DEVELOPMENTS IN PRECAMBRIAN GEOLOGY 4

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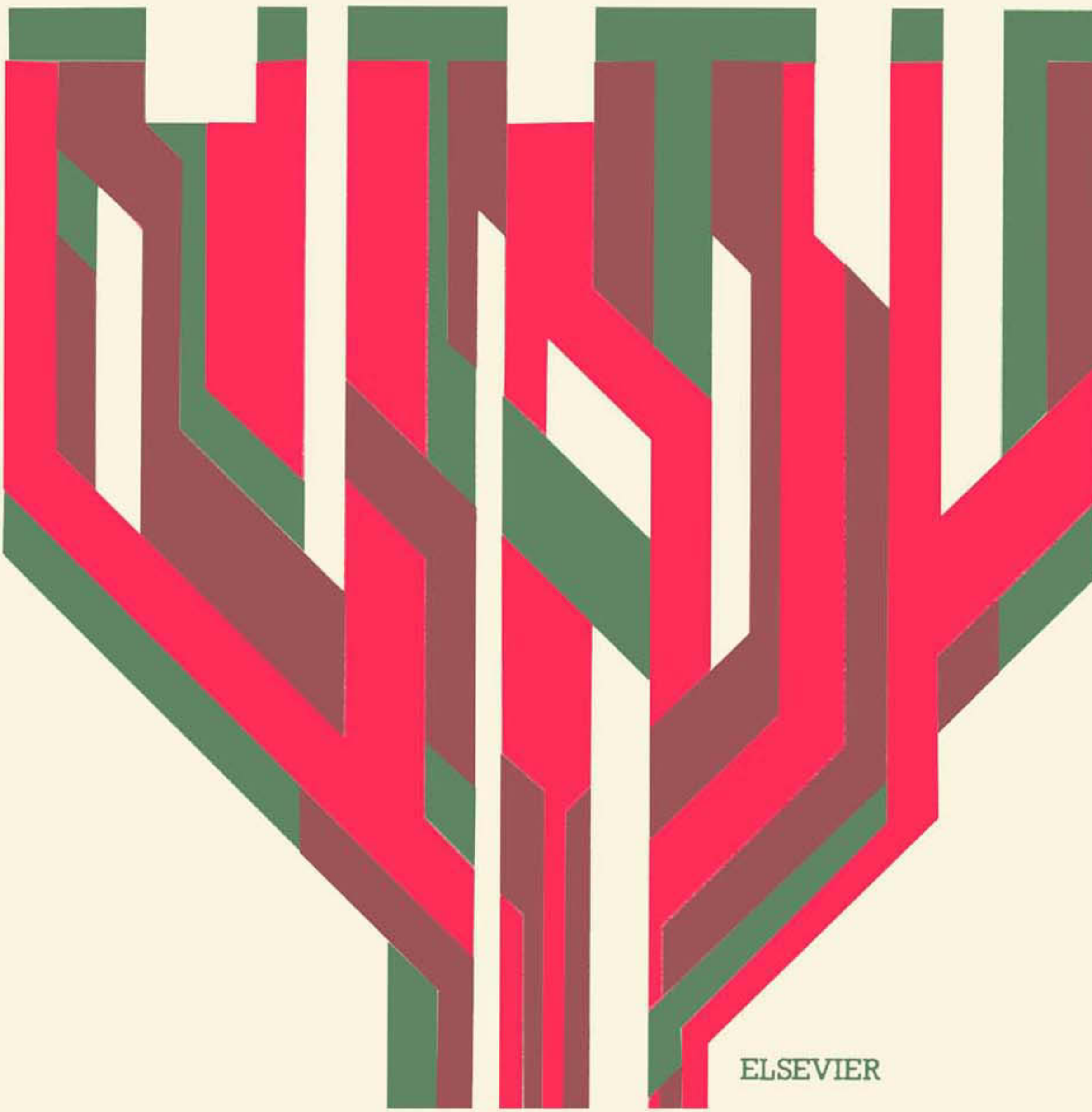
PRECAMBRIAN PLATE TECTONICS

A. KRÖNER (EDITOR)



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DEVELOPMENTS IN PRECAMBRIAN GEOLOGY 4

PRECAMBRIAN PLATE TECTONICS

Edited by

A. KRÖNER

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FOREWORD

In 1968 W. Jason Morgan, in a paper published in the Journal of Geophysical Research, introduced the concept of plate tectonics according to which the earth's surface is considered to be divided up into a number of rigid plates. The plate boundaries consist of mid-ocean ridges, oceanic trenches, great faults, and active mountain belts. It is now generally believed that the movements of the continents and of the sea floor are part of large-scale movements of plates.

I was introduced to sea-floor spreading in 1965 in Canberra in Australia when, during a brief visit, I met Professor J. Tuzo Wilson who, over a glass of sherry in his flat at the University House, produced from his pocket, along with his ever-present notebook, a most convincing cardboard model of sea-floor spreading.

Since those times, the fundamental problem whether it is possible to extend plate-tectonic processes back into the Precambrian has caused considerable, often heated, discussion. Opinion is still divided: did the evolution of the crust during the Precambrian follow the uniformitarian principle, consequently being compatible with Phanerozoic plate tectonics, or did global tectonic mechanisms undergo considerable changes since the early Precambrian? The problem is caused by the uncertainty of the thermal history of the earth, growth of the continental crust, tectonic style, and formation of calc-alkalic igneous rocks. Data exist indicating that the crust of the earth was both lithologically and chemically complex as far back as the available geologic record indicates.

This volume deals with various aspects of Precambrian plate tectonics, whether considered uniformitarian or not. It contains a wealth of new data dealing with Precambrian crustal evolution. It is probably not surprising that attempts to fit Precambrian crustal evolution into the New Global Tectonics have met with variable success and have produced many contrasting views.

The Authors and the Editor alike are to be congratulated on producing a most interesting volume at this most appropriate stage in the study of plate-tectonic models applied to Precambrian crustal evolution. May this book stimulate much more research in its fascinating field.

Helsinki, April 30, 1980 KALERVO RANKAMA

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INTRODUCTION

The application of present-day plate-tectonic models to Precambrian crustal evolution is still a matter of considerable debate. In the early 1970's many models were proposed on the uniformitarian assumption "the present is the key to the past" and invoked processes closely analogous to present crust-forming and destruction mechanisms right back to the Archaean. These interpretations were heavily criticized by field-oriented geologists who claimed that many Precambrian rocks and their tectonic relationships were strikingly different from those in Phanerozoic terrains and that non-plate-tectonic processes must therefore have operated during most of the earlier history of the earth.

Considerable international and interdisciplinary research has since been focussed on the problem of Precambrian crustal evolution as a part of the Geodynamics Project, the International Geological Correlation Programme and other efforts, and during the last few years opinions have changed so much that the question no longer is whether plate tectonics operated in the Precambrian but what form of plate tectonics took place and how the crustal interaction processes changed through time in response to the declining heat flow of the earth.

With this in mind contributors to this volume were asked to present their views on Precambrian crustal evolution and were encouraged to speculate on geodynamic processes that may have operated during specific periods of Precambrian time. As manuscripts were submitted to the Editor, it became apparent that the major theme of this book would be whether uniformitarian concepts are applicable to the geodynamic evolution of the continental crust or, as one contributor put it, did uniformitarian causes have non-uniformitarian effects?

There is as yet no clear answer to this question, but it is to be hoped that the data and conclusions presented in this volume will assist the reader in assessing the validity of present and past speculations on the early crustal history of our planet. The student of the earth sciences will also find that the same set of data can be interpreted in entirely different ways, as is particularly well demonstrated by the papers on Archaean evolution and Precambrian metallogeny, and that there are no, and probably never will be, standard models for crustal evolution of world-wide applicability.

The reader will notice that aspects of the Precambrian evolution of North America and Africa are particularly well covered while other shields are not or only poorly discussed. This is unintentional and partly results from the fact that several invited contributors withdrew at short notice. Nevertheless, it is felt that the papers submitted give a good reflection of present thinking of Precambrian geodynamics.

As in all multi-author volumes of this kind, there are problems of standardization of language, style and the presentation of data. The spelling of words follows the English usage, e.g. Archaean, not Archean, etc., also "farther" is preferred to "further" if used as comp. of far. The expressions "terrain" and 'terrane" are used synonymously since some North American contributors prefer the latter as defined in the AGI Dictionary of Geological Terms.

Since the manuscripts were received, Rhodesia has been renamed Zimbabwe. Alterations have been made to the text but it was not possible to amend the text figures, in which references to Rhodesia should now read Zimbabwe. The term Rhodesian craton has been retained as a formal geological name because of its established place in geological literature.

All ages are abbreviated Ma (million years) and Ga (billion years), respectively, and contributors were asked to quote radiometric age data on the basis of the new decay constants. In a few cases the recalculated figures were inserted by the Editor. Cross-references were also inserted by the Editor where appropriate and are printed in italics.

Almost every paper was reviewed by two referees, frequently fellow contributors, and the Editor expresses his sincere appreciation to the colleagues listed herunder who have devoted considerable time to the improvement of the manuscripts submitted: A. J. Baer, W. R. A. Baragar, J. M. Barton Jr., A. Berthelsen, J. C. Briden, D. Bridgwater, K. Burke, R. Caby, E. H. Chown, W. R. Church, C. Craddock, E. Dimroth, R. F. Emslie, W. S. Fyfe, I. G. Gass, A. Y. Glikson, A. M. Goodwin, J. A. Hallberg, R. B. Hargraves, C. J. Hawkesworth, K. Heier, J. B. Henderson, J. V. Hepworth, A. Hofmann, W. Jacoby, B. Jahn, D. D. Klemm, T. E. Krogh, R. St J. Lambert, H. Martin, M. O. McWilliams, W. R. Muehlberger, E. G. Nisbet, J. D. A. Piper, K. Rankama, R. W. R. Rutland, F. J. Sawkins, D. F. Strong, G. G. Suffel, J. Sutton, D. H. Tarling, W. R. Van Schmus and B. F. Windley.

This volume is a contribution to IGCP Project No. 92 (Archaean Geochemistry) and to the work of the IUGS Commission on Tectonics. It is hoped that it will stimulate further research into the fascinating subject of Precambrian crustal evolution.

Mainz, March 21, 1980 ALFRED KRÖNER

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Chapter 1

PRECAMBRIAN ROCKS IN THE LIGHT OF THE PLATE-TECTONIC CONCEPT

BRIAN F. WINDLEY

ABSTRACT

Two geophysical constraints limit the style of tectonic processes in the Precambrian: heat production and palaeomagnetism. A high rate of heat production by radiogenic decay implies mantle convection and growth of new oceanic lithosphere, and the definition of apparent polar wander paths implies the existence of rigid plates at least in the Proterozoic.

The Archaean (3900—2500 Ma) was a period of vigorous growth (accretion) and destruction (subduction) of oceanic lithosphere with consequent production of voluminous calc-alkaline melts with tonalitic end-products deformed and metamorphosed into continental gneisses and granulites in lower crustal levels. Extensive rifting in marginal basins gave rise to multiple greenstone belts as proto-ophiolites. No stable cratons formed in these early proto-plate times.

The Archaean—Proterozoic boundary represents a transitional period during which the Archaean-thickened continental crust was uplifted and eroded to give rise to abundant clastic debris. Many intrusions and dykes were injected into this stabilizing crust.

The Proterozoic (2500—600 Ma) was a period transitional in type between the Archaean and the Phanerozoic. Some greenstone belts continued to form in the early Proterozoic, but from this time the Wilson cycle can be recognized, designating the advent of modern-style plate tectonics. Stable cratons were bordered by narrow fold belts formed by Cordilleran—Himalayan collisions which led to the formation of the first decipherable indentation (slip-line) fracture systems. Many abortive attempts at rifting in stabilized continental plates gave rise to aulacogens, alkaline igneous activity and the mid-Proterozoic anorthosite-rapakivi granite suite. Continental breakup and rearrangement of plates after 1100 Ma produced the Grenville—Dalslandian and Pan-African—Braziliano fold belts.

INTRODUCTION

Because the Precambrian occupies about seven-eighths of geological time, an understanding of the tectonic development of Precambrian mobile belts is essential for an understanding of the mode of evolution of the earth's crust. In recent years there has been an increased focus on problems attached to the formation of many Precambrian mobile belts, in particular with respect to the possibility that they may have formed by some process akin to modern plate tectonics (Windley, 1977a). We know that mantle convection controlled lateral lithospheric plate motions and the resultant formation of orogenic belts, certainly in the last 200 Ma and most probably in the last 700—800 Ma.

The question that we now have to ask is, did mantle convection occur prior to the late Proterozoic, or was some other non-uniformitarian tectonic mechanism responsible for the formation of orogenic belts for the bulk of the Precambrian?

In order to answer this question we can most easily go back to the rocks and assess such parameters as their mutual associations, age relationships, structural development and geochemistry, and establish whether or not they are consistent with what we can expect from an evolution via a Wilson cycle. If our Precambrian belts are intelligible in such a manner, there is no need to search elsewhere for some non-uniformitarian mechanism.

But first let us consider some independent geophysical constraints on Precambrian tectonic development: heat production and palaeomagnetism.

THE HEAT FACTOR

It is generally accepted that heat production by breakdown of radiogenic material was greater in the past because of the exponential decay rate of radioactive isotopes. The question arises, how was this greater amount of heat used up or lost from the earth? For an answer we can turn to the modern earth for which there is a great deal of data on heat production and loss. Today about a third of the heat loss escapes by conduction through the continental lithosphere, a third by conduction through the oceanic lithosphere, and a third (about two-thirds according to J. G. Sclater, pers. commun., 1979) is used up in generation and aging of new crust/lithosphere at plate accretion boundaries (Burke and Kidd, 1978). Also, the oceanic crust cools about ten times faster than the continental crust and about 45% of the earth's total heat is lost by the oceanic plate creation—subduction process (Bickle, 1978). All this heat associated with the plate-tectonic process moves upwards via the convection cells in the mantle and it is not surprising that the oceanic lithosphere is the main area of heat loss because its base is effectively the upper boundary of a mantle convection cell.

Conduction through continental crust is not only an inefficient process of heat loss today, but also it could not have been substantially greater in the Archaean because if temperatures in the lower part of the crust, which consists mostly of granulites of dioritic composition (Tarney and Windley, 1977), were raised much above 800°C extensive partial melting would have given rise to substantial minimum-melting granites (the temperature at the base of the present continental crust in the shields is at the most 500°C). Not only is there a lack of evidence for such bulk melting in presently exposed Archaean granulites (which have mineral assemblages suggesting formation at c. 10 ± 1 kb at c. 30 km depth, Tarney and Windley, 1977), but there is a lack of such granites in higher-level Archaean terrains such as the Superior Province of Canada (Burke and Kidd, 1978), Alternatively, Davies (1979) suggested that there may have existed beneath stable continental crust a root zone at least 200 km thick which acted as a thermal buffer between the crust and the convecting mantle, and that the greater Archaean heat flux would have been removed, mainly through faster sea-floor spreading.

In contrast to conduction, convection is an extremely efficient means of using up and dissipating heat, and it is easily possible to increase the amount of heat dissipated by this process, either by increasing the total length of mid-oceanic ridges or by increasing the rate of plate production (Bickle, 1978). Therefore it is difficult to escape the conclusion that some form of plate-tectonic activity was in operation during the Precambrian and that it would have operated on a more extensive scale than at present. The current average rate of plate production is 3 km^2 a^{-1} and, after careful consideration of constraints and variables, Bickle (1978) concluded that plate production was at least 18 km^2 a^{-1} 2800 Ma ago. In a non-expanding earth this increase must have been complemented by a similar increase in plate subduction and, in turn, in the degree of partial melting of the downgoing slab and, consequently, a comparable increase in the production of the new magmas that gave rise to either andesites at high crustal levels or tonalites in batholithic proportions in deeper levels. These suggestions are in fact entirely consistent with, firstly, the very high proportion of tonalitic rocks intruded in the Archaean into deep crustal levels; after deformation they were transformed into largely tonalitic gneisses at high amphibolite or granulite grade, such as the Nûk gneisses in Greenland and the Scourie gneisses in Scotland. These rocks tend to have low initial strontium isotope ratios of c. 0.701, implying short crustal residence times since their derivation from a low Rb/Sr source, probably the mantle (Moorbath, 1975; Moorbath and Taylor, this volume, Chapter 20). Secondly, they are consistent with the large quantities of volcanic rocks produced in Archaean greenstone belts. In short, the Archaean appears to have been the principal period of crustal growth in earth history. A significant part of the present-day continents was created by the end of the Archaean (Windley, 1977b). McCulloch and Wasserburg (1978) concluded from their study of Rb-Sr and Sm-Nd systematics that the bulk of the Superior, Churchill and Slave Provinces of the Canadian Shield was formed in the period 2700—2500 Ma ago.

Today much of the resultant heat production is directly related to mantle convection and plate production and the long-term thermal and tectonic evolution of the earth must be related to the abundance and life span of these heat-producing isotopes (Pollack and Chapman, 1977). In short, the isotopes provide us with the link between the tectonic activity of the present and the past.

PALAEOMAGNETISM

Apparent polar wander paths have been constructed for several continental blocks such as North America, Australia and Africa back to early Proterozoic time (e.g. Fig. 1-1), but it has not yet proved possible to extend such paths into the Archaean. Irving and McGlynn (1976) demonstrated a close correlation between palaeomagnetically determined Proterozoic latitudes and the occurrence of glacial deposits, and the rates of relative polar motion are similar to those known for the Phanerozoic; this is clear evidence for the existence of mantle convection in the Proterozoic. Also, Proterozoic polar wander paths typically have hairpins comparable to Phanerozoic paths which define major changes in the horizontal direction of movement (relative to

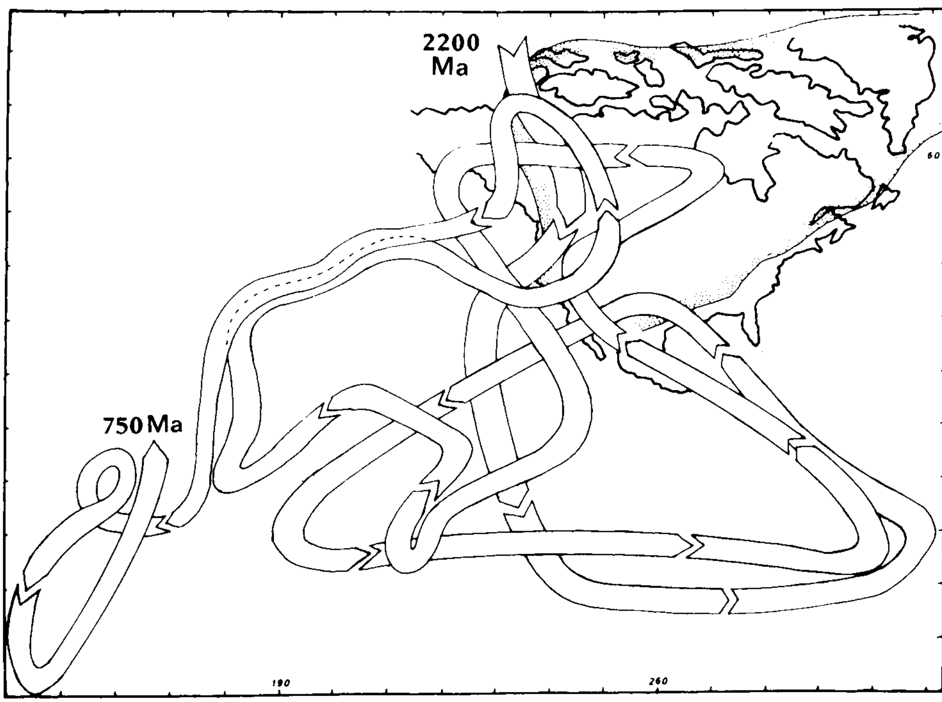

Fig. 1-1. Apparent polar wander path for the period 2200-750 Ma for the Laurentian craton (after Piper, 1978).

the pole) of the lithospheric plates and which correspond to major tectonic events such as the Grenville Orogeny and the orogeny of the Coronation Geosyncline. Irving (1979) estimates that if true polar wandering has been small, the average N—S drift of North America was 4—5 cm a^{-1} in the Proterozoic compared with 1—1.5 cm a^{-1} for the Phanerozoic.

To summarize so far, the Precambrian thermal and palaeomagnetic data taken together meet the essential geophysical requirements for plate growth and lateral motion. We shall now see if the Precambrian rock record is consistent with these constraints.

ARCHAEAN CRUSTAL EVOLUTION

There are two types of Archaean tectonic regime: greenstone-granite belts and granulite-gneiss belts. Both formed at various times throughout the Archaean and their characteristic features are so different that they must have formed in different tectonic environments or crustal zones; in other words, the granulite-gneiss belts are not just highly metamorphosed greenstone-granite belts.

Greenstone-granite belts

Greenstone belts are prominent in Zimbabwe (Rhodesia), South Africa, West Australia, Canada, India and Finland. In Zimbabwe there were three periods of formation (Sebakwian 3600 Ma, Lower Bulawayan 3000 Ma, Upper Bulawayan and Shamvaian 2800—2700 Ma) (Wilson et al., 1978; *see also Nisbet et al., this volume, Chapter 7, ed.*), but in several continents only one age group has so far been defined, particularly in the range 2800—2600 Ma.

Characteristic features of the belts include a synformal structure (except for early nappes in some Rhodesian belts), a size ranging from 40—250 km across to 120—800 km long, a mostly low-pressure type of greenschist-grade regional metamorphism, intrusion by plutons that range from tonalites and trondhjemites through granodiorites to potassic granites, and a threefold stratigraphic division:

(a) A lower group of pillow-bearing komatiitic ultramafic and mafic volcanics, the basaltic komatiites having bulk compositions broadly comparable with modern mid-ocean ridge basalts.

(b) A central group comprised largely of andesites and silicic volcanics with a calc-alkaline chemical affinity, whose trace and rare-earth elements are very similar to those in modern island arc volcanics (Jahn et al., 1974) or Andean volcanics (Taylor and Hallberg, 1977) except for higher contents of transition elements.

(c) An upper group made up of clastic greywackes, sandstone and conglomerates and chemically precipitated banded iron formations, cherts and limestones.

Also important are Cambrian greenstone belts in closed marginal basins in the Lachlan fold belt of SE Australia (Crawford and Keays, 1978) and early Proterozoic greenstone belts of the Amisk group in Canada (Bell et al., 1975; Stauffer et al., 1975).

Many tectonic models have been proposed to explain the formation of greenstone belts (Windley, 1977a). The one most consistent with geological relationships and geochemical data on the one hand and with a priori expectations of plate-tectonic behaviour on the other, is based on the remarkable similarity of greenstone belts with modern closed marginal basins that lie behind the arcs of the Cordilleran fold belt of North and South America (Burke et al., 1976; Tarney et al., 1976). Marginal basins form by crustal extension behind a volcanic island arc or behind an active arc on a continental margin. On the west side of the Pacific they are still open but on the eastern side they have been closed by movement of the adjacent arc towards the continent. Characteristic features of these basins, which are relevant to comparison with Archaean greenstone belts, are:

(a) A synformal structure.

(b) A lower group of serpentinized ultramafics together with pillow-bearing basalts whose composition is similar to that of mid-oceanic ridge basalts except for higher abundances of large ion lithopile elements which are more comparable with greenstone belt basalts.

(c) An upper group of sediments derived partly from an adjacent continental basement and partly from the evolving arc.

(d) Intrusion by plutons ranging from tonalites to potassic granites.

(e) A greenschist grade of low-pressure metamorphism.

From the above data it is clear that modern marginal basins do not differ to any significant degree from Archaean greenstone belts, and what differences there are can be related to variables such as heat flow and geothermal gradient which are known to have been higher in the Archaean.

The marginal basin model is particularly apposite for greenstone belt formation because it satisfactorily accounts for predictable tectonic relationships. Firstly, it provides a mechanism for the rifting and extension by mantle diapirism behind an arc and for closure by movement of an arc towards a continent. Secondly, considering the previous argument for increased plate generation and subduction in the Archaean caused by the higher heat production, the main oceanic plates would predictably have been destroyed by subduction, just as they would be at any other time of earth history and yet the lower greenstone volcanics have the chemical character of modern oceanic crust. The best way we know today of forming oceanic crust and preserving it to a remarkable degree (i.e. not like that in tectonic mélanges in subduction trenches) is in marginal basins. This is the same reason as why many Phanerozoic ophiolites, which in general were once thought to be remnants of main oceanic crust, are now considered to be remnants of marginal basins, e.g. Baie Verte in Newfoundland, Cordilleran fold belt, Troodos complex, Cyprus, and S. Chile (Proc. Ophiolite Symposium, Cyprus, 1980). The geological features (rock types, rock associations, mutual age relationships, structure and metamorphism) and geochemical parameters of the older and younger belts are, by and large, so similar that the marginal basin model stands up remarkably well to testing on a variety of fronts. The environment is ideal for the close association of oceanic basalts, island-arc andesites, clastic sediments derived from a continental terrain and a volcanic arc, and tonalite-granite plutons of Andean type. No other tectonic model can so satisfactorily account for the interrelationships of these diverse features.

Granulite-gneiss belts

These highly metamorphosed belts contain three rock components: quartzo-feldspathic gneisses, layered peridotite-gabbro-anorthosite or leucogabbro-anorthosite complexes, metavolcanic amphibolites and metasediments (marbles, quartzites and micaschists). Prominent belts are in West Greenland and Labrador, the Scourian of NW Scotland, the Limpopo belt of southern Africa, the Kola, Aldan and Anabar Shields of the USSR, and southern India. (*See also Barton and Key, this volume, Chapter 8, and Moralev, this volume, Chapter 10, ed.*)

Most of these regions went through a major accretionary isotopic event in the period 2700—3100 Ma (Moorbath, 1976). The following account concerns the three components of the predominant 2700—3100 Ma rock groups that constitute the bulk of these terrains.

The quartzo-feldspathic gneisses make up about 85% of these terrains. In amphibolite-grade areas they contain hornblende and biotite, and in granulite areas they have hypersthene, being acid granulites or charnockites. Predominantly they are tonalitic to granodioritic in composition. Mostly they are well foliated, but in Greenland, Labrador, Scotland, Zimbabwe (Rhodesia) and southern India they are locally less deformed and less foliated in low-strain zones. Such zones map out as lenses from a few metres to a kilometre across, and within them the gneisses are in their original undeformed state as homogeneous tonalites or granodiorites. Thus by studying the strain patterns it is possible to follow the transition from undeformed granitoids to foliated gneisses. In the low-strain zones the tonalites clearly have discordances and apophyses against older rocks (which may range from gneiss to metavolcanic amphibolites and peridotite-gabbro-calcic anorthosite complexes, McGregor, 1973). Furthermore, such gneisses from Scotland and Greenland have calc-alkaline bulk compositions and TiO_2-SiO_2 abundance levels that are appreciably different from those of Archaean or post-Archaean sediments, although they are comparable with those of calc-alkaline plutonic rocks from the British Caledonian active continental margin (Tarney, 1976).

In attempting to explain the mode of origin of Archaean granulite-gneiss belts the most critical problem is how to generate the voluminous quantities of igneous tonalite that now make up the bulk of most of the belts. If we turn to the modern earth we find that there is only one tectonic regime that contains comparable volumes of tonalite — the Cordilleran fold belts of South America with their vast Mesozoic batholiths, at least 46—55% of which consist of intrusive tonalite (Pitcher, 1978).

When considered in detail it is clear that the geological and geochemical features of the more deeply eroded batholiths are indeed remarkably similar to those of the early Precambrian granulite-gneiss belts (Tarney, 1976; Windley and Smith, 1976). The most appropriate examples are the British Columbian (Roddick and Hutchinson, 1974), Chilean (Hervé et al., 1974) and southern Californian (Gastil, 1975) batholiths, the first two of which have probably been eroded at least 15—20 km. In fact granulites are just beginning to appear at the present erosion level within the British Columbian batholith with assemblages indicating formation at 5—8 kb (Hollister, 1975) as compared with 30 km (10 kb) estimated from mineral assemblages for Archaean granulites (Tarney and Windley, 1977). A surprisingly large proportion of these deep-seated batholiths are highly deformed; foliated tonalites and tonalitic gneisses containing ellipsoidal pancake-shaped dioritic xenoliths are abundant and these gneisses have risen diapirically into nappes and fold interference patterns (MacColl, 1964; Hutchinson, 1970). The batholiths contain a great many layered igneous complexes that consist predominantly of hornblende-bearing peridotite-gabbro-calcic anorthosite (Mullen and Bussell, 1977).

In summary, the batholithic root zones of the main arcs of Mesozoic—Cenozoic Cordilleran fold belts provide an excellent modern analogue for most Archaean granulite-gneiss belts.

EARLY—MID-PROTEROZOIC GEOSYNCLINES, MOBILE BELTS AND PLATE COLLISIONS

In the period 2100—1700 Ma ago several geosynclines and mobile belts were probably created by some form of plate opening and closure (Fig. 1-2). The following are key examples:

(1) In NW Canada the Coronation Geosyncline was the early expression of the Wopmay Orogeny which may have formed by a remarkably complete Wilson cycle of events, according to Hoffman (in press).

(2) Molnar and Tapponnier (1975) have shown how the major wrench faults of Asia can be interpreted in terms of slip-line theory as the result of the northward indentation of India into the Asian plate. From several recent studies it seems likely that some Proterozoic collisional mobile belts have left a record of their associated indentation fracture patterns in their foreland regions. For example, regular and symmetrical wrench fault patterns led Gibb (1978) to suggest that the Slave craton, acting as a mini-plate, had indented southeastwards into the Churchill Province of Canada in the early Proterozoic. Comparable fault patterns resulted from the collision of the Coronation continental margin and the Bear microcontinent (Hoffman, in press), and Gibb (1975) suggested that large dextral faults cutting the margin of the Superior craton were the direct result of collision between the suturing Superior and Churchill plates. The lateral convergence of the Superior and Slave Provinces is supported by the apparent polar wander paths of Cavanaugh and Seyfert (1977). The Circum—Ungava geosyncline

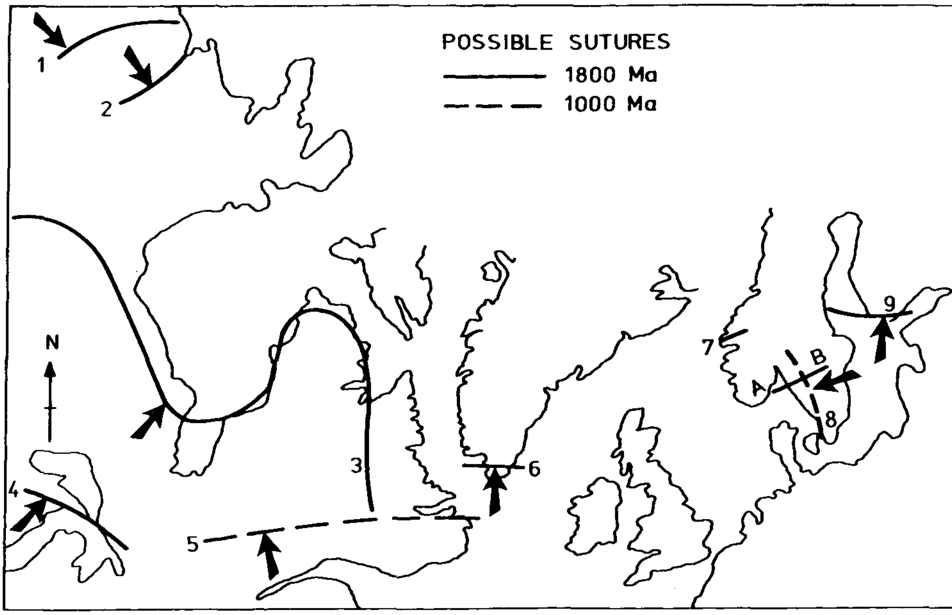

Fig. 1-2. Map of the North Atlantic region showing the position of possible early and late Proterozoic sutures. *1* = Hoffman (in press); *2* = Gibb (1978); *3* = Gibb and Walcott (1971); *4* = Van Schmus (1976); *5* = Irving et al. (1974); *6* = Watterson (1978); *7* = Krogh (1977); *8* = Berthelsen (1976); *9* = Hietanen (1975). Arrows indicate movement direction of indenting plates and possible original dip direction of subduction zones. *A-B* gives location of section in Fig. 1-3.

has long been regarded as a continent—continent collisional suture (Gibb and Walcott, 1971)*, and more recently regional gravity profiles have been used to define the crustal margins thickened by ductile flow of the over-riding plate (Gibb and Thomas, 1976; Kearey, 1976). The 1800—1750 Ma Amisk basin, which has the stratigraphic and chemical character of an Archaean greenstone belt (Bell et al., 1975), may be an early Proterozoic island-arc deposit (Stauffer et al., 1975).

Watterson (1978) proposed that the two sets of deep-seated, 1800 Ma old, orthogonal shear belts in Greenland and NW Scotland were the result of a northward continental collision in South Greenland that was responsible for the formation of the contemporaneous Ketilidian mobile belt and Julianehaab granodioritic batholith. This is a very appealing explanation of these intra-continental Proterozoic shear belts which was first suggested by Molnar and Tapponnier (1975, p. 425). The Ketilidian mobile belt was briefly compared with the Andean belt of western America by Bridgwater et al. (1973). In

* For alternative interpretations see also Baragar and Scoates (this volume, Chapter 12) and Dimroth (this volume, Chapter 13), ed.

particular, the Julianehaab batholith (Allaart, 1967) has many features that are extremely similar to those of the American Cordilleran batholiths, the most prominent of which are synplutonic-syntectonic hornblende-bearing basic dykes that quench at high water pressures (c. 4 kb) (Roddick and Armstrong, 1959; Watterson, 1968) and net-veined acid-basic complexes (Windley, 1965; Cobbing and Pitcher, 1972 in Peru; my observations in the southern Californian batholith).

(3) In the 1800—1600 Ma old Svecofennian belt of Finland widespread, early trondhjemitic-tonalitic gneisses that contain remnants of hornblende gabbro complexes pass northwards into late-tectonic microcline granites with eutectoid compositions. According to Hietanen (1975) the potassium content of the differentiation series increases with decreasing age northwards (and thus this geochemical-chronological pattern is similar to that passing eastwards across the Sierra Nevada batholith) and she suggests that a comparable plate-tectonic model can adequately explain its development. The example is interesting because it suggests that the tonalite-trondhjemite gneiss belts, previously described as typical of the Archaean, may have continued to form in the Proterozoic, albeit in more narrowly confined mobile belts.

(4) In the 1700—1800 Ma old terrain of SW Norway eclogites occur as lenses in the gneisses and in enclosed ultramafic-anorthosite layers. On considering the stability fields and compositions of coexisting minerals, Krogh (1977) calculated that the former type, which contains glaucophane, formed at pressures of 20—22 kb, whilst Lappin and Smith (1978) estimated 30—40 kb for the same rocks, and Carswell (1974) obtained pressures of 22—37 kb for garnet lherzolites in the gneisses. Derivation of mantle material or metamorphism at such high pressures and subsequent incorporation into the continental crust can most readily be explained in terms of continent—continent collision tectonics, perhaps related to major interthrusting and obduction processes.

(5) In the Great Lakes area of North America the sequence of events in the early—mid-Proterozoic mobile belt bordering the Archaean Superior craton may compare with that expected for a continental margin which later becomes involved in a collision-type orogeny (Van Schmus, 1976; Cambray, 1978; *see also Van Schmus and Bickford, this volume, Chapter 11, ed.*). Rifting and block faulting created several basins and tholeiitic dykes were intruded into adjacent basement. The Huronian and Marquette Supergroups and the Animikie Group were laid down in the subsiding back-arc shallow epicontinental basins on an attenuated continental basement bordering a volcanic-plutonic arc. Deformation, high-grade regional metamorphism and extrusive and intrusive igneous activity occurred during the Penokean Orogeny 1850—1900 Ma ago in the Cordilleran stage, and this was followed 1790 Ma ago by the extrusion of rhyolitic and ignimbritic lavas and the intrusion of high-potash granites in the Himalayan stage of development. A possible western

continuation of this mobile belt extends into southern Wyoming where a Proterozoic subduction zone brought older continental crust, overlain by a Proterozoic coastal plain, into collision with a volcanic arc between 1760 and 1720 Ma ago (Hills et al., 1975).

(6) In Arizona, U.S.A., at least three 1800—1700 Ma andesitic arcs are identifiable with petrochemical polarity overlying sub-arc oceanic crust and there are remnants of synchronous arcs in six states farther east. These intra-oceanic arcs are flanked by thick flysch-like greywacke and volcanoclastic sequences analogous to trench-slope deposits, and by finer clastics in back-arc basin-like deposits, and are capped by molasse-basin assemblages. Syn-tectonic, synvolcanic granodiorite, minor tonalite and granite batholiths intrude the base of the arc assemblages and are restricted to an intra-arc setting. Basic volcano-plutonic arc generation and subduction was followed by more silicic volcano-plutonic arc generation and subduction which accreted the earlier arcs to the northern Archaean block. Accompanying the plutonism and accretion was widespread deformation and metamorphism from 1750 to 1650 Ma ago (Anderson, 1977).

MID-PROTEROZOIC RIFTING AND MAGMATISM

Anorthosite-rapakivi granite suites were intruded in Scandinavia about 1700 Ma ago and they are aligned with a group in North America that has an age of 1500—1400 Ma. A similar belt extends across the continents of the southern hemisphere (Windley, 1977a).

The anorthosites have andesine-labradorite and the granites, in places substituted by quartz mangerites, are enriched in K (potash feldspar) and Fe/Mg (fayalite), and are associated with ferrodiorites. These rocks are locally associated with, and in places succeeded by, acidic-basaltic extrusives, alkaline and peralkaline plutonic complexes and continental sandstones. In an extensive review Emslie (1978) concludes that the anorthosite-granite suite represents the product of bimodal magmatism intruded in an anorogenic environment, and thus was associated at a late stage with alkaline complexes intruded into graben which controlled the outpouring of rhyolitic lavas and the sandstone sedimentation.

The rock associations are remarkably similar to the Palaeozoic (470—412 Ma) "Younger Granite" complexes of Niger and Nigeria in which fayalite granites are associated with labradorite anorthosites, gabbros, syenites and alkaline granites (Moreau et al., 1978). These rocks were intruded in a linear network of subvolcanic ring complexes in an anorogenic setting.

Dewey and Burke (1973) suggested that the Proterozoic anorthosite-granite suite was intruded into the thickened overriding plate of a continental collisional mobile belt. However, it now seems more likely from detailed consideration of rock associations and geochemistry that the suite was generated by partial melting of deep crustal rocks, and intruded in the early stages of major intracontinental rifting during the mid-Proterozoic (Emslie, 1978).

GRENVILLE—DALSLANDIAN MOBILE BELT

There has been much dispute about the mode of formation of this 1000 Ma old orogenic belt and it is not possible to debate here the opposing ideas and tectonic models (see Windley, 1977b, pp. 135—138; also Baer, this volume, Chapter 14); nevertheless, "most experts agree that the Grenvillian orogeny may be explained in terms of plate tectonics" (Baer et al., 1974).

The palaeomagnetic data of Irving and McGlynn (1976) suggest that the bulk of the Grenville belt moved about 5000 km in a northwestward direction between 1125 and 1000 Ma ago to collide eventually with the Laurentian Shield. However, McWilliams and Dunlop (1978) point out that the variability in blocking temperatures of magnetic systems makes it unlikely that palaeomagnetic data from within the Grenville belt will be useful in testing collisional models, and that even if the existence of a suture were proven, the opening and closing of a *small* ocean would be undetectable palaeomagnetically. A 1200 km long linear negative Bouguer anomaly 100 km inside the Grenville belt may indicate the suture position because it defines a boundary between crustal blocks of different mean density and thickness (Thomas and Tanner, 1975). Part of the suture probably lies between Bancroft and Renfrew along the margin of the Elsevir batholith where pillow basalts of oceanic origin overlie a mafic-ultramafic meta-igneous complex. The basalts are intruded by granodioritic to granitic plutons (Chapell et al., 1975) and low-K tholeiites and andesites have trace-element chemistry consistent with production in an arc environment (Condie and Moore, 1977). Following uplift and erosion these rocks were overlain by mio-geoclinal sediments of the Flinton Group which were then subject to deformation and metamorphism during continental collision (Chesworth, 1972). The effects of the Grenville collision are manifest farther west in Arizona and New Mexico (Anderson, 1977). Evidence of the pre-Grenville rift zones lies in the Keweenawan, Seal Lake and Gardar magmatic provinces, and Harp Lake and Abitibi dyke swarms.

Considerable advance has recently been made in understanding the evolution of the Dalslandian belt in southern Sweden and Norway in terms of Himalayan collision tectonics. Zeck and Malling (1976) and Berthelsen (1976) demonstrate the presence of major low-angle thrusts on which slabs of crustal thickness were piled upon each other. The possible suture zone is outlined by a regional belt of thrust stacks 200—300 km long and 50 km wide (Fig. 1-3).

THE PAN-AFRICAN—BRAZILIANO MOBILE BELTS

These mobile belts began their development in the late Precambrian and their peak of orogenic activity was in the period 750—550 Ma ago. Recently there has been considerable success in establishing in a preliminary manner

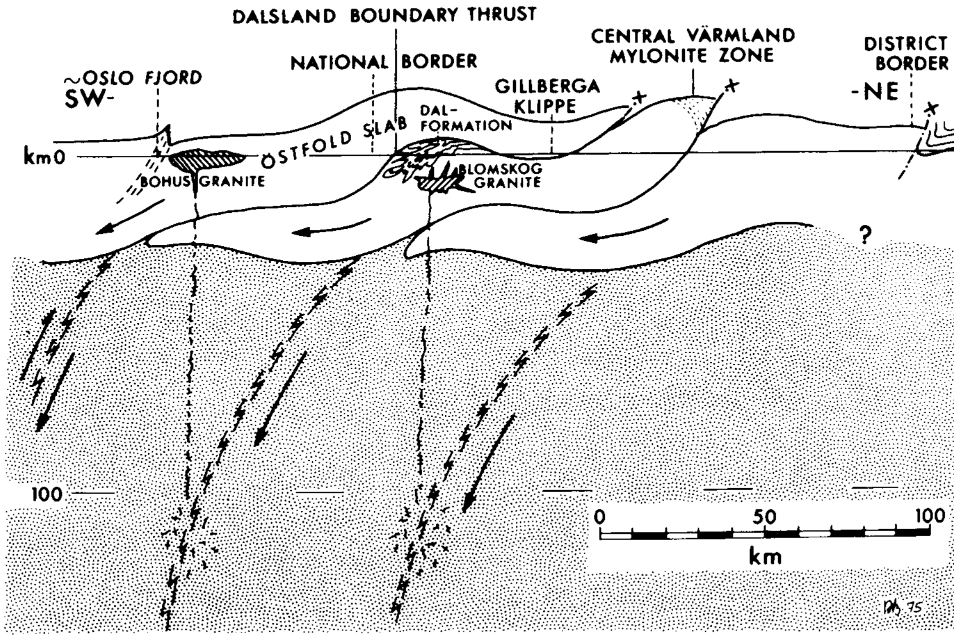

Fig. 1-3. Profile through the Swedish part of the Dalslandian orogen (after Berthelsen, 1976). For location, see *A-B* in Fig. 1-2.

geological relationships which point cogently towards a plate-tectonic mode of development. For example, gabbros, sheeted dykes (dolerites) and basalts are known in the Arabian Shield and NE Africa (Frisch and Al-Shanti, 1977; Gass, 1977) and a great many possible ophiolites in Africa were proposed by Shackleton (1976), although no details were given to corroborate the proposal (Fig. 1-4). In Morocco there is a 4—5 km thick ophiolite at Bou Azzer (Leblanc, 1976, this volume, Chapter 17). At Bleida in Morocco there is a well-preserved trailing continental margin sequence of shelf limestones and quartzites overlain by alkaline tholeiitic basalts related to early break-up of continental crust (dated at 788 ± 9 Ma, Rb-Sr, Clauer, 1976), followed by stratiform distal massive copper sulphide deposits whose position was controlled by sedimentation of off-shore shales and acid volcanism (Leblanc and Billaud, 1978). The Bou Azzer ophiolite was thrust over this continental margin sequence 685 Ma ago (Leblanc, this volume, Chapter 17). In the western Hoggar of Algeria there are volcanoclastic deposits, andesites to dacites and calc-alkaline batholiths interpreted as having formed in connection with island-arc—Andean subduction and high-level nappes and synkinematic granites related to crustal thickening and crustal melting during continental collision (Bertrand and Caby, 1978; Caby et al., this volume, Chapter 16). In NE Sudan the ophiolites and rhyolitic-andesitic

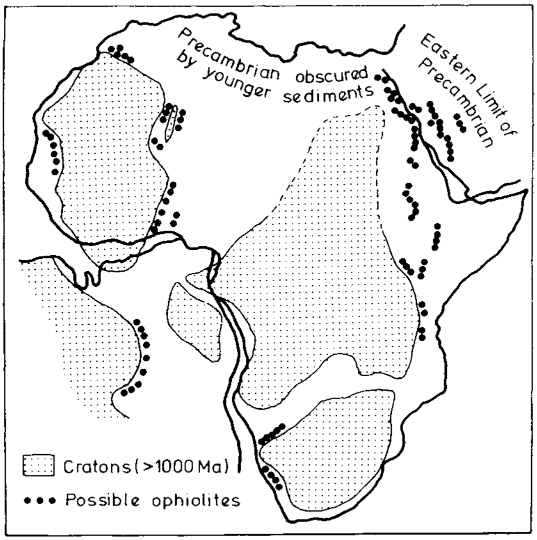

Fig. 1-4. Map showing possible ophiolites in the Pan-African belts of Africa and Brazil (after Shackleton, 1976).

pyroclastics are intruded by calc-alkaline batholithic granites (Neary et al., 1976). In southern Libya the Ben Ghnema batholith exhibits a lateral compositional variation from tonalite and granodiorite to adamellite and granite, identical to the west-to-east variation in the Sierra Nevada batholith (Ghuma and Rogers, 1978), and in Egypt there are intrusions with porphyry-type Cu-Mo mineralization and Kuroko-type massive sulphides (Garson and Shalaby reported in Neary et al., 1976). In Ghana there is a continental-margin sequence of conglomerates, shales, sandstones and carbonates and an ophiolite suite of spilites, pillow lavas and serpentinites (Grant, 1973), whilst in Nigeria there is the complimentary arc belt of calc-alkaline dacites and rhyolites (McCurry and Wright, 1977). In the c.700 Ma old Gariep belt of Namibia there are glaucophane-bearing schists (Kröner, 1975), and calc-alkaline lavas and volcaniclastics were intruded by granite batholiths in a c.1300 Ma old volcano-plutonic arc situated at a cratonic margin in Namibia (Watters, 1976).

On the basis of the above diverse evidence from widely separated regions (at least six nationalities working in ten Pan-African countries), it appears that crustal evolution in most Pan-African belts can very reasonably be interpreted in terms of plate-tectonic processes. However, we must remember that we are constrained by the palaeomagnetic data of Piper (1973) and McWilliams (this volume, Chapter 26) and therefore that most Pan-African oceans must have been relatively narrow. Neary et al. (1976) conclude that the oceanic "closing" across the NE African belts was less than 1000 km.

The data presented in this paper are all consistent with the operation of plate-tectonic processes of various types in Precambrian time. Of course there are some fixist non-uniformitarian viewpoints, according to which certain Precambrian mobile belts formed as in situ, ensialic structures, but significantly these individual models have failed to gain general support.

I think it is important to see the present situation in historical perspective. The lessons of the sea-floor spreading—continental drift scenario of the 1960's were not applied to the understanding of Phanerozoic fold belts until 1969 when J.F. Dewey made his rigorous benchmark analysis of the Caledonide—Appalachian belt. There were few attempts in the early seventies to apply the geological relationships and sequence of events in Cordilleran—Himalayan belts to the rock record in Precambrian mobile belts, but in the last few years an increased understanding of the mode of development of young orogenic belts has clearly given rise to a great many plausible interpretations of Precambrian belts. Evidence of palaeorifts, oceanic crust and ophiolites, trailing-edge sedimentation, volcanic arcs and Cordilleran batholiths, marginal and epicontinental basins, collisional tectonics and sedimentation, eclogites and sutures, is all found in Precambrian mobile belts in the correct spatial association and with geochemical characteristics that are little different from those of modern equivalents. However, there has been far greater success in interpreting high-level Precambrian belts in which sedimentary and volcanic piles, ophiolites and nappes are still preserved, than the deeply eroded belts which contain much sialic basement; this is probably because most Phanerozoic belts, the standard for comparison, have not been highly uplifted and eroded and thus the origin of their root zones is poorly understood. In this way the Pan-African Mozambique and Zambesi and the Grenville belts, with their high proportion of continental basement rocks, have been interpreted (wrongly, in my opinion) as ensialic belts. However, we know that all modern orogenic belts, in both their Cordilleran and Himalayan stages, develop on continental crust — all collisional mobile belts must be expected to expose much old basement at deep erosion levels.

Of course, it is much easier to interpret Proterozoic than Archaean geology in terms of the Wilson cycle. Nevertheless the isotopic and trace-element characteristics of Archaean rocks are consistent with a primitive plate-tectonic process (Jahn et al., 1974; Moorbath, 1976; Tarney and Windley, 1977). But it would be naive to suggest that Archaean and present-day rocks do not have significant differences — this is to be expected from the fact that certain variables were dissimilar, e.g. the heat production rate and the plate growth and spreading rate were probably higher, the angle of subduction zones was shallower, and the mantle was less depleted than now. These kinds of variables will be useful to constrain future models of tectonism and magma genesis.

TABLE 1-I

Stages in earth history

1. 4500—3900 Ma	No rock record preserved, therefore extreme speculation. Formation of proto-crust and proto-lithosphere of unknown thickness. Non-reducing atmosphere and hydrosphere evolve rapidly
2. 3900—2700 Ma	High rate of radiogenic decay, vigorous small-scale mantle convection and new oceanic crust which was subducted to give rise to voluminous calc-alkaline melts with predominantly tonalitic end-products, deformed to gneisses. Extensive rifting in marginal basins gives rise to multiple greenstone belts as proto-ophiolites
3. 2700—2300 Ma	Transitional and diachronous period during which much uplift and erosion of late Archaean thick continental crust gives rise to voluminous clastic debris. Stabilization of crust-lithosphere following aggregation of Archaean proto-plates. Many dykes and stratiform intrusions injected into stabilizing crust
4. 2300—600 Ma	Lower mantle temperatures and slower global spreading rate led to "normal" Wilson-cycle from early Proterozoic. Narrow geosynclines and mobile belts formed by Cordilleran—Himalayan collisional tectonics giving rise to extensive more stable plates
5. 600 Ma to Present	Wilson cycle controls all geological phenomena

Table 1-I summarizes the main stages in Precambrian crustal development in the context of earth history.

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Chapter 2

PRECAMBRIAN TECTONIC STYLE: A LIBERAL UNIFORMITARIAN INTERPRETATION

R. B. HARGRAVES

ABSTRACT

Major uncertainties impeding clarification of the Precambrian geologic record are: (1) its thermal history; (2) the history of the growth of continental crust; and (3) tectonic style. The only certainty is that average surface heat flow has declined.

From an appraisal of current ideas on mechanics of contemporary plate tectonics, it seems that the pull associated with the subduction of cold, negatively buoyant oceanic lithosphere is the dominant force (buoyancy-powered subduction). If this is valid, then subduction is less likely earlier in earth history when heat flow was higher. However, with higher heat flow, the rate of convective circulation in the mantle was faster, and viscous drag forces at the base of lithosphere may have been stronger.

Simple assumptions and calculations suggest a model in which:

(1) Initially, in a vigorously convecting earth, the viscous drag forces were sufficient to drag buoyant crust back into the mantle (viscous drag subduction). No crust survived this recycling, but it enhanced the upward concentration of sialic components in the earth.

(2) With cooling, slowing of convection and the passage of time, a buoyant, possibly continuous global scum-crust of intermediate composition would have developed. This crust was effectively decoupled from the mantle and was recycled as a result of volcanic addition on surface and remelting and removal at the base.

(3) With further cooling crust became coupled to the mantle; the average density of lithosphere increased and ultimately buoyancy-powered subduction of crust could begin. Subduction of segments of the entire lithosphere, where crust was thin, initiated the contemporary plate-tectonic style and growth of the continent-ocean dichotomy. Simultaneously, decoupling and sinking of mantle lithosphere separated from crust caused intracontinental mobile-belt orogeny.

(4) With continued decrease in continental heat flow the strength of the boundary between continental crust and the mantle has increased and decoupling has become more difficult; as a result intracontinental orogeny is now rare or impossible. This leaves only buoyancy-powered subduction of oceanic lithosphere as the predominant contemporary orogenic force.

These four stages are considered to be consistent with the available geologic record and can be roughly equated with the (1) Pre-Archaean, (2) Archaean, (3) Proterozoic and (4) Phanerozoic.

INTRODUCTION

Earth history is recorded in rocks, and the record, although increasingly fragmentary and blurred with age, is now known to extend back to at least 3.6 Ga (Moorbath, 1977). All such rocks to which we have access were formed at or near the surface of the solid earth, and hence the geologic record concerns most directly the history of the crustal environment. It is from the study of rocks that the geologist attempts to infer the past configuration of the crust and the style and intensity of processes operating on it.

The traditional uniformitarian approach to the task is to assume that the earth has always looked and operated approximately as it does today (Kitts, 1963) and this has proved to be extremely effective in advancing the science. When concerned with Precambrian history, however, one is more sensitive to the fact that there was a beginning, about 4.6 Ga ago, and that in all probability the earth looked very different then than it does now. It seems to me to be unlikely that the earth could have evolved completely from its hypothetical hot, relatively homogeneous, beginning to a configuration and dynamic style approximating that of today, in the geologically short time interval between 4.6 and 3.8 Ga. I expect this development to have been slower and to have continued long into the geologic record. It is my prejudice, therefore, to seek evidence of secular change rather than uniformity in Precambrian history.

There are many uncertainties inherent in speculation on the Precambrian; three major problems reviewed in some detail in this chapter concern: (1) the earth's thermal history; (2) the origin of continents; and (3) Precambrian tectonic styles. Thereafter, contemporary ideas pertaining to the forces driving plate tectonics and their relationship to terrestrial heat flow are reviewed. This leads to an evolutionary tectonic model which is consistent with possible (perhaps probable) interpretations of most geological and geophysical constraints.

MAJOR UNCERTAINTIES IN PRECAMBRIAN SYNTHESIS

Thermal history of the earth

Initial state

The rate at which the earth grew by accumulation of matter from the solar nebula influenced its initial thermal state. If slow, the kinetic energy released upon infall of planetesimals could be continuously dissipated by radiation back into space. If accumulation was sufficiently rapid, however, at least the outer layers would melt as appears to have occurred on the moon (Wetherill, 1972; Smith, 1979).

With regard to the pattern of growth, in inhomogeneous accretion models (Clark et al., 1972) it is suggested that the earth accumulated first an iron core and then a silicate mantle. This stands in contrast to homogeneous protoplanet accretion models (Ringwood, 1975; Barshay and Lewis, 1976) from which core and mantle were subsequently differentiated. The thermal implications of these contrasting hypotheses reside in the fact that the gravitational energy released by rapid differentiation of a metallic iron core could heat the entire earth by as much as 2000°C (Birch, 1965; Flasar and Birch, 1973). Such a thermal event would presumably not occur in the inhomogeneous accretion theory.

The possible role of short-lived radioactive nuclides, or lunar capture, in causing rapid heating of the earth near the time of its origin is not clear (Singer, 1977).

Based on the assumption that at least one of these possible heating mechanisms operated during or soon after formation, the majority of scientists (see Ringwood, 1975, p. 550) seem to favour the idea that the earth formed (or soon became) hot, rather than cold. In addition to the possibility of chemical differentiation of the outer layer of the earth (if, by analogy with the moon, it was once molten), a high initial temperature would ensure mantle-wide convection from the beginning as favoured by McKenzie and Weiss (1975).

Thermal evolution

Apart from its initial temperature, the greater abundance of long-lived radioactive nuclides in the earth at its beginning ensured a higher heat generation than now. In this respect, at least, the earth has undoubtedly changed (Hubbert, 1967). This is a principal energy source, however, for processes operating in the crust, and so knowledge of the inventory of heat-producing elements in the earth is of extreme importance.

In addition to estimates using actual analyses of rocks, appraisals of the radioactive element content of the earth have always been influenced by one remarkable relationship: that the apparent total heat flux from the earth was approximately that which would be generated if its complement of heat-producing elements was equivalent to that of average chondritic meteorite (Urey, 1956; Hurley, 1957). Latest estimates of oceanic heat flow, however, have been considerably increased (from 1.46 to 2.2 HFU) due to the recognition of the important role of hydrothermal circulation and heat dissipation at ridges (Sclater et al., 1980). If a balance between heat loss and heat production is still assumed, the earth's heat-producing capacity must be considerably greater than that of chondritic meteorites (Parsons and Richter, 1979); alternatively, some of the heat being lost is primordial.

Based on the chemistry of all crustal and upper mantle rocks to which we have access, there are cogent reasons to argue that the earth could not, in any case, be exactly equivalent to chondrites (Gast, 1960). The one that is most pertinent from a heat production point of view concerns the ratio of potassium to uranium. In chondritic meteorites this ratio is approximately 80.000:1 whereas the average value in all terrestrial rocks analyzed is 10.000:1. As Wasserburg et al. (1964) pointed out the heat flow from the earth can also be matched by assigning appropriate amounts of K and U in

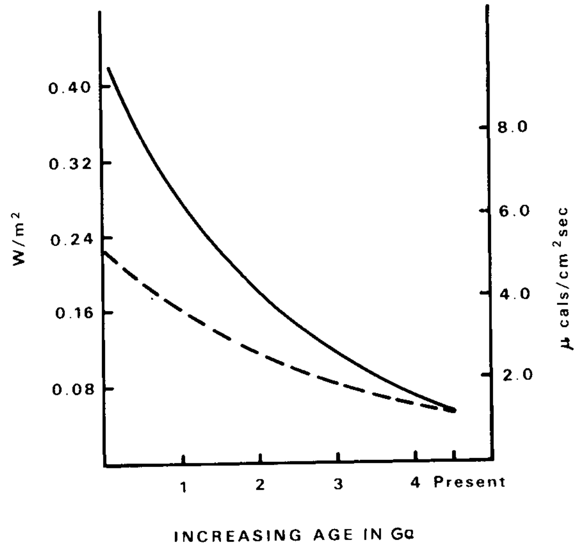


Fig. 2-1. Variation of average terrestrial heat flow (assumed to be 1.4 HFU) with time in a model of the earth with an initial temperature sufficiently great to permit convection throughout the mantle. The solid curve is for a model with chondritic abundances of radioactive elements; the broken curve is for one with heat-flux equivalent to present values, but with a K/U ratio derived from measurements of crustal rocks — the model of Wasserburg et al. (1964). Figure adapted from fig. 7 of McKenzie and Weiss (1975).

the ratio of 10^4. Because of the shorter half-life for the decay of ^{40}K to ^{40}Ar and ^{40}Ca, however, there is a very marked contrast in the heat generation earlier in earth history implied by the two models (Fig. 2-1). Assuming thermal equilibrium a chondritic earth had eight times the present heat flow 4.5 Ga ago, whereas according to Wasserburg et al. (1964) the flow then was only four and a half times greater. The consensus seems to be that the earth's radioactive element inventory is best represented by the Wasserburg model with its lower extrapolated heat-flow curve (Fig. 2-1), but it is acknowledged that we do not know the exact value and distribution of radioactive heat sources in the earth, even today.

Continental accretion

The hypothesis that continents had slowly grown by lateral accretion was applied by J. Tuzo Wilson (1954) to the evolution of North America, with its Archaean (> 2500 Ma) Superior Province bounded to the southeast first by the Grenville Province (1200—1000 Ma) and, in turn, by the Appalachian orogenic belt (500—200 Ma). At that time the concept of continental accretion proved extremely popular, particularly as it was construed as evidence precluding continental drift (e.g. see Chadwick, 1962).

The Gondwana continents, where the best evidence for continental drift has always been recognized, likewise were the least consistent with this concept of growth around the stable continental nuclei of today. In fact Hurley and Rand (1973) have since demonstrated from radiometric data that the distribution of Precambrian orogenic belts in South America and West Africa are such as to almost require continental drift. They show that while individual orogenic belts of the same age in both continents are sharply truncated by the present coast lines, they would be continuous if the Atlantic were closed.

Isotopic arguments

The more directly geochemical basis for inferring accretion of continental crust stems from the pioneering work of Hurley et al. (1962) concerning the evolution of $^{87}Sr/^{86}Sr$ ratio in crustal versus mantle systems. Hurley et al. (1962) showed that most granites in orogenic belts had mantle-type initial Sr ratios, precluding their derivation by remelting of old Rb-enriched pre-existing crust, and supporting the view that continents had grown by primary additions from the mantle.

The considerable research since then (e.g. Hurley and Rand, 1969) has abundantly confirmed the Sr-isotopic relationships found by Hurley et al. (1962). The Rb/Sr in the continents, however, appears to decrease with depth and seems to approach the mantle value (~ 0.04) towards the base of the crust (Zartman and Wasserburg, 1969). Thus the Sr isotopic evolution of lower crustal rock may be little different from the mantle, and on this evidence alone, an upper mantle versus lower crustal source for a granite with low initial $^{87}Sr/^{86}Sr$ cannot always be determined.

The ratios of the isotopes ^{206}Pb and ^{207}Pb to ^{204}Pb, and ^{143}Nd to ^{144}Nd likewise increase with time as a function of the U/Pb and Sm/Nd ratios in the host. For U/Pb it appears that, relative to the mantle, this ratio tends to be higher in the upper crust but significantly lower in the lower crust (Zartman and Wasserburg, 1969); the Sm/Nd ratio, on the other hand, is much lower in the crust than in the mantle and shows little variation throughout the crust (De Paolo and Wasserburg, 1979). Thus, on the basis of lead and neodymium isotope ratios, the mantle versus lower crust source of a granite with low initial Sr ratios may possibly be distinguished. Moorbath (1977, 1978) and Moorbath and Taylor (this volume, Chapter 20) infer that these data together indicate that the volume of continental crust has grown progressively (if episodically) and irreversibly with time. They relate the differentiation and accretion of continental material solely to processes associated with subduction of oceanic lithosphere and the associated calc-alkaline igneous activity at island arc and continental cordillera.

Hurley and Rand (1969, 1971) synthesized all K-Ar and Rb-Sr radiometric and isotopic data on the crystalline basement rocks of continental

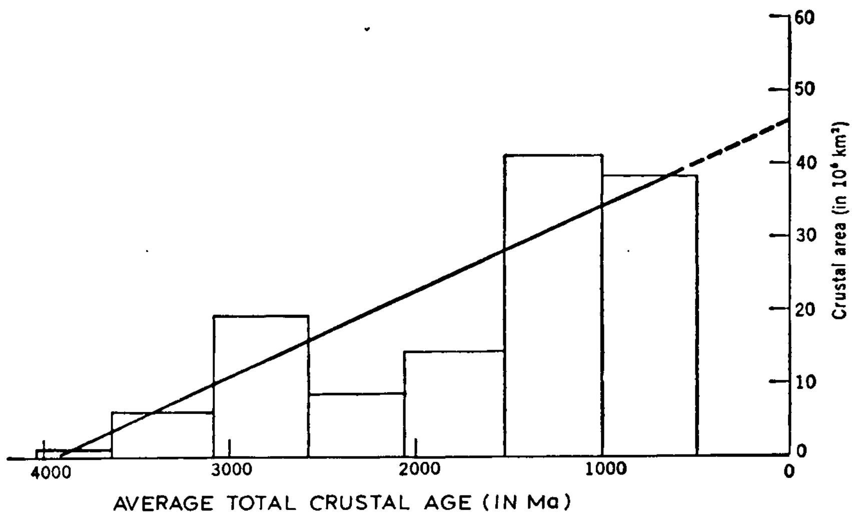


Fig. 2-2. Histogram showing the area of continent underlain by rocks within successive intervals of total crustal age: figure 13 of Hurley and Rand, 1969. The pattern is interpreted to indicate an accelerating generation of crustal material amounting to 20 km^2 per million years.

crust as a function of their time of formation, in 450 Ma time blocks (Fig. 2-2). Isotopic evidence suggesting the presence of admixture of reworked, old sialic crust is so common in Phanerozoic orogenic belts, however, that they considered only the Precambrian terrains. Ignoring the hints of episodicity in the graph, Hurley and Rand (1969) drew the straight line shown in the figure and inferred that the increase in area of continental crust during the Precambrian accelerated at a rate of 20 km^2 /Ma. The integrated area of continent through geologic time derived from this relationship is illustrated in Fig. 2-3.

Today orogeny is attributed to the interaction of lithospheric plates, and it is widely presumed that plate tectonics is a consequence of thermally driven convective motions in the mantle. One could then equate intensity of plate-tectonic activity with rate of heat generation and dissipation (Burke et al., 1977). But the rate of heat generation has declined with time and so, to a first order, the rate of plate-tectonic turnover should have declined with time. Thus, if plate tectonics generates continental crust, the growth of continental crust should slow as heat generation has slowed (Fig. 2-3). The accelerating continental growth curve is completely antithetic to this model and is more consistent with a slowing rate of continent *destruction* rather than irreversible construction.

Recent geochronologic work (see Hart and Allègre, 1980) indicates

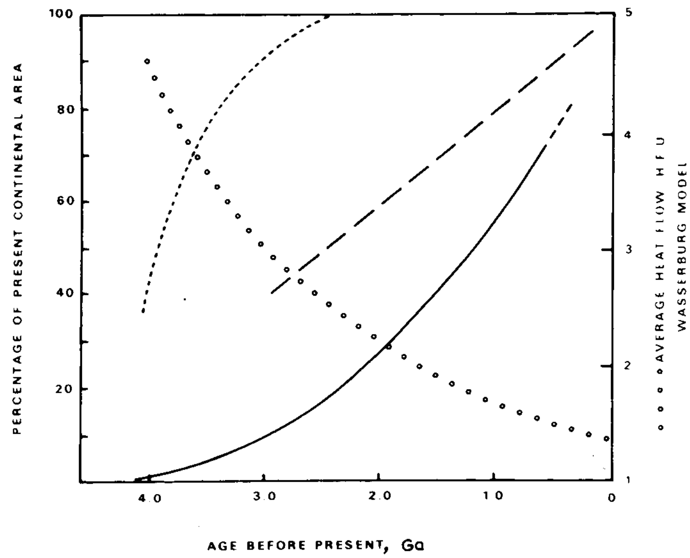


Fig. 2-3. The solid curve indicates the integrated area of continental crust as a function of time, expressed as a percentage of today's total, according to the accelerating growth rate of Hurley and Rand (1969) shown in Fig. 2-2. For comparison, the long-dashed line portrays a relatively linear growth curve, while the short-dashed curve portrays a history in which segregation of continental crust was essentially complete by 2.5 Ga before Present. The curve marked by open circles (scale on the right) indicates the declining average heat flow according to the model of Wasserburg et al. (1964).

generally that reworked older sialic crust is more abundant in younger Precambrian mobile belts than was originally thought and the validity of Hurley and Rand's accelerating growth curve may be disputed. But their view is more in accord with those who consider that continents have grown throughout geologic time (Moorbath, 1978, p. 401).

Crustal recycling

Armstrong (1968) and Armstrong and Hein (1973) pioneered a crustal recycling model that is more consistent with the apparent progressive growth curve and simultaneously satisfies constraints imposed by the record of both strontium and lead isotopes. Armstrong proposed that the bulk of the material composing continents and hydrosphere was segregated in approximately their present day volumes very early in earth history (at least by 2.5 Ga). Since then continued mixing and isotopic equilibration of crust and mantle has occurred via the mechanism of plate tectonics.

In a related manner O'Nions et al. (1979, 1980) have successfully modelled the evolution of K-Ar and Sm-Nd (in addition to Rb-Sr and U-Pb) in mantle and crust—atmosphere—oceans by means of a simple two-reservoir model (mantle and a 50 km outer layer of the earth which includes the continents). Two-way transport of incompatible elements between these two reservoirs is proposed with declining time-dependent flux rates, each with different time constants. Acceptable models were those in which: (1) only 50% of the mantle, by mass, contributed to crustal formation (the gradual accumulation of K and the other lithophile elements in the outer layer); (2) the return flux transport coefficient is initially much greater than that for the outward flux, but to provide for the segregation of the bulk of the continental crust between 3.5 and 3.0 Ga, as is assumed in their model, the two transport coefficients become equal about 3.0 Ga ago. The residence (process) time thereafter becomes long in both reservoirs, and today is very much longer in the outer crustal layer (Fig. 2-4).

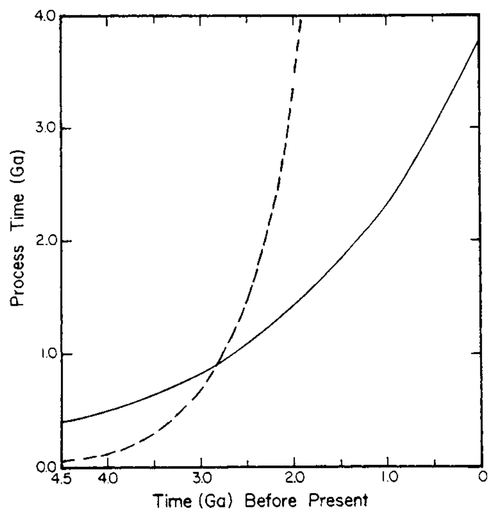


Fig. 2-4. Process (or "Residence") time for potassium in the mantle (solid line) versus 50 km crustal layer (broken line), according to the two-reservoir recycling model of O'Nions et al. (1980). Adopted from figs. 7 and 8 of O'Nions et al. (1980).

The mechanism for recycling of continental material invoked by Armstrong and O'Nions is by the subduction of sediment derived from the continent and deposited on oceanic crust. The feasibility of this process is doubted by Moorbath (1977, pp. 173—174), but there is evidence that it is actually occurring, to some extent at least, at trenches today (Whitford et al., 1977; see also Fyfe, this volume, Chapter 22). It seems unlikely, however, that this erosional mechanism alone could account for the complete absence of crustal remnants older than 3.8 Ga, or be adequate for the complete and rapid recycling called for in the O'Nions' models. If no other mechanism is possible, direct subduction of continental crust (if it is thin or of slivers, if it is thick) would be much more effective, and the feasibility of this is discussed in the next section. By whatever means, the scale of the recycling called for in these models poses serious questions as to how crust is mixed and isotopically homogenized with the mantle reservoir. The mantle heterogeneities recorded today may well be a result of this mixing process (Richter and Ribe, 1979).

The success of these recycling models in explaining the isotopic systematics pertaining to continental growth convinces me that recycling of continental crust must be occurring, although at a much slower rate now than earlier. As Armstrong (1968) puts it: the residence time of continental crust is increasing.

In summary, if the segregation of sialic material is directly related to convective mantle circulation, then isotopic data on continental accretion presents a paradox: one must either: (a) invoke a mechanism for thorough mixing and isotopic re-equilibration of recycled continental sialic material, a process which slows with time; or (b) find a mechanism to delay the start of irreversible continental segregation until 3.8 Ga ago, and then have it accelerate (albeit episodically). The fact that heat flow has declined with time inclines me to favour the first alternative.

Precambrian tectonic style

A current controversy pertaining to uniformity and the Proterozoic tectonic style is sparked by interpretations of the palaeomagnetic record.

The more aggressive uniformitarianists (e.g. Burke et al., 1976a, b, 1977) argue that all geologic and palaeomagnetic data are consistent with the continuing operation of the Wilson cycle (i.e. the opening and closing of ocean basins, with associated rifting and orogeny) throughout the Proterozoic at least.

Most appraisals of currently available palaeomagnetic data however (McElhinny and McWilliams, 1977; *see also Irving and McGlynn, this volume, Chapter 23; McWilliams, this volume, Chapter 26, ed.*) find that the Precambrian record is not consistent with development of orogenic belts *solely* by closure of oceans. For North America (Irving and McGlynn, 1976), Africa (Briden, 1976) and Australia (McElhinny and Embleton, 1976) apparent polar wander "swathes" can be constructed which accomodate all the palaeomagnetic data from several dispersed cratons. This, in turn, implies that the younger orogenic (or mobile) belts which now separate the older cratons on these continents were either entirely ensialic or resulted from the opening and (geographically precise) reclosing of oceans no wider than about 1500 km (McElhinny and McWilliams, 1977). Piper (1976) has proposed the existence of but one super-continent in the Precambrian, but others (McGlynn et al., 1975; Irving and McGlynn, 1979; *see also Irving and McGlynn, this volume, Chapter 23, ed.*) conclude that there are significant differences between the apparent polar wander paths for at least North America (Laurentia) and Africa, implying their independent drift.

The palaeomagnetic record suggests therefore that although plate tectonics was operating to some degree, continents were even more coherent in the Proterozoic than they are now and that an additional mechanism operated which could lead to ensialic orogenesis. If these data and conclusions are valid, then some differences in tectonic style are indicated even during the Proterozoic. To deny this is to reject the palaeomagnetic data or claim that its resolution is presently inadequate to demand such departures from uniformity (Burke et al., 1976a; *see also Irving and McGlynn, this volume, Chapter 23, ed.*).

CONSTRAINING ASSUMPTIONS

With respect to the controversial topics discussed above I will assume that:

(a) The earth was hot enough for mantle-wide convection from the beginning and that its radioactive heat-generation history has been at least as high as that implied by the Wasserburg model.

(b) Mechanisms exist or existed for recycling, through the mantle, of once-segregated sialic material and the overall rapidity of this recycling has slowed with time.

(c) Ensialic orogenesis did occur during Proterozoic time.

In good uniformitarian style, however, the next section will review current understanding of contemporary tectonic processes and use this as a basis for speculating about the past.

CONTEMPORARY GEODYNAMICS

We now accept that the earth's crust is made up of a mosaic of comparatively rigid lithospheric plates whose relative mobility can be described in simple geometric terms (Morgan, 1968). We understand that the motion and interaction of these lithospheric plates can explain much of the contemporary volcano-tectonic process and the configuration of continental and oceanic crust. In our fraternal excitement at this recent discovery, however, we geologists should not forget that geophysicists are still uncertain as to exactly why plates move (see Kerr, 1978).

It is generally acknowledged that plate motion is a direct or indirect response to convective motions in the interior of the earth. The simple models pictured by Griggs (1939) and Hess (1962), with convective streams rising beneath ridges and descending at trenches, dragging crust with them, provide a qualitatively plausible mechanism for the forces which drive plates. One intuitively suspects that the detailed reality, whatever it may be, is merely some variation of a theme whereby the viscous drag of convecting mantle moves the plates. Recent attempts to model mantle convection, to determine the plate-driving forces and to ascertain how closely plate motions are coupled to mantle convection, indicate that the simple concepts are likely to be very misleading.

Scale of mantle convection

There has been considerable debate as to whether the whole mantle or only the upper mantle convects (McKenzie, 1969; Turcotte and Oxburgh, 1972). Depending upon whether or not the seismic discontinuity at ~ 650 km depth is considered to be a phase change only or to mark a compositional change in addition, whole mantle convection can occur through the boundary, or only in layers above and below (McKenzie and Weiss, 1975). If, as many now seem to believe, whole mantle convection can and does occur (O'Connell, 1977; Davies, 1977; Stevenson and Turner, 1979; Elsasser et al., 1979), there remains considerable uncertainty as to its form. Morgan (1972) has suggested that hot plumes rise from the core—mantle boundary and spread radially at the base of the lithosphere (mantle hotspots). As an extreme opposite Gough (1977) has suggested that departures of the geoid from a spheroid exhibit the symmetric pattern of a tennis ball and that this pattern may reflect single-cell whole-mantle convection.

It should be noted, however, that the geochemical earth-evolution models of both O'Nions et al. (1979, 1980) and Wasserburg and De Paolo (1979) satisfy the isotopic constraints best if only about half, rather than the whole mantle has differentiated to form crust. If true, this would suggest that the upper and lower mantle convect independently, with little material transfer between them. The scale and form of the convection in the earth are obviously not well understood.

The possible occurrence of two superimposed scales of convection in the upper mantle above 700 km, one whose horizontal scale is large compared with 700 km and of which plates are part, and one whose scale is comparable with or smaller than 700 km, has been proposed by McKenzie and Weiss (1975), Richter and Parsons (1975), McKenzie and Richter (1976), and Richter (1973). Richter and Parsons (1975) proposed that beneath fast moving plates this small-scale convection could take the form of two-dimensional rolls, as illustrated in Fig. 2-5. The question of the existence of small-scale convection in some form is important as it is a mechanism for providing heat to the base of the lithosphere everywhere, as seems to be called for by analyses of heat-flow data (see below).

Above all, while plate tectonics today constitutes the most obvious evidence for thermal convection in the mantle, it is not yet clear whether or not there is any direct relationship between the pattern of surface activity (ridges and trenches) and deep mantle flow (Stevenson and Turner, 1979); there is certainly no easy or obvious correlation.

Heat flow

The pattern of heat flow from the earth today shows high values at ridges and young orogenic belts with relatively low values in areas of old

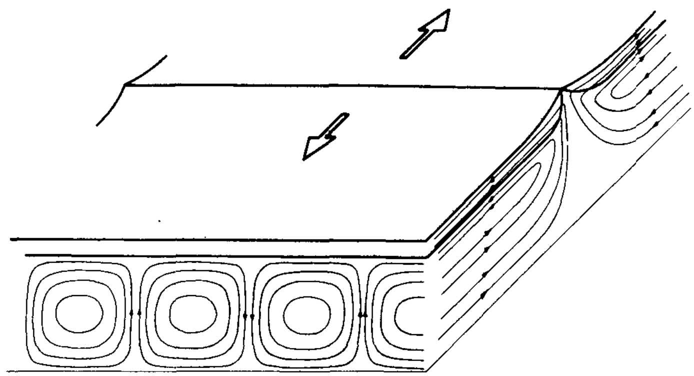


Fig. 2-5. Block diagram illustrating the small-scale longitudinal convective rolls superimposed on large-scale flow in the vicinity of a ridge, as proposed by Richter, 1973 (copy of his fig. 12).

oceanic (100–150 Ma) and old continental (> 1000 Ma) crust (Polyak and Smirnov, 1968; Pollack and Chapman, 1977a). Allowances for the heat lost at ridges by hydrothermal circulation give an average heat flow for the oceans of 92 mW/m^2 [= 2.2 HFU] (Sclater et al., 1980). Together with the continental average of 59 mW/m^2 this gives an average for the whole earth of 79 mW/m^2 (Parsons and Richter, 1981).

The remarkable correlation of both heat flow and ocean-floor elevation with age of oceanic crust (Sclater and Francheteau, 1970; Sclater et al., 1971) testifies to the thermal importance of oceanic lithosphere creation and cooling. This process is estimated to account for as much as 60% of the total heat flux from the earth today (Sclater et al., 1980).

There are discrepancies, however, between the elevation of old oceanic crust predicted on the basis of a simple cooling model and that observed (Parsons and Sclater, 1977; see Fig. 2-6). These can best be explained if there is a ubiquitous steady-state flux from the asthenosphere to the base of the oceanic lithosphere (Sclater, 1972; Parsons and Richter, 1981). The immediate origin of the heat flux to the base of the lithosphere is not clearly understood; it might be transported by small-scale convection (Parsons and McKenzie, 1978; Parsons and Richter, 1981), be generated by viscous dissipation—shear heating, associated with flow in the upper asthenosphere (Schubert et al., 1978) or by time-averaged hot-spot heating (Crough, 1979). In areas of old oceanic crust, where the lithosphere has cooled and approaches apparent thermal equilibrium, the heat flow still exceeds 1 HFU (= 41.8 mW/m^2), suggesting that the flux to the base of the oceanic lithosphere must, according to Sclater (1972), approach 1 HFU. Crough (1975) and Crough and Thompson (1976) have proposed thermal

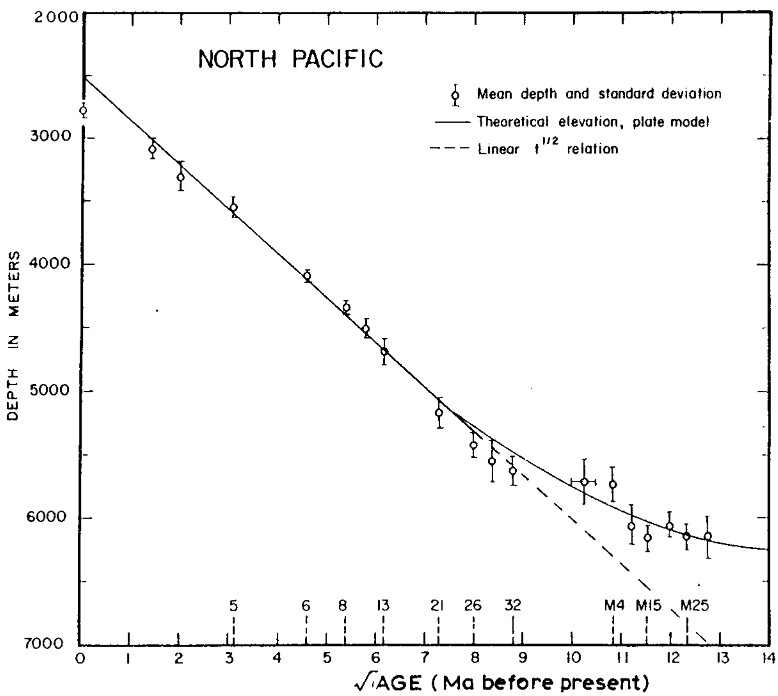


Fig. 2-6. Mean ocean depth (and standard deviations) plotted versus the square root of age for the North Pacific: fig. 8A of Parsons and Sclater (1977). Note the departure from the linear $t^{1/2}$ depth dependence for oceanic crust older than about 75 Ma.

models for cooling of both oceanic and continental lithosphere and conclude that the flux from asthenosphere to lithosphere is approximately the same everywhere and averages 0.5 to 0.6 HFU. If this lower value is accepted, it implies that the ubiquitous asthenosphere—lithosphere flux contributes about 30% of the total heat flow from the earth.

The third major contribution to the earth's heat flow is from radioactive elements concentrated in the lithosphere. In oceanic areas the maximum contribution, even from old crust and lithosphere, is not likely to exceed 0.2 HFU (Pollack and Chapman, 1977a) but the concentration of radioactive elements in continental crust is of course much higher. Allowing for the marked concentration of incompatible heat-producing elements towards the top of continental crust (Lachenbruch, 1970) and varying degrees of erosion of this enriched layer, it is estimated that this source contributes from 0.5 to 0.6 HFU to the continental heat flux (Smithson and Decker, 1974; ~ 40% of the observed heat flux, according to Pollack and Chapman, 1977a). As continents constitute roughly 40% of the earth's surface, this is equivalent to about 0.20 HFU over the whole surface or 10% of the overall average, 1.9 HFU (Sclater et al., 1980).

There are thus three principal components in the contemporary overall heat flux (see also Bickle, 1978):

(1) Formation and cooling of oceanic lithosphere ~ 60%.

(2) Sublithosphere mantle heat flow ~ 30%.

(3) Radioactive elements in continental crust ~ 10%.

Simple extrapolation of these three components back to the Precambrian, when the radiogenic heat production was higher (Fig. 2-1), would imply higher geothermal gradients, thinner lithosphere and more intensive plate tectonic turnover (e.g., Burke et al., 1977).

In an incisive paper Bickle (1978) constructed a quantitative thermal model along these lines and considered the relative importance of the three components in the Archaean. His conclusion (1978, p. 311) that heat loss "by some process analogous to plate tectonics" was significant in the Archaean depends critically upon his assumption of a *P-T* gradient of 20–25°/km as being representative of equilibrium gradients through cratons at that time. This value, which is obtained from a study of metamorphic mineral assemblages in Greenland, is considered to be representative of Archaean high-grade metamorphism. While acknowledging the uncertainty he argues that regional metamorphic gradients place maximum limits on the equilibrium gradients. The value he chooses (20–25°/km) is so low that he feels his conclusions — favouring plate tectonics — are inescapable. He states (1978, p. 311) "if no heat were lost by plate processes at 2.8 Ma, the equilibrium heat flow into the base of the lithosphere would be about 3.4 HFU giving a thermal gradient of 50°C km^{-1} in the continents" — and he considers this unacceptable. It is noteworthy, however, from Grambling's (1979) compilation of Precambrian metamorphic *P-T* data that an average metamorphic gradient of around 50°C/km (rather than 20–25°C/km) is very reasonable.

Burke et al. (1977) also assume that higher heat generation in the mantle — requiring more vigorous mantle convection — inevitably translates into more rapid plate-tectonic turnover of crust but, as discussed below, this is not necessarily valid.

Plate-driving force

In all recent analyses of the relative magnitude of the various forces which can be imagined to contribute to the operation of plate tectonics, the negative buoyancy or "pull" of the cold oceanic lithosphere is the major (Harper, 1975; Forsyth and Uyeda, 1975; Richardson et al., 1976, 1979; see Fig. 2-7), if not predominant, force (Richter, 1977; Chapple and Tullis, 1977; Richter and McKenzie, 1978; Parsons and Richter, 1979). Viscous drag at the base of the lithosphere (which was intuitively, to me at least, the favoured mechanism) appears to contribute little under oceanic areas and even causes a noticeable retardation under continents. The push from

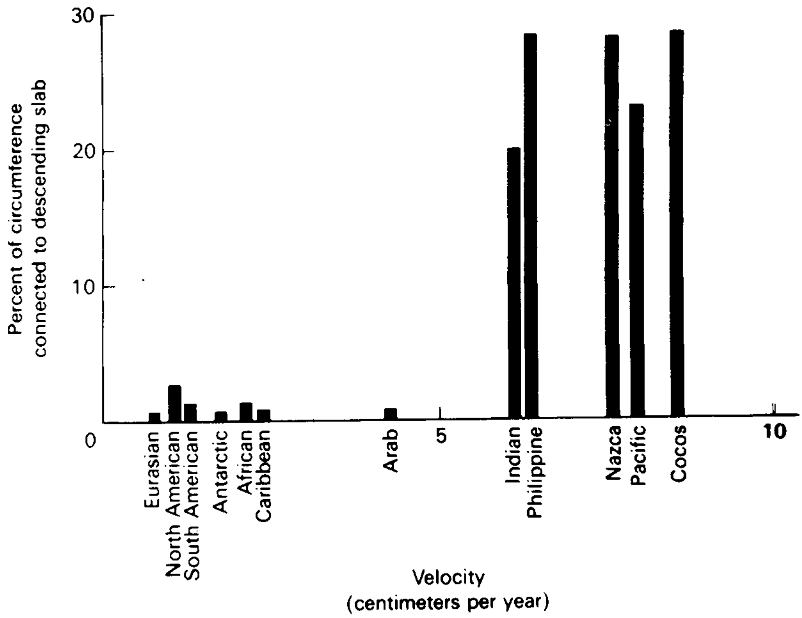


Fig. 2-7. The percentage of plate circumference connected to the downgoing slab versus the absolute average velocity of plates (from Forsyth and Uyeda, 1975). The positive correlation between plate velocity and fractional length of subducting boundary (or trench) is interpreted to indicate that pull of the negatively buoyant subducting slab is the dominant driving force for plate tectonics.

ridges is likewise small according to Chapple and Tullis (1977) but comparable to the pull at trenches according to Richardson et al. (1976, 1979).

One might conclude from this that all that is required to start subduction is for oceanic lithosphere to grow by cooling until some critical value of negative buoyancy is achieved. This is clearly not the case, however, because the age of crust being subducted today ranges from less than 5 to 160 Ma (Deffeyes, 1972; Bickle, 1978) and Atlantic Ocean crust even older than this has not even started to founder. Some combination of these forces (ridge push, viscous drag, etc.), and perhaps circumstances such as convergence and overriding of ridge by continent, must suffice to trigger subduction and cause the spectrum of subducting-crust ages. Once underway, the inversion of basalt to eclogite at 70—100 km depth will overcome any buoyancy problem. Nevertheless, Molnar and Atwater (1978) have pointed out that Cordilleran-type tectonics, with trench at the continental edge, is occurring today only where the subducting oceanic lithosphere is *less* than 50 Ma old, whereas the development of island arcs and clear evidence of back-arc spreading are evident only where the subducting crust is *older* than 50 to 100 Ma. These facts testify to the importance of negative buoyancy in dictating the style of oceanic lithosphere subduction, and if it is accepted that, on a world-wide basis, this is the major force driving this process, then this view has important ramifications with respect to the Precambrian.

Buoyancy

Given that mantle convection occurs, today's dichotomy of oceanic and continental crust reflects the fact that oceanic lithosphere can subduct, while continental lithosphere apparently cannot. Granite is much less dense than basalt, so it is intuitively reasonable that only oceanic crust should be consumed. But basalt itself is also much less dense than peridotite, i.e. has positive buoyancy, and hence an important role has been assigned to the inversion of basalt to eclogite in promoting subduction of oceanic crust (Ringwood and Green, 1966; Ringwood, 1975). This inversion, however, is not likely to occur until the crust has reached depths of 70 to 100 km (Ringwood, 1975) and hence cannot play a role in initiating subduction.

The negative buoyancy required for subduction arises primarily as a result of cooling of oceanic mantle-lithosphere (McKenzie, 1969; Molnar and Atwater, 1978; Molnar and Gray, 1979) and in fact, according to Oxburgh and Parmentier (1977), a requisite amount of lithosphere growth by cooling has to occur in order even to overcome the positive buoyancy of both the crust and complementary depleted mantle.

Using a value of 2.7 g/cm^3 for the density of continental crust, McKenzie (1969) showed that a thickness of at least 4.5 km would more than compensate for the negative buoyancy of "cold" lithosphere 50 km thick. Today one might question the assumption of a 2.7 g/cm^3 density for average continental crust and also the thickness for subcontinental lithosphere. Any increase in density of the crust or thickness of subcrustal lithosphere should favour an increase in the minimum crustal thickness required for net positive buoyancy. The subductibility of the earth's mechanical boundary layer is, in present-day terms, a function of mantle-lithosphere thickness and its negative buoyancy combined with crustal thickness and its positive buoyancy. Relationships between these are qualitatively illustrated in Fig. 2-8 in which the negative buoyancy of mantle lithosphere with respect to asthenosphere has been assumed to average 0.08 g/cm^3 (i.e. average asthenosphere $\rho = 3.22$, average mantle lithosphere $\rho = 3.30$; see Crough and Thompson, 1976, but note that it is the difference which is important, not the absolute value). It shows the isostatic boundaries of crustal and lithospheric thickness for various average crustal densities. Consistent with the contemporary situation, Fig. 2-8 illustrates that oceanic crust ($\rho = 3.0$) 5 km thick can theoretically subduct when lithosphere beneath exceeds 20 km in thickness (i.e. achieves net negative buoyancy), whereas continental crust ($\rho = 2.70$) 35 km thick requires lithosphere $>$ 220 km thick. Molnar and Gray (1979) give a more rigorous derivation of these relationships,

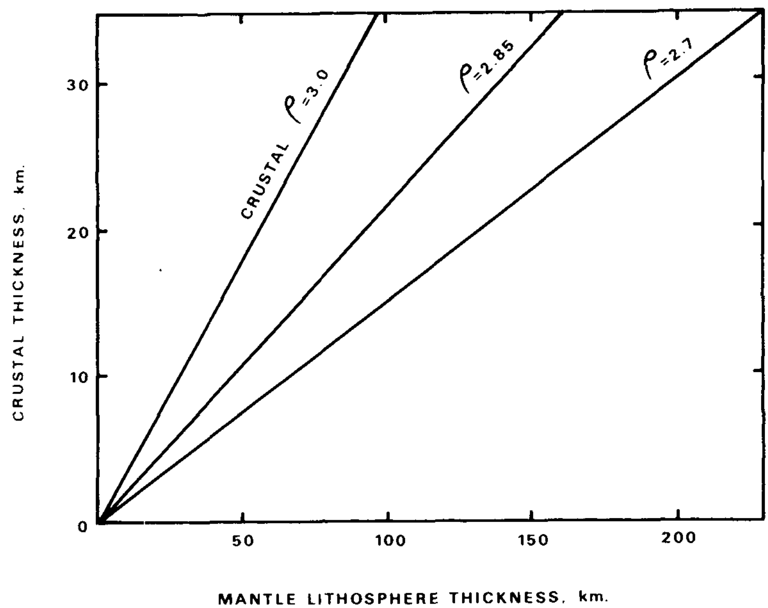


Fig. 2-8. Thickness of positively buoyant crust ($\bar{\rho}_{lc} < \rho_a$) relative to negatively buoyant mantle lithosphere ($\bar{\rho}_{lm} = 3.30; > \rho_a$) required for the entire lithosphere to be isostatic with asthenosphere ($\rho_a = 3.22$). Three curves are drawn, for average crustal densities of 3.0, 2.85, and 2.7 g/cm^3.

concluding (p. 59): "The calculations do show, however, that we cannot eliminate the possibility of subducting a large fraction of continental crust, if it could be detached from the upper part."

IMPLICATIONS FOR THE PRECAMBRIAN

Buoyancy-powered plate tectonics

With higher heat generation in the past it is inescapable that convective overturn of mantle was more active. Plate tectonics results from the fact that today thin, positively buoyant oceanic crust is transported along with the cold overturning mantle (subduction of lithosphere). This overturn of light crust is not necessarily inevitable, however, as evidenced by the convective circulation of water beneath an ice-covered (positively buoyant crust) pond. But if contemporary plate tectonics (in which overall negative buoyancy plays a crucial role) is identified with convective overturn, and this mechanism is required to dissipate more heat in the past than it does now, either a greater length of spreading ridge, more rapid spreading, or some combination must be invoked (Burke et al., 1977). The first implies less average distance and the second less average time, from ridge to trench. In either case less negative buoyancy could be achieved, and in this sense subduction of crust is *less* likely to occur rather than be accelerated (Molnar and Atwater, 1978; Hargraves, 1978).

It is conceivable that some "critical" world-wide average degree of negative buoyancy is required for plate tectonics to operate and high heat flow may at one time have precluded it. Currently subducting oceanic lithosphere averages about 80 km in thickness and is about 75 Ma old. This thickness is commensurate with a heat flow of about 1.4 HFU (Pollack and Chapman, 1977b; Chapman and Pollack, 1977). If, for example, 2.8 HFU were the maximum average heat flow value (minimum lithosphere thickness ~ 30—40 km, ~ 20 Ma old) permitting buoyancy-powered subduction then plate tectonics, as we understand it, could not have started until long (at least 1500 Ma, Fig. 2-1) after the earth formed.

Subduction by viscous drag

With higher heat flow mantle convection was undoubtedly more vigorous in early earth history and the force due to viscous drag at the base of the lithosphere could have been much stronger. Perhaps it was sufficient to drag down even positively buoyant lithosphere (Hargraves, 1978).

McKenzie and Weiss (1975, p. 152) attempted to evaluate the magnitude of the mechanical stresses that would be associated with their small-scale convection under conditions of higher temperature and more rapid circulation. With estimated convective velocities up to 10 times greater than now and initial mantle temperatures 200° higher, they conjecture that viscous stresses on the base of the lithosphere would be in the range 5×10^7 N/m^2 (~ 500 bar). These, they point out, would be sufficient to cause fracturing and shearing of the lithosphere. Could they, however, literally pull positively buoyant lithosphere consisting principally of basaltic crust back down into the mantle?

Consider a positively buoyant lithospheric crust of thickness t and density ρ_c transported on asthenosphere of density ρ_a, flowing with a velocity such that it could exert a viscous drag stress, σ, on stationary lithosphere (Fig. 2-9a). The gravitational force due to the positive buoyancy of the crust is $gt(\rho_a - \rho_c)$ and unless this exceeds the viscous force σ, the crust will be dragged down into the mantle.

The equilibrium relationships between t, σ and $\Delta\rho$ are illustrated in Fig. 2-9b. It can be seen that with a viscous drag of 5×10^7 N/m^2 a basaltic crust ($\Delta\rho = 0.25$), if thinner than 20 km, would be subducted. If the stress was only 100 bar, any such crust thicker than 4 km would float.

If therefore, early in earth history, convective heat flow was high, the lithosphere was thinner and probably consisted largely of the basaltic crust, it is possible that, despite its positive buoyancy, such a crust could have

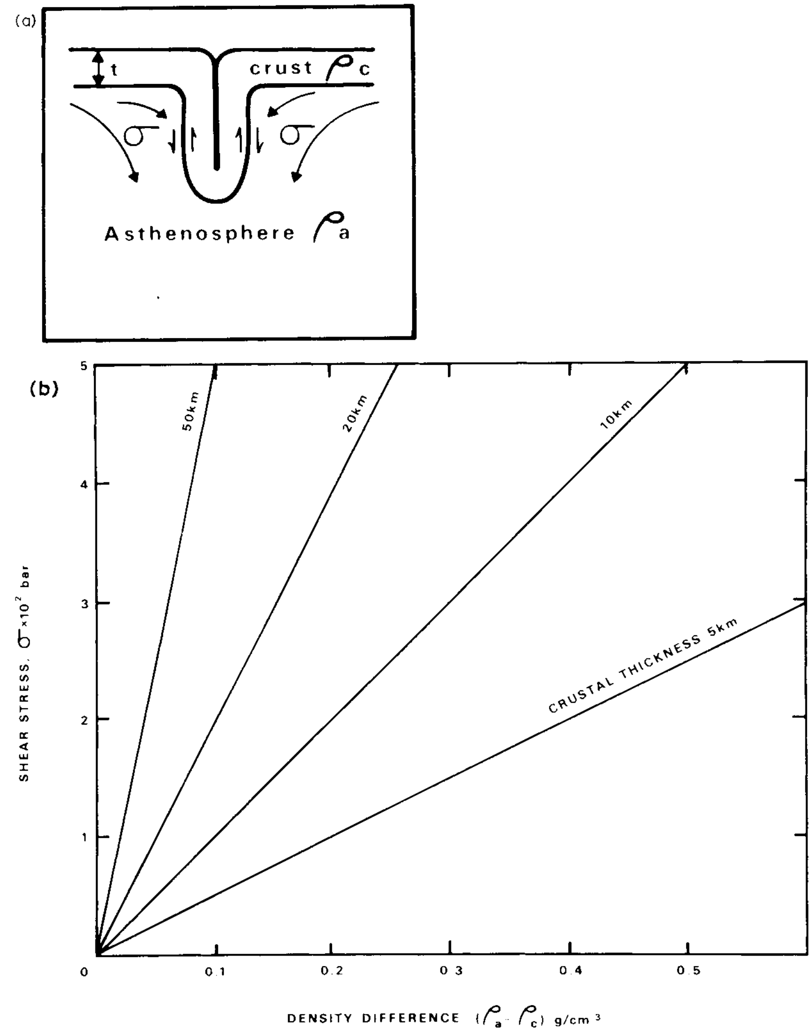


Fig. 2-9. a. Model of crust (= lithosphere) ρ_c, thickness t, positively buoyant with respect to asthenosphere (ρ_a), subject to downward dragging shear stress σ, due to convecting asthenosphere.
b. The magnitude of the viscous drag shear stress (σ) required to compensate the buoyancy of crusts (ρ_c) of varying thickness, as a function of the magnitude of the buoyancy difference ($\rho_a - \rho_c$). For example with a viscous drag shear stress of 500 bar ($= 5 \times 10^7\,N/m^2$) a basaltic type lithosphere crust ($\rho_a - \rho_c \simeq 0.25$) thinner than 20 km would be dragged down.

been dragged down sufficiently for the basalt-eclogite inversion to take place and render the slab negatively buoyant. The surface of the early earth may then have consisted of a mosaic of transient basaltic plates, circulating with a vigorously convecting asthenosphere.

With time, however, heat flow decreases and, assuming constant viscosity, the drag stress decreases. The subductibility of lithosphere increasingly depends upon its thickness and $\Delta\rho$. Any growth of cold mantle lithosphere beneath the crust would decrease $\Delta\rho$, making it more subductible (for a given shear stress), but accumulation of more sialic crust by second-cycle melting above the downwelling zones would tend to counteract this. Lithosphere dominated (with regard to density) by more sialic crust would be buoyant and survive as scum.

The evolution from viscous drag subduction to the buoyancy-powered subduction of today, and whether or not the change was continuous, may well have depended upon the rate of differentiation of sialic material and its distribution. The simplest interpretation of the record would be that prior to 3.8 Ga whatever sial was differentiated was completely recycled, and only at about that time did it start to survive. Recycling of the sial requires re-equilibration with the mantle, which may involve or entail remelting, but it does not necessarily require complete rehomogenization with the mantle. The manner in which this recycling is accomplished is most uncertain and direct subduction of sialic crust, while it may indeed be envisioned if the crust is thin and mantle lithosphere beneath is thick (see below), is not necessarily the only mechanism possible. The tectonic — and recycling — style in the early earth may well have been dictated by the compositional evolution of the protocrust and its configuration.

Evolution of the protocrust

There seems to be no doubt that the outer part of the moon was or became molten early in its history (> 4.0 Ga), and fractional crystallization of this layer differentiated the anorthositic lunar crust (Smith, 1979). Without any equivalent early terrestrial record the moon is our best guide, which suggests that some degree of melting and differentiation of the outer part of the earth (prior to and independent of that resulting from plate tectonics) must also have occurred; this would have tended to enrich the upper mantle and protocrust in incompatible elements. Thus the upper mantle undergoing partial melting above upwelling convection zones is at least likely to have been particularly "fertile" in terms of its basalt-producing capacity, and early "oceanic" crusts are likely to have been thicker than their modern analogue. A thicker crust relative to the overall lithosphere would compound the buoyancy problem as far as viscous drag subduction is concerned.

Secondly, even if this thick "basaltic" crust was subducted, partial melting of it would enhance the overall differentiation of sialic material; second-cycle melting products are inherently buoyant regardless of phase changes at high pressure (density of eclogite facies equivalent of diorite and granodiorite are 3.13 and 3.00 g/cm^3, respectively, Ringwood, 1975, p. 40).

Even thin crust of this composition would inhibit viscous drag subduction unless the lithosphere below was very thick (see Fig. 2-8).

The question becomes: how fast was sialic crustal material differentiated during the first 1 Ga of earth history and, given the high heat flow, in what thickness could it accumulate without remelting at the base?

Estimates of the rate of production of sialic (= continental) crust at island arcs today (e.g. Anderson, 1974), if extrapolated linearly for 4.5 Ga, are at least sufficient to produce the present continental volume. If the rate of production is correlated with heat flow this volume would at least double. Thus, without even considering prior differentiation associated with accretion, the volume of sial likely to have been differentiated by plate-tectonic processes is apparently so great as to "require" that it has undergone recycling, as called for in the geochemical models described earlier.

The view favoured here is that the segregation of sialic crust may have been so rapid, and its thickness so constrained by high thermal gradients, that it spread to cover most of the globe (Hargraves, 1978). Without subduction, however, how was this crust recycled? I picture an earth covered by an almost complete, nonsubductible, scum-crust lithosphere floating above a still vigorously convecting (small-scale plumes?, see Fyfe, 1978) asthenosphere and consider that Archaean granite-greenstone terrains may be a relict of this stage in earth history. Apart from copious outpourings of basalt above upwelling convection currents, the high heat flow must have been reflected in steeper conductive thermal gradients through the crust. Without lateral transport of crust away from the upwelling (convection) zones, the continued outpouring of lava on the surface would eventually thicken the surface crust to the point where partial melting and/or density increasing phase changes occurred at depth. It is suggested (ad hoc) that if to some extent crustal material reaching this boundary became re-incorporated into, or re-equilibrated with, the circulating asthenosphere beneath, then the required recycling might have been achieved.

Onset of negative buoyancy

As average heat flow declined further the oldest lithosphere thickened and its susceptibility to foundering increased. Subduction due to negative buoyancy, when it began, may have taken two forms:

(a) Those parts of the crust which were thinner, and relatively denser, and coupled to thick mantle lithosphere founder completely, initiating buoyancy-powered plate tectonics such as dominates today. This starts the development of the present continent—ocean dichotomy.

(b) Differential stresses between thicker less dense, positively buoyant crust and negatively buoyant mantle lithosphere below would favour their decoupling (delamination), initiating intracontinental orogeny (see also Molnar and Gray, 1979).

Thickening of lithosphere occurs as a result of cooling and accretion of cold mantle material beneath positively buoyant crust. There are thus three parameters in the buoyancy balance: thickness and density of (1) the crustal component of the lithosphere (t_{lc}, ρ_{lc}), (2) the mantle component (t_{lm}, ρ_{lm}), and (3) the density of the asthenosphere (ρ_a). Neglecting viscous drag forces, the minimum requirement for subduction of lithosphere is that its average density $\bar{\rho}_l = [(t_{lc}\rho_{lc} + t_{lm}\rho_{lm})/(t_{lc} + t_{lm})] > \rho_a$. While the lithosphere can then sink, the crustal part of it resists this tendency.

Consider the model illustrated in Fig. 2-10a. The shear stress σ_{m0} along the boundary at the base of the crust (the Moho) is the component of the positive buoyancy force parallel to that boundary:

$$\sigma_{m0} = \Delta\rho \cdot g \cdot t_{lc} \cdot \sin\theta$$

where $\Delta\rho$ is the difference in density between average crust and asthenosphere, t_{lc} is thickness of the crust and θ is the inclination of the boundary. Using the values given earlier ($g = 1000\,\text{cm/sec}^2$, $\rho_{lc} = 2.85$, $\rho_a = 3.22$, $\Delta\rho = 0.37\,\text{g/cm}^3$) and if $10^7\,\text{N/m}^2$ ($100\,\text{bar/cm}^2$) is taken as the critical shear stress for failure:

$$t_{lc} = \sigma_{m0}/\Delta\rho \cdot g \cdot \sin\theta$$

when $\theta = 30°$, $t_{lc} = 5.4$ km; and $\theta = 6°$, $t_{lc} = 27$ km.

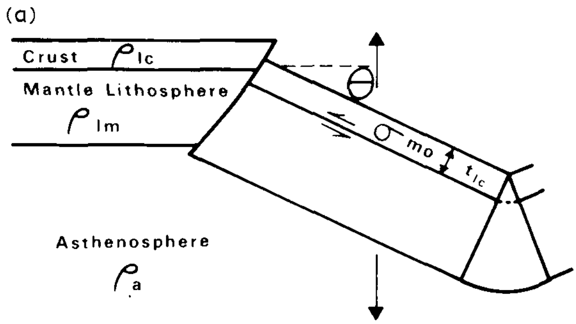
The strength of the boundary between chemical crust and mantle lithosphere will depend upon the temperature. If, as is argued here, the potential thickness of the chemical crust was in part constrained by the thermal gradient, the boundary would be hot close to the melting point, and hence weak. Decoupling would be easier. Today, however, the Moho is relatively cold and the boundary strength probably greater than $10^7\,\text{N/m}^2$ (up to 10^8). The likelihood of decoupling is therefore much lower.

The effective shear stress would also be a function of the inclination of the boundary; for a 6° slope a depression of 15 km (1/2 amplitude) would require a half wave length of ~ 300 km (Fig. 2-10b).

Fracturing of the lithosphere is necessary to free the subducting segment in the first place and it is assumed that such breaks are available. The main point to emerge from this simplified analysis is that shear stresses at the Moho related to the differential buoyancy of chemical crust and mantle lithosphere may well have been sufficient to cause their decoupling. This is a mechanism to cause intraplate orogeny which is otherwise similar to subduction at plate boundaries (see Molnar and Gray, 1979).

Crustal thickening

Implicit in the preceding discussion is the assumption that high heat



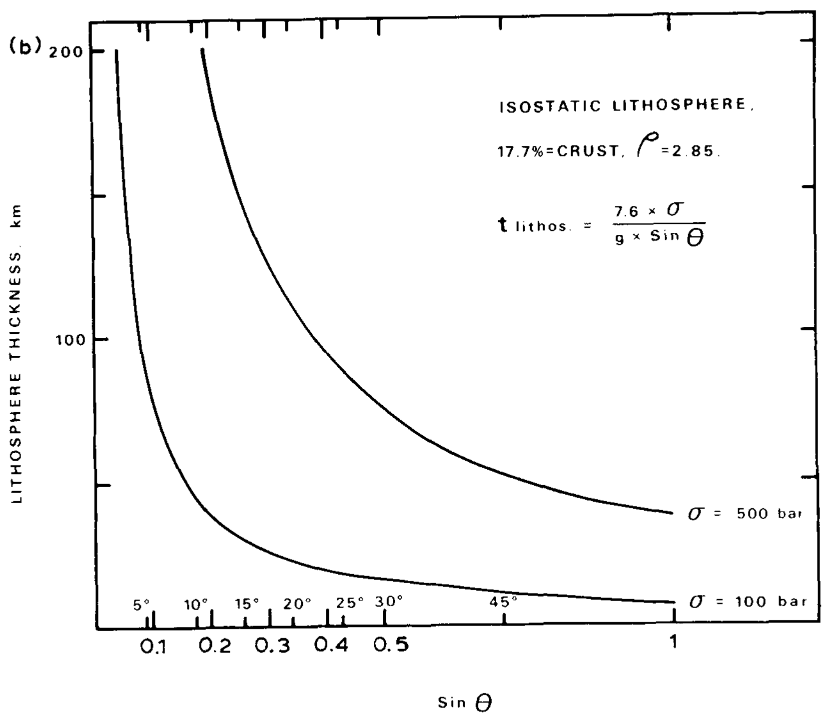


Fig. 2-10. a. Model depicting shear stress on an inclined (angle θ) Mohorovicic discontinuity (σ_{mo}), resulting from the contrasting buoyancies of crust (thickness t_{lc}, density ρ_{lc}) and mantle lithosphere, when the average overall lithosphere is isostatic with asthenosphere (ρ_a). $\sigma_{mo} = (\rho_a - \rho_{lc}) g t_{lc} \sin\theta$.

b. Differential shear stresses on Moho as a function of (isostatic) lithosphere thickness (t_l) and Moho inclination (θ); curves for shear stresses of 500 bar and 100 bar (5 and 1×10^7 N/m^2) are drawn.

flow means both thin lithosphere and thin crust. Extrapolation backwards 3.0 Ga of even the average 1.0 HFU recorded in Archaean cratonic areas today implies steady-state gradients soon after their formation and isostatic adjustment of at least 50°/km in crust with conductivity of 2.5 W/m°C (Pollack and Chapman, 1977a; cf. Bickle, 1978). Therefore, it is considered most unlikely that early Archaean cratons (i.e. after orogeny, erosion and isostatic adjustment) could possibly have been more than 20 km thick and were possibly much less (Hargraves, 1976). Yet as all Archaean crust today appears to average about 35 km in thickness this means that by some means old cratons have been thickened from below without any obvious surface manifestations of the process.

Alternatively one may assume that this is impossible and that minimally eroded Archaean greenstone-belt crust, which is ~ 35 km thick today, actually formed at at least that thickness 3.5 Ga ago (Burke and Kidd, 1978). If that assumption is made then deep, cold, non-radioactive mantle-lithosphere keels must be invoked to form simultaneously beneath these old continental crust segments to shield them from the undoubtedly higher Archaean mantle heat flow (Davies, 1979).

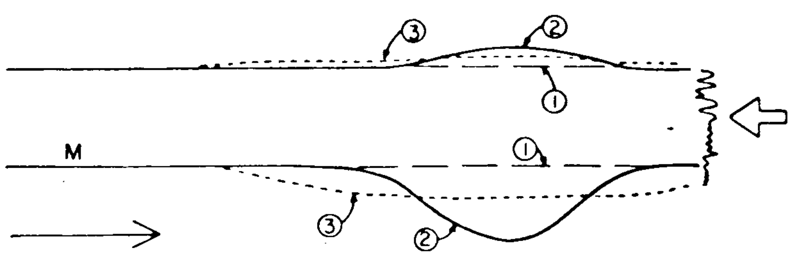


Fig. 2-11. Model depicting basal thickening of crust by lateral spreading of crustal root compensating mountain range at the surface. Figure from Hess (1962). *1* = original crust; *2* = orogenically thickened crust; *3* = crust after erosion above and lateral spreading at the base.

The process of crustal thickening favoured here has been conceived (e.g. Hess, 1962; see Fig. 2-11) or actually invoked by several people (e.g. sialic underplating of Engel, 1970; "slow subcrustal accretion" of Holland and Lambert, 1975) but, to my knowledge, there is no direct evidence to support it. Such evidence could only be obtained from the base of the contemporary Archaean cratons, however, which is not easy. But I consider the general validity or otherwise of this hypothetical process of thickening of continental crust by sialic underplating as one of the most pivotal questions in Precambrian geology.

Summary

Predicated on the declining heat flux, the preferred model for the tectonic evolution of the earth is therefore as follows (Hargraves, 1978; see Fig. 2-12):

(a) Following accretion and some degree of outward differentiation of sialic material the earth was rapidly brought to a state in which protocrust was continuously being produced and consumed above vigorously circulating small-scale convection cells. Although positively buoyant, the crust was pulled down by the viscous drag forces associated with the down currents.

(b) The viscous drag subduction slowly led to the accumulation of a globe-encircling, lithosphere scum-layer of intermediate composition and moderate thickness (10—20 km), completely decoupled from a still vigorously convecting asthenosphere-mantle below. Its thickness constrained by the geothermal gradient, the lithospheric crust is recycled by adding lava at the top and remelting and removing it at the base.

(c) With further decline in heat flow and increase in the mantle fraction of the lithosphere two distinct tectonic styles evolved:

(1) The combination of thin, positively buoyant crust and thick, negatively buoyant mantle lithosphere necessary to give net negative buoyancy and permit subduction eventually arose somewhere on the globe. The buoyancy-powered plate-tectonic regime was initiated.

(2) Elsewhere, foundering of mantle-lithosphere slabs decoupled from thicker crust above also occurred and caused intracontinental orogeny.

(d) Constrained by the ambient thermal gradients, continental crust thickened by orogeny and sialic underplating simultaneous with the enlargement of the oceans.

Such a tectonic evolution model will be appraised on the basis of geologic evidence in the next section.

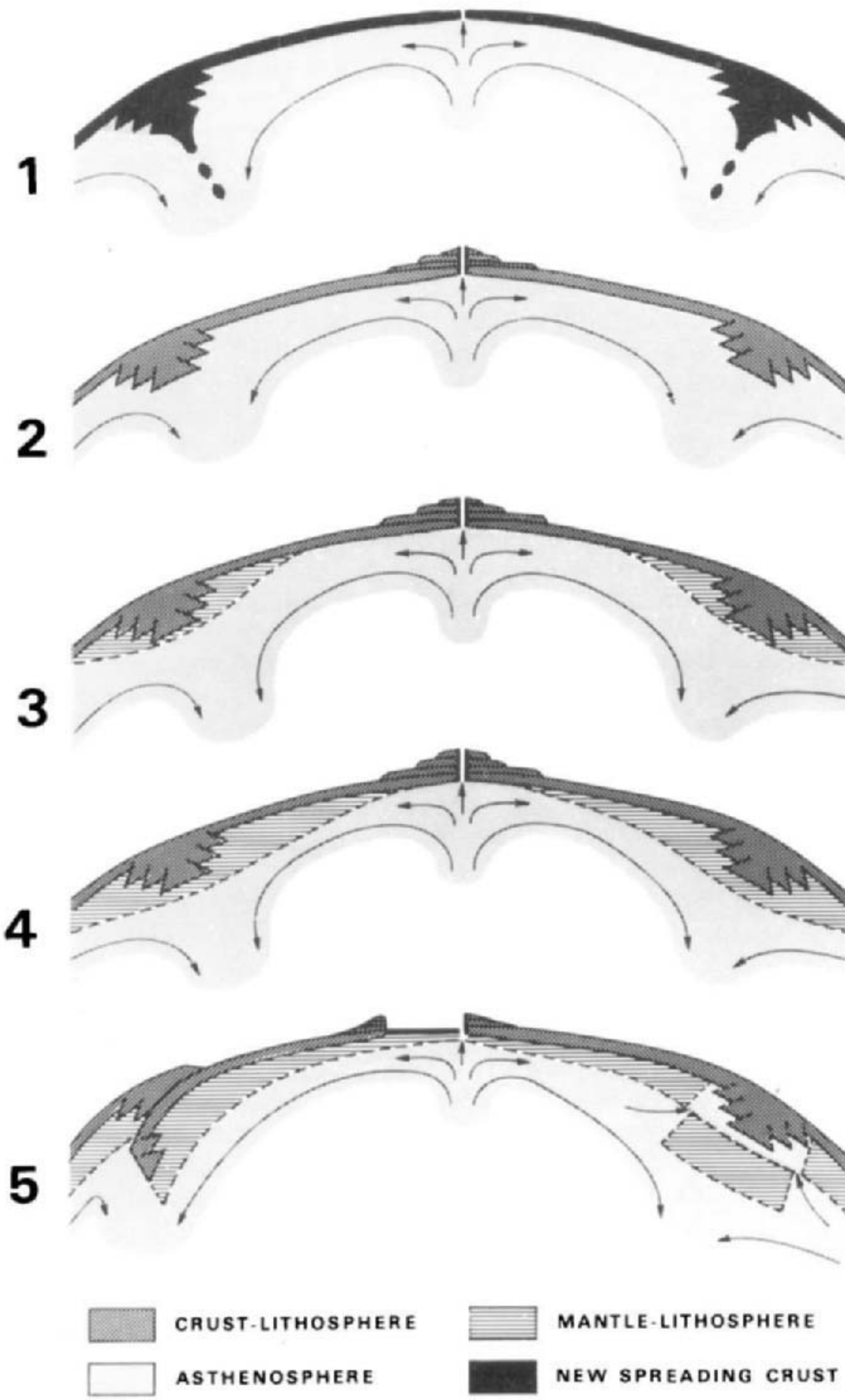
GEOLOGIC EVIDENCE

Pre-Archaean

The absence on earth of rocks older than about 3.8 Ga could have many explanations including intense meteorite bombardment. Such absence is consistent, however, with complete recycling of crust inférred to have been caused by viscous drag forces imposed by vigorous asthenosphere convection.

Archaean

Archaean crust is characterized by predominantly volcanic greenstone belts interspersed with quartzo-feldspathic gneisses, and much controversy


1
2
3
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5
CRUST-LITHOSPHERE
MANTLE-LITHOSPHERE
ASTHENOSPHERE
NEW SPREADING CRUST

has raged over their origin (e.g. Windley, 1973; this volume, Chapter 1). The duality of this assemblage (granites and greenstones) has led to their respective identification with continental and oceanic crust and has fostered conflicting interpretations of the relations between them (see Tarney et al., 1976). One view considers that the greenstones represent primitive oceanic-type crust into which tonalites, granodiorites and granites were intruded; the other view holds that the ancient tonalitic gneisses represent the oldest continental-type basement onto which the greenstone volcanics were erupted. This diversity of opinion testifies to the insidiousness of the uniformitarian presumption that there must be a continent—ocean dichotomy — because we have such today.

The oldest rocks dated on most continents are tonalitic or trondhjemitic gneisses, but as Moorbath eloquently points out, these contain variably digested inclusions of what he considers were once greenstone-type metabasaltic volcanics (the "pre-Amitsoqs", Moorbath, 1977, p. 157). There is thus no "vestige of a beginning" to the processes which produced this kind of bimodal assemblage.

Barker and Arth (1976), however, emphasize that the early Archaean gneiss terrains (which they consider to be older than most greenstone belts), consist of metatonalites and metabasalts from which the equivalents of andesite are conspicuously lacking. To explain this bimodal assemblage, as distinct from one comprising the complete calc-alkaline series such as is found in several younger greenstone belts, they propose that continued eruption of basalt at the surface was accompanied by the intrusion and extrusion of tonalitic liquids derived by anatexis of amphibolites (metabasalts) at the base of the crust. The primitive scum model suggested here, with its concomitant crustal recycling, is consistent with such an evolution: subaqueous extrusion of basalts would promote the formation of partially

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Fig. 2-12. Pictorial cartoon illustrating postulated evolution of tectonic style. *1* = Viscous drag subduction stage: newly formed basaltic crust (= lithosphere) formed above upwelling hot-spot zones is dragged down at convergent zones, but partial remelting distills more sialic material.

2 = Progressive accumulation of sialic differentiates concomitant with decreasing heat flow and viscous drag leads to globe-encircling, positively buoyant, scum-crust (= lithosphere) decoupled from mantle asthenosphere below.

2—3 = Eruption of basalt above upwelling zones (hot-spots?) is no longer accompanied by lateral spreading as buoyant scum-crust is continuous: outpouring of basalt onto scum-crust causes thickening and remelting at base; this stage is characteristic of Archaean crustal evolution.

3—4 = With further slowing of convection lithosphere thickens to include denser mantle above cooler downwelling zones, causing progressive change in buoyancy from positive toward negative.

5 = This thickening culminates in (left) buoyancy-powered subduction, with concomitant growth of new basaltic (= oceanic) crust, and (right) decoupling of mantle lithosphere from overlying crust causes intracontinental type orogeny.

hydrated greenstones. With continued burial by younger flows the greenstones would transform to amphibolite and ultimately melt, at least partially, giving the tonalitic magmas required. Even if initially the crust was entirely basaltic in composition, with continued cycling in this manner it would eventually come to be a bimodal mixture relatively enriched in tonalitic components. This model thus leads in the direction of a globe-encircling "sialic" crust as has been previously invoked (Hargraves, 1976), which is, however, constantly undergoing recycling and regeneration, rather than being permanently differentiated and merely "reworked" (Collerson and Fryer, 1978).

The greenstone belts themselves typically consist of thick synclinal sequences of basic volcanic and sedimentary rocks. Tending to be mafic to ultramafic at the base, the volcanic rocks become more calc-alkaline upwards with an increasing admixture of clastic sedimentary debris (Annhaeusser, 1971). They may unconformably overlie older tonalitic gneisses (Baragar and McGlynn, 1976; *Nisbet et al., this volume, Chapter 7, ed.*) and are in turn intruded by younger granitic plutons. Greenstone-granite assemblages can be repeatedly emplaced in one craton as is most clearly demonstrated by Wilson et al. (1978) in Rhodesia. Here, there are greenstone belts of at least three distinct ages between 3400 and 2700 Ma (*Nisbet et al., this volume, Chapter 7, ed.*).

Greenstone belts could be regarded as relics of the scum-tectonic style whereby thick lava piles accumulated above upwelling zones by eruption through fissures in the globe-encircling crust (see Fyfe, 1978). Undergoing recycling as a consequence of outpouring of basalt above and remelting and recycling at the base, the scum-crust would evolve first to a mixture of tonalite/trondhjemite and basalt (the bimodal assemblage of Barker and Arth, 1976, and Condie, 1976). Continued cycling and eventual remelting of tonalites would generate granite sensu stricto. If the scum-crust only began to form around 3.8 Ga ago, and its residence time is $\sim$ 200 Ma (the time required to extrude 20 km of basalt over the earth), it is no surprise that the first true granites — the product of two stages of anatexis — only formed some 400 Ma later (cf. Burke and Kidd, 1978).

With regard to chemistry Engel et al. (1974) have synthesized major element data on igneous, sedimentary and metamorphic rocks of all ages and expressed the results in terms of the changing K/Na ratios. This ratio is deemed to be a measure of maturity or degree of fractionation of the crust. Their results are shown in Fig. 2-13 (top), and vividly demonstrate the change between Archaean and Proterozoic terrains, the older rocks being less mature or more mafic.

Veizer and Compston (1976) and Veizer (1976) have contributed data on the change with time of $^{87}Sr/^{86}Sr$ of carbonates which are presumed to be related to that of contemporary seawater (Fig. 2-13, bottom). They point out that these changes are consistent with the data of Engel et al.

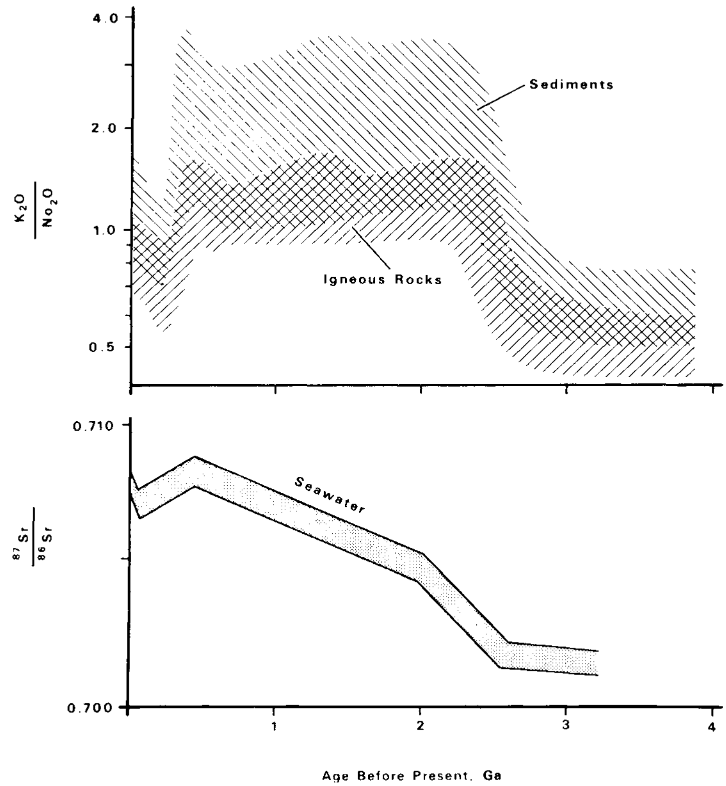

Fig. 2-13. Diagrams illustrating: (top) — secular evolution of K_2O/Na_2O ratios in sediments and igneous rocks, according to Engel et al. (1974); (bottom) — $^{87}Sr/^{86}Sr$ in seawater according to Veizer (1976).

(1974) and favour a more basaltic erosional provenance in the Archaean or at least one which is isotopically indistinguishable from the mantle.

Again these chemical data are consistent with a scum-tectonic regime in which basic volcanic eruptions constitute a major fraction of the crust and the crustal residence time of even felsic components is short..

Although there is considerable temporal overlap the change in geology, geochemistry and tectonic style between Archaean and Proterozoic terrains is striking. There has been considerable speculation in the literature as to the cause of this change, much of it connected in some way with the declining heat flux.

It is suggested here that this change coincides with the initiation of buoyancy-powered lithosphere subduction causing interplate and intraplate orogeny.

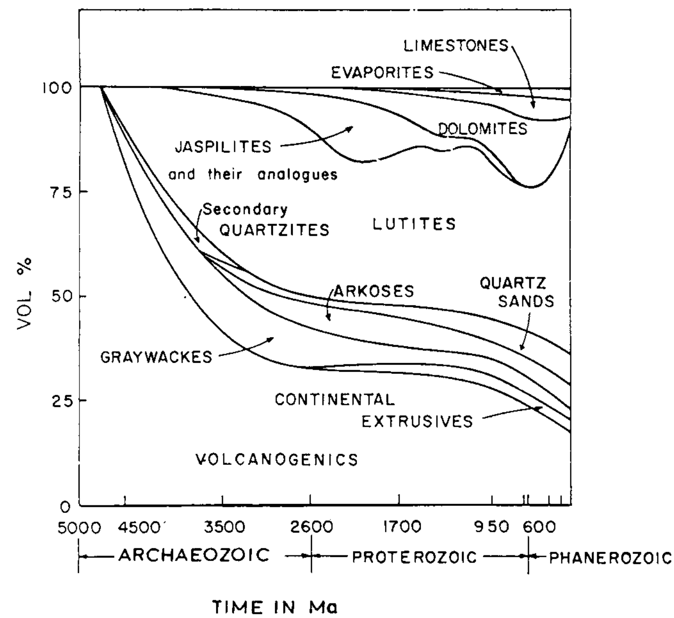

Fig. 2-14. Ronov's (1964) estimate of the relative proportion of different sedimentary rock types as a function of age.

Proterozoic

There is much evidence of long-term change in the geologic record. Ronov (1964) has estimated the relative proportions of different kinds of sedimentary rock as a function of time (Fig. 2-14). These data indicate a preponderance of greywacke and volcanogenic sediment early in earth history, with an increasing proportion of more mature sediment in recent times. Whether this change is real and records a secular evolution in the nature of continental crust and geologic processes or whether it is apparent and an artifact of selective erosion and recycling (Garrels and MacKenzie, 1969, 1971) is much debated. While he does not doubt the importance of selective recycling of sedimentary rock as advocated by Garrels and MacKenzie, Veizer (1973) argues that the changes implied by these kinds of data are real and testify to a change from a more basaltic erosional provenance in Archaean times to more sialic in the Proterozoic.

Compressional thickening of pre-existing positively buoyant crust would tend to occur adjacent to or above the sites of lithosphere subduction: the locus of orogeny. Erosion of these welts would provide the principal source of sedimentary debris because basalt eruptions would be localized at the complementary spreading ridges. With the initiation of the growth by plate tectonics of the oceanic—continental crust dichotomy a substantial change in clastic sedimentary debris could be expected.

The Precambrian palaeomagnetic data, however, are interpreted to indicate that the presently surviving continents were more closely aggregated into possibly only a few major segments throughout much of the Proterozoic (*see also Irving and McGlynn, this volume, Chapter 23, ed.*). This coherence suggests the dominance of some simple, large-scale convection mode (superimposed on the small-scale mode), perhaps analogous to that proposed by Gough (1977), functioning to keep surviving boundary-layer segments confined to, and thickening in, restricted parts of the globe-surface. Alternatively, if the fraction of the earth's surface formed of oceanic crust proper was much less in the Proterozoic, this coherency may reflect the fact that the possible range of intercontinental drift would have been much more restricted (Hargraves, 1976).

Contemporaneously, the foundering of mantle lithosphere decoupled from overlying cratons is manifest in mobile belts, the characteristic Proterozoic orogens. While the wave length of this activity, possibly related to small-scale convection, may have increased with time (Clifford, 1968), the horizontal displacements were relatively small.

COMMENTARY

The arguments presented here emphasize aspects of the early earth which may have caused or allowed the development of a crustal configuration and tectonic style other than strictly uniformitarian. The model may be quite wrong. But it is all too easy to "interpret" geological data in a modern Wilson-cycle framework. Alternative models, such as the one proposed, may serve to encourage a search for, and evaluation of, critical data which might otherwise be overlooked.

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Chapter 3

PRECAMBRIAN PLATE TECTONICS

A. KRÖNER

ABSTRACT

Opinion is still divided whether Precambrian crustal evolution followed uniformitarian principles and is thus compatible with contemporary plate tectonics, or whether global tectonic mechanisms have changed progressively since the Archaean. Major uncertainties are the earth's thermal history, growth of continental crust and the generation of igneous rocks of calc-alkaline affinity.

If Archaean heat flow was 2.5—3 times its present value *all* lithosphere was buoyant and there could have been no gravitational instability and subduction. This implies the early development of a globe-encircling scum crust through recycling and vertical accretion, an evolution which is in conflict with isotope data. An alternative appears to be a hot-spot model with the formation of closely spaced (100—500 km) thermal plume systems and vigorous mantle convection where viscous drag pulls juvenile lithosphere partly back into the mantle. Partial melting and differentiation yields tonalite, the oldest surviving continental crustal rock yet known. The Archaean oceanic crust consisted of piles of komatiitic to tholeiitic lavas and ultramafic intrusives produced over the plumes and was not layered in the same manner as today.

Aggregation of sialic proto-continental material (tonalite with remnants of mafic rocks) yielded buoyant semi-rigid and migrating miniplates which grew through underplating and subsequent attenuation.

The following granite/greenstone phase was characterized by still vigorous sublithospheric convection which induced varying degrees of crustal stretching, rifting and limited ocean opening. Under favourable conditions sag-subduction, partial melting and crust—mantle mixing processes added new crust to the early nuclei and culminated in a series of "continental accretion—differentiation events" (Moorbath, 1977) which established continents or even supercontinents by the end of the Archaean some 2500 Ma ago, resembling their modern counterparts in size and elevation.

With further cooling and thickening of lithosphere in the Proterozoic the average density increased and limited buoyancy-powered subduction began. Relative motion between rigid oceanic plates caused the intervening weaker continental blocks to take up most of the necessary deformation through internal distortion and generated fundamental fractures and shearzones. Differential stresses between light continental crust and underlying negatively buoyant mantle lithosphere favoured their decoupling and may have led to intracontinental orogeny involving underthrusting or subduction of substantial segments of sialic crust (see Molnar and Gray, 1979).

Towards the end of the Precambrian the continuous decline of heat flow resulted in present-day lithospheric thicknesses and the establishment of plate-generation mechanisms which produced layered oceanic crust. With progressive thickening of continental lithosphere through cooling and basalt depletion (Jordan, 1979) the likelihood of crust—mantle decoupling became much lower and, consequently, ensialic orogeny and intracontinental subduction were less frequent.

Finally, further cooling enhanced the negative buoyancy of oceanic lithosphere and established the Phanerozoic Wilson-cycle regime. The worldwide late Precambrian to early Palaeozoic events reflect a transition from predominantly intracontinental to predominantly modern-type plate-margin orogeny.

It is suggested that plate tectonics is non-uniformitarian if the entire history of the earth is considered and began in the Archaean with the generation of small semi-rigid plates. The style and mechanism of plate tectonics changed progressively with the gradual decline in heat flow and increased lithospheric rigidity, and it is therefore not necessary to invoke alternative models in order to explain Precambrian crustal evolution.

INTRODUCTION

Plate tectonics has taken less than 10 years to conquer the world and there are few who do not accept the geophysical and geochemical evidence in favour of presently moving rigid plates and ocean opening and closing. It is surprising, therefore, that attempts to fit Precambrian crustal evolution into the New Global Tectonic scheme have met with variable success. Two extreme views have been expressed over the last few years.

One accepts uniformitarianism for the tectonic evolution of the earth, at least since the formation of some form of rigid crust in the Archaean, and explains crustal development in terms of presently operating (i.e. post-Mesozoic) global mechanisms (e.g. Windley, 1977, and this volume, Chapter 1). The other sees major differences in the geology of Precambrian terrains as compared with Phanerozoic settings and suggests a variety of models for crustal tectonics, based either on dominantly vertical movements (e.g. Ramberg, 1967; Stephanson and Johnson, 1976; van Bemmelen, 1977; Schwerdtner et al., 1979) or on horizontal displacements (Watson, 1976; Sutton, 1977; Dimroth, this volume, Chapter 13) within large continental segments.

A third view has now emerged recently and tries to reconcile the different interpretations on the assumption that global tectonics did not strictly follow uniformitarian principles, mainly because the physical conditions in the lithosphere and underlying mantle have changed through geologic time. This view is supported by the majority of contributions to this volume but accepts the basic concept that global tectonics and crustal evolution result from the interaction of lithospheric plates. However, the present scenario of plate interaction is not accepted as a model for the early Precambrian, and a variety of "primitive" plate-tectonic processes have been envisaged, many of which consider only limited ocean opening and closure and question the operation of the Wilson cycle because of thermal constraints and the rock record (e.g. Lambert, 1976 and this volume, Chapter 18; Baer, 1977 and this volume, Chapter 14; Fyfe, 1978 and this volume, Chapter 22; Young, 1978; Kröner, 1979b).

Many authors who follow this non-uniformitarian approach also question the validity of using geochemical parameters as established for modern tectonic settings in explaining Precambrian crustal development (e.g. Taylor and McLennan, this volume, Chapter 21) since it is becoming increasingly apparent that certain rock types such as bimodal volcanic suites and andesites as well as their intrusive equivalents may form in more than one environment by a variety of processes (e.g. Hallberg et al., 1976; Eichelberger, 1978; Gastil, 1979; Chauvel et al., 1979; Betton and Cox, 1979). Lastly, a great step ahead in the understanding of Precambrian intraplate tectonics is the recent recognition that subduction of continental crust (A-subduction) may play an important role in orogenic processes (e.g. Hatcher, 1978; Hsü, 1979; Kröner, 1979b; Molnar and Gray, 1979; Bally, 1980). This mechanism provides a plausible explanation for ensialic orogeny and basement reactivation which has been postulated for several Precambrian mobile belts (e.g. Baer, 1977; Kröner, 1977a, b, 1979b; Plumb, 1979) and for which the Labrador belt of Canada may serve as one of the best documented examples (Dimroth, 1972 and this volume, Chapter 13).

It would appear, therefore, that virtually all variations in Precambrian tectonic style and the great diversity of ancient rock associations can be explained by the interaction of horizontally moving plates but it seems evident that each era of crustal evolution was characterized by specific tectonic regimes that reflect the response of the crust to the gradual change in the thermal history of our planet. The following account, as well as this entire volume, should demonstrate that no alternative global tectonic theory is required to explain Precambrian geodynamics, but that plate tectonics has changed through time and is therefore non-uniformitarian.

THE ARCHAEAN (> 2.5 Ga)

The formation of the earliest solid crust remains a matter of considerable speculation (e.g. Windley, 1976; Grieve, 1980) and will not be discussed here. Whether this crust was a globe-encircling scum of anorthosite (Shaw, 1976) or granitoid (Fyfe, 1978) and was unsubductable due to a high positive buoyancy or was pulled down back into the mantle by viscous drag forces associated with vigorously circulating convection cells (Hargraves, this volume, Chapter 2) is equally uncertain. All that can be said is that by about 3 Ga ago recognizable protocontinents of sizes up to several thousand km^2 were in existence (e.g. Rhodesia–Kaapvaal, Kröner 1977a; East Africa–Northern Zaire, Cahen et al., 1976; Western Australia, Gorman et al., 1978; Gee, 1979; Southern India, Ramakrishnan et al., 1976; Siberia, Moralev, this volume, Chapter 10; the Superior and Churchill Provinces of the Canadian Shield, Baragar and McGlynn, 1976; Young, 1978; Henderson, this volume, Chapter 9) and these continental segments consisted of granite-greenstone associations and gneissic terrains. Almost 1000 Ma of this early history is now documented, though rather fragmentary and from widely scattered localities in several shields, and geological relationships as well as geochemical data and geophysical considerations make it possible to propose a variety of crustal evolution models for the Archaean some of which are presented in this volume (e.g. Anhaeusser, Barton and Key, Hargraves, Lambert, Moralev, Nisbet et al.).

The following constraints and assumptions must be borne in mind on the basis of presently available data:

(1) Most of the oldest rock complexes yet known consist of tonalitic orthogneisses with very low initial Sr-isotopic ratios and enclose a variety of mafic to felsic metavolcanic remnants of equally primitive geochemistry, which range in size from small xenoliths to individual greenstone belts (e.g. Isua, West Greenland; Barberton, South Africa). Isotopic constraints preclude that the granitoid rocks are derived from much older pre-existing crust (Moorbath and Taylor, this volume, Chapter 20) and at least the mafic to ultramafic complexes must be regarded as mantle differentiates (e.g. Green, 1975, and this volume, Chapter 19). Field relationships and geochemistry tend to favour a virtually concomitant generation of these rock types in settings which some authors have compared with modern destructive plate margins (e.g. Glikson, 1976; Anhaeusser, this volume, Chapter 6; Windley, this volume, Chapter 1).

However, there is at least one documented case from the Limpopo belt of South Africa where 3.8 Ga old siliceous paragneisses (Sand River gneiss) do not contain remnants of older mafic rocks and geochemistry and isotopic data suggest their derivation from still older differentiated crustal material (Barton et al., 1978; Barton and Key, this volume, Chapter 9).

(2) Many of the pre-3 Ga granulite-gneiss terrains consist of a heterogeneous assemblage of quartzo-feldspathic gneisses, partly converted to charnockite-enderbite complexes, as well as thin (several tens to several hundreds of metres) sequences of shallow-water metasediments and associated metavolcanics (e.g. Limpopo belt, Barton and Key, this volume, Chapter 6; Aldan Shield, Moralev, this volume, Chapter 10; the Androyan and Graphite systems of Malagasy, Besairie, 1967; Labrador, Collerson et al., 1976; see also Windley, 1977). Although unconformities at the base of these supracrustal sequences are not preserved, structural relations suggest that they are younger than at least some of the gneisses and were deposited in small basins on a fairly stable sialic floor (e.g. Messina Formation in the Limpopo belt, Barton and Key, this volume, Chapter 8).

(3) All investigators agree that the earth had a higher rate of heat generation in the Archaean than today and that the average terrestrial heat flow has decreased from about 125—180 mW/m^2 at about 3.8 Ga ago to about 60—80 mW/m^2 at present, or by a factor of 2—3 (Hargraves, this volume, Chapter 2; Lambert, this volume, Chapter 18; West and Mareschal, 1979). This implies a vigorously convecting mantle during the early Precambrian with generation of small-scale convective rolls in the asthenosphere below a relatively thin lithosphere (40—60 km) with low torsional rigidity.

In spite of this higher heat flow and higher mantle temperatures (McKenzie and Weiss, 1975) Archaean continental geotherms appear to have been much the same as those of today (Bickle, 1978; England, 1979), and it has been suggested that most of the additional heat was lost either through faster plate production or along longer or more numerous ridge or plume systems (Bickle, 1978; Windley, this volume, Chapter 1). However, Davies (1979) has proposed that the mantle lithosphere under Archaean continents acted as a thermal buffer between the ancient crust and the hot convective asthenosphere since basalt extraction would leave a residuum less dense and cooler than the starting material (Clark and Ringwood, 1964; Jordan, 1978). This sub-continental "tectosphere" (Jordan, 1978) could have been more than 100 km thick as suggested by the presence of Archaean kimberlite-derived diamonds in the Witwatersrand conglomerates of South Africa (Hallbauer et al., 1980).

A considerable proportion of the heat may therefore have been lost through conduction across a thin and positively buoyant oceanic lithosphere. It must also be assumed that separation of the early crust into continental and oceanic types began at least 4 Ga ago.

(4) If all early Archaean crust was buoyant there could have been no gravitational instability and modern-type subduction. Instead, Hargraves (1978 and this volume) has shown that viscous drag forces at the base of the lithosphere must have been much stronger then and could have dragged down at least portions of the early crust into the mantle. Depending on the density difference between this crust and the asthenosphere even thin layers of early sialic differentiate could have been swallowed and thus became recycled into the mantle reservoir. Once continental crust was too thick, positive buoyancy would inhibit subduction but the underlying older and denser mantle lithosphere could still decouple (Hargraves, this volume, Chapter 2) or delaminate (Bird, 1978, 1979; Molnar and Gray, 1979) and sink in response to convective drag. This process may have been important in the generation of greenstone belts and will be discussed further below.

Lambert (this volume, Chapter 18) also considers modern-type subduction unlikely in the early Archaean and favours crustal growth through vertical accretion governed by hot-spot activity.

(5) In the modern plate-tectonic regime generation of new continental lithosphere occurs predominantly along destructive margins and applications of uniformitarian principles would imply successive growth of sialic crust from small Archaean nuclei to present-day continents (Hurley and Rand, 1969; Clifford, 1970). This conclusion is at variance with the observation that large continental segments were already established in the Archaean (see p. 59; also Rutland, 1976; Kröner, 1979a) and that as much as 50—60% of the total mass of existing continental crust was already generated in discrete "accretion—differentiation events" prior to about 2.5 Ga ago (Moorbath, 1977; Moorbath and Taylor, this volume, Chapter 20). This leaves 40—50% for the post-Archaean, and since Jacobsen and Wasserburg (1979) showed

that the growth of continents for the last 500 Ma was *much less* than the average growth rate over the entire history of the earth, only 10% or less of new crust may have been added to the continents in Phanerozoic times through contemporary subduction-related processes.

If crustal growth has decelerated as a result of decreasing plate generation and destruction as some have argued (e.g. Moorbath and Taylor, this volume, Chapter 20), plate interaction or orogenesis must have been considerably faster in the Precambrian than today, yet available data suggest much longer periods of crust-forming events than in the Phanerozoic. For example, the Limpopo belt had a 1000 Ma long orogenic history (Barton and Key, this volume, Chapter 8) and the Barberton greenstone belt and environs may have evolved during more than 500 Ma (Anhaeusser, this volume, Chapter 6). Therefore, modern subduction processes may not be the most efficient means of generating continental crust and alternative mechanism have probably operated through most of the earth's history.

(6) The operation of contemporary plate tectonics requires rigid plates and gravitational body forces on the descending lithosphere to drive the Wilson cycle of plate generation and destruction (Turcotte, 1979). There is little doubt that many segments of Archaean upper crust become stabilized at an early stage as shown by the presence of early Precambrian shallow basins and mafic dyke swarms, but it is not certain whether the Archaean lower crust had the same physical properties as that of today.

Lambert (this volume, Chapter 18) and Hargraves (this volume, Chapter 2) suggest vertical accretion during which the crust is basically thickened from above while the depressed lower part melts to produce tonalite plus granodiorite. West and Mareschal (1979) showed that if Archaean lower crust is covered by a "blanket" of upper crustal volcanic layers, its temperatures would rise considerably to reduce the viscosity and allow for gravity-driven diapric tectonics to take place.

Given these conditions it is doubtful whether plate boundaries of present dimensions and geometries could develop (Glikson, 1976) and it may be significant that no typical "geosynclinal" Archaean successions are known which resemble Phanerozoic stable continental margin deposits.

Similar arguments may apply to Archaean oceanic crust. A higher geothermal gradient and hotter mantle mean a higher proportion of partial melting and magma at shallow depth. Since solidification processes in the magma chamber below ridges determine the structure of the oceanic crust (Kidd, 1977; Dewey and Kidd, 1977) it is unlikely that Archaean thermal conditions and plate driving forces, which must have differed substantially from those of today, would have generated the same layered oceanic crust as now observed. It is not surprising, therefore, that ophiolites in the present meaning of the term (Penrose Conference definition, 1972) have not been identified anywhere in Archaean mafic to ultramafic sequences, and I consider it significant that the most essential evidence of Wilson-cycle tectonics, the presence of sheeted dyke complexes to document laterally spreading oceanic crust, has not yet been recognized.

These points and the lack of a sufficiently rigid oceanic lithosphere may support the contention that large ocean ridge systems were not present in the Archaean and that growth of oceanic crust was caused less by lateral accretion then by vertical accumulation of lava piles over mantle plumes (Lambert, this volume, Chapter 18) whose internal structure was significantly different from that in Phanerozoic ophiolites and in present-day ocean crust. A further implication of Archaean thermal conditions is that 15 km thick layers of dense basaltic to komatiitic lava as described from the Barberton greenstone belt could not have formed above a ridge system of modern type on relatively thin and probably ductile 3.5 Ga old oceanic lithosphere.

(7) A point of considerable importance is whether Archaean rocks of oceanic affinity or calc-alkaline character were necessarily generated in the same environments as those of today. Indeed, much "evidence" has been cited for the operation of Wilson-cycle tectonics in the early Precambrian by demonstration of apparent geochemical similarities between primitive greenstone volcanics and modern ocean-floor basalts, bimodal greenstone volcanism and recent island arcs and tonalitic to trondhjemitic Archaean plutonic rocks and modern Andean-type granitoid batholiths (e.g. Condie and Harrison, 1976; Glikson, 1972; Tarney and Windley, 1977; Windley, this volume, Chapter 1).

Doubt is cast on the practice of using geochemical parameters of contemporary rock associations to infer ancient tectonic settings since it is becoming increasingly apparent that many so-called "distinctive rock types" may occur in more than one environment. For example, many Archaean komatiites and tholeiitic basalts have been interpreted as primitive oceanic crust (e.g. Windley, 1976, 1977), yet such rocks also occur in sequences which demonstrably rest on a granitoid basement (e.g. Henderson, this volume, Chapter 9; Nisbet et al., this volume, Chapter 7; Kröner et al., in press) and komatiites have even been described from the intracratonic early Proterozoic Ventersdorp basin of the Kaapvaal Craton (McIver, 1975). All that can be said is that such rocks are of primitive origin, represent partial melts produced from the mantle and can occur *in any tectonic setting that allows these melts to rise to the surface.* A modern analogue to this situation is given by the basalts of Baffin Island in northeastern Canada which, although clearly resting on Precambrian sialic basement, have all characteristics of mid-ocean ridge basalts (O'Nions and Clarke, 1972).

Similar arguments may be advanced for the ancient bimodal suites and andesites, according to some obvious remnants of Archaean subduction-related island arcs. However, these rocks also occur in rift environments (Hallberg et al., 1976), sometimes even *underlying* komatiitic flows (Schulz, 1978; Hickman, 1980), and also form part of areally extensive intracratonic basin sequences such as the Pilbara Supergroup of Western Australia (Hickman, 1980), the Pongola, Dominion Reef, Witwatersrand and Ventersdorp successions of the Kaapvaal Craton (Haughton, 1969) and some of the Dharwar "geosynclinal piles" of the Indian Shield (Naqvi, 1976) which are clearly not related to subduction.

There is a deplorable lack of detailed stratigraphic and palaeogeographic analyses on the basis of physical volcanology and sedimentology in most Archaean greenstone belts (E. Dimroth, pers. commun., 1980) from which depositional environments could be deduced. Instead geochemical data are frequently employed as a substitute for laborious fieldwork. For example, Archaean subduction zones and even polarities have sometimes been postulated solely on the basis of a few analyses apparently displaying "trends" in certain elements such as K, Rb, Sr, Ti. Yet it is well known that chemical compositions of modern and ancient submarine volcanic rocks are often, if not always, affected by extensive low-temperature alteration, even within individual flows or pillows (e.g. Dimroth and Lichtblau, 1979), with variations greatly exceeding the apparent trends established.

Eichelberger (1978), Fyfe (1980a) and others (for summaries see EOS, 61: 67–68, 1980) have shown that andesitic to rhyolitic rock types may be produced by varying degrees of mixing of lower crust and mantle material, and Betton and Cox (1979) demonstrated that basalts and rhyolites with low initial $^{87}Sr/^{86}Sr$ ratios were probably derived from previously solidified picritic to gabbroic rocks which were emplaced at the base of the crust shortly before eruption of the above-mentioned differentiates during processes of crustal attenuation.

As noted by Fyfe (this volume, Chapter 22) Archaean thermal conditions demand a high degree of melting in the upper mantle and large volumes of dense ($> 3.0\,g/cm^3$) melt products must have accumulated beneath and in the ancient lighter crust. In some areas of extreme tension, fractions of primitive magma may have reached the surface as komatiite but in most cases remained deep in the crust and cooled, allowing the melt to evolve to a basaltic composition and thus promote the necessary decrease in density to permit the remaining magma to rise to the surface as tholeiitic basalt (Betton and Cox, 1979). The newly generated and underplated mafic lower crust would also solidify and become the source of andesitic to rhyolitic volcanism and of granitoid batholiths (Gastil, 1979; Hanson, in press).

Chauvel et al. (1979) concluded on the basis of REE data as well as Sr and Nd isotopic growth values that vast areas of c. 2.7 Ga old granitic gneisses of a typical Archaean granite-greenstone terrain in east-central Finland have been derived by partial melting of short-lived basaltic materials from the lower crust which were separated from their mantle sources no more than 150 Ma prior to granite emplacement. Igneous activity as a result of magma underplating may therefore have been one of the most efficient crust-forming processes in the Archaean and could also have produced the lower crustal heterogeneity as observed by Oliver (1978) and Smithson and Brown (1977).

(8) Lastly, the relationship between Archaean granite-greenstone complexes and high-grade gneissic terrains or belts is still a matter of considerable debate. In most Archaean cratons the granulitic assemblages appear to form a substratum at least to the c. 2.6—2.7 Ga old late Archaean greenstone generations (e.g. Zimbabwe, Nisbet et al., this volume, Chapter 7; Western Australia, Gee, 1979; northern Finland, Kröner et al., in press; Canadian Shield, Baragar and McGlynn, 1976; Young, 1978; Krogh and Gibbins, 1979; Henderson, this volume, Chapter 9; Aldan Shield of Siberia, Moralev, this volume, Chapter 10). This is particularly well demonstrated by the Kaapvaal Craton/Limpopo belt boundary, a transition zone from upper to lower crust (Kröner, 1980a; Barton and Key, this volume, Chapter 8) and the same relationship is evident between the Inari granulite belt and neighbouring greenstones in northeastern Finland (Kröner, 1980a) and the Dharwar-type granite-greenstone terrain and the granulite-charnockite complex of southern India (Ramiengar et al., 1978).

The rocks of the high-grade terrains consist of a variety of orthogneisses as well as remnants of predominantly shallow-water metasedimentary sequences and anorthosites (see Windley, 1977 for a detailed description). Modern seismic work shows that much of the present-day lower crust of the continents is composed of similar heterogeneous assemblages (Heier, 1973; Fairhead and Scovell, 1977; Smithson and Brown, 1977; Schilt et al., 1979). The real problem is how these complexes were generated. Obviously many of the metasediments were once deposited on a granitoid crust, then deformed and transferred to great crustal depth while becoming modified, injected by melts and finally dehydrated.

It seems impossible and implausible to me that *all* granulites of the earth's crust were produced by modern-type continental collision orogenies, and other mechanisms must be invoked, the more so since many Archaean and Proterozoic high-grade terrains lack the pronounced linearity of Phanerozoic orogenic belts and reveal a tectono-metamorphic evolution that extended over hundreds of millions of years (Kröner, 1979a).

One possible mechanism for the formation of non-linear early Archaean granulite complexes appears to be crustal "overplating" by large amounts of greenstone volcanics and high-level granitoid batholiths (Fyfe, this volume, Chapter 22; Goodwin, this volume, Chapter 5; Hargraves, this volume, Chapter 2; Lambert, this volume, Chapter 18) through which the older rocks were depressed to lower crustal levels and transformed to granulites. This process may be enhanced through a considerable rise in lower crustal temperatures due to the thermal blanket effect of overlying dense volcanic piles (West and Mareschal, 1979) and cause migmatization, melting and the onset of gravity-induced granite diapirism.

Alternatively, crustal thickening through over- or underthrusting of segments of continental crust may be visualized to generate quasi-linear high-grade belts (Myers, 1976; Kröner, 1980a). Molnar and Gray (1979) and

Hargraves (this volume, Chapter 2) have shown that, given sufficiently high temperatures, upper crustal plates could be detached from their base if the density difference between the two segments exceeds a certain critical value. The subcrustal lithosphere would then be pulled down due to viscous drag forces while the upper crust would delaminate and be thrust over the adjoining lithosphere which remained intact. Such interstacking by over- and/or underthrusting of crustal segments has been reported from the Appalachians (Cook et al., 1979) and the Alpine-Himalaya collision belts (Bird, 1978; Hsü, 1979; Reutter et al., 1980) but would have been more frequent in the Archaean since crustal delamination was easier then due to a reduced boundary strength between crust and hotter mantle in a thinner lithosphere (Hargraves, this volume, Chapter 2).

The difference between Archaean and modern collision-induced crustal thickening is also evident in the fact that none of the ancient granulite terrains has a typical molasse in its foreland. Even in the low-grade greenstone belts molasse-like sediments including clastics were deposited *into* the evolving greenstone basins and there is no evidence for mountain building, quick uplift and erosion with subsequent deposition into foredeeps. This indicates that Archaean uplift was extremely slow, thus perhaps also inhibiting the preservation of high-pressure and low-temperature metamorphic assemblages, and crustal thickening was caused either largely by vertical accretion as postulated above, or the thickening through thrusting was short-lived and caused almost immediate melting at the base of the crust to restore the original crustal thickness. The latter mechanism may explain why the original basement-cover surface is now again exposed in several ancient high-grade terrains such as the Limpopo belt, after it was apparently transferred to crustal depths of more than 20—25 km during Archaean tectonic events.

Archaean plate tectonics — dominance of rifting and sag-subduction

Given the above constraints and assumptions and the geological field relationships of early Precambrian rock assemblages (e.g. Windley, 1976, 1977; Kröner, 1977a; Hunter, 1981; Chapters in this volume) the following model for the evolution of Archaean continental crust is preferred, which draws heavily on the speculations of Goodwin (1977 and this volume, Chapter 5), Hargraves (1978 and this volume, Chapter 2), Kröner (1979a) and Lambert (this volume, Chapter 18).

The first solid layer of crust on earth may have been a cumulate of anorthosite (Shaw, 1976), produced as a buoyant scum from a terrestrial magma ocean (Warner, 1979). On this crust piles of mafic lava accumulated, with internal structures unlike those of modern ocean floors, above numerous thermal plumes controlled by a system of closely spaced convection cells in a vigorously convecting mantle. Growth of the volcanic pile following partial melting of the subcrustal mantle immediately below resulted in basalt depletion and cooling of this region so as to form a mini-tectosphere (Jordan, 1979) which, together with the anorthositic cumulate and the volcanic pile, makes a coherent unit that can be regarded as a buoyant miniplate. As the pile thickens its lower parts and the anorthosite began to melt, with tonalite and granodiorite intruding their own parent, some highly differentiated lavas reaching the surface and a dense residue remaining at the base of this early crust. If thick, old and cool enough viscous drag may succeed in decoupling this residue together with the underlying thin tectosphere to allow recycling into the convecting mantle. Continuous growth of this kind, coupled with further basalt-depletion of the mantle over plumes, will establish coherent plates which "float" on the turbulent asthenosphere. At the same time repeated melting at and near the base of the crust with detachment and removal of dense residue due to viscous drag in the asthenosphere will gradually produce a two-layered structure with the lower part transformed into high-grade assemblages and the upper part consisting of both primitive and differentiated volcanics and granitoid intrusives. Gravitational instability, meteorite bombardment and frequent breakup of these thin miniplates will facilitate the continuous rise of tonalitic and granodioritic diapirs to the surface which are cut by mafic dykes and are again overplated by mafic volcanic rocks.

Eventually a situation may be reached where a mixture of granitoid intrusives and differentiated volcanics constitutes the surface of this now predominantly sialic early upper crust. If it emerges above sealevel, erosion ensues and early sediments with strong calc-alkaline affinities that reflect the composition of their parents are deposited. The c. 3.8 Ga old Sand River tonalites of the Limpopo belt could be examples of this type, and I visualize that the crust had evolved to this stage by about 4 Ga ago. However, no extensive sedimentary basins formed at that time since only small parts of the microcontinents emerged significantly above sea level because vertical continental accretion was compensated by fusion at depth and, perhaps, sideways flow by ductile spreading of the lower crust (cf. Hess, 1962).

All this crust was buoyant, and the miniplates may have moved around in an irregular "jostling" fashion in response to the small-scale convection system. Neighbouring oceanic crust, where insufficient thicknesses of mafic volcanics prohibited differentiation into felsic components to form sialic nuclei, grew older and denser, and jostling plate motion may have facilitated plate breakup at the continent—ocean boundary with subsequent near vertical downward pull of oceanic lithosphere due to viscous asthenospheric drag. Collision of closely spaced and small continental plates followed and the oldest greenstone belt assemblages may now mark the sutures of these processes. Elsewhere new oceanic crust was formed along fractures or oceanic rift zones over plumes to compensate for the subducted older material and may have become the locus for the formation of new sialic crust by differentiation, perhaps analogous to modern Iceland.

The scenario is now set for the evolution of the widespread Archaean granite-greenstone terrains between about 3.8 and 2.5 Ga ago. The old belts of the 3.8 to 3.5 Ga generations such as the Isua of Greenland, the Sebakwian of Zimbabwe, the Barberton of South Africa and the Warrawoona of Australia apparently have no identifiable basement and the predominance of mafic to ultramafic rocks in some of these may suggest that they represent remnants of some form of early oceanic crust. However, the ubiquituous presence of sialic components with polyphase deformation patterns often not seen in the neighbouring greenstones indicates that continental crust existed at least somewhere nearby and that these ocean basins must have been relatively small and/or were probably floored by older sialic material. The small size of individual greenstone belts in comparison with the large crustal segments in which they occur also seems to support this contention.

The younger greenstone belts of the late Archaean are best explained by the plume model of Lambert (this volume, Chapter 18) and the rift-and-sag-model of Goodwin (1977, and this volume, Chapter 5), whereby crustal attenuation and fissuring above hot-spots or "hot lines" in the mantle generates oval or elongate basins which may attain considerable dimensions of several hundred km (e.g. Warrawoona Group, Australia, Hickman, 1980) and which collect shallow-water sediments as well as mafic to ultramafic lavas (komatiites) produced by a high degree of melting in the subcrustal mantle (Fig. 3-1a). Since the first mantle differentiates of komatiitic composition are significantly denser than the overlying crust only a small proportion can rise to the surface and extrude as high-Mg lava while a much larger volume remains in the crust or at its base and may become the source for later, more differentiated volcanics and intrusives. Further rifting causes crustal subsidence in the proto-greenstone basin and subsequent rise of the crust—mantle boundary. This process is enhanced by both the weight of the already extruded dense komatiitic and tholeiitic volcanics and the downward drag by underplated peridotitic magmas (Fyfe, 1978; Fig. 3-1b). Eventually the spreading and sagging process may lead to the decoupling of crustal blocks with further production of great volumes of basaltic rocks. At that stage fusion and crust—mantle mixing become important and lead to the production of andesites and more highly differentiated volcanics higher in the greenstone pile as well as consanguineous calc-alkaline granitoid intrusives (Fig. 3-1c). This mechanism explains why rocks of ocean-floor affinity (komatiitic and tholeiitic basalts) now occur in the same stratigraphic sequence as island-arc-type assemblages (Fyfe, 1978).

If the crust becomes sufficiently thin during stretching and sagging it may eventually break apart with the formation of small ocean basins in which ensimatic greenstone belts can form (Goodwin, 1977); see also Fig. 3-1d. The mechanism of fissuring and rupture proposed here may be analogous to the formation of modern rift systems and small ocean basins of Red Sea type. The ensimatic greenstone belts of this type would consist of oceanic

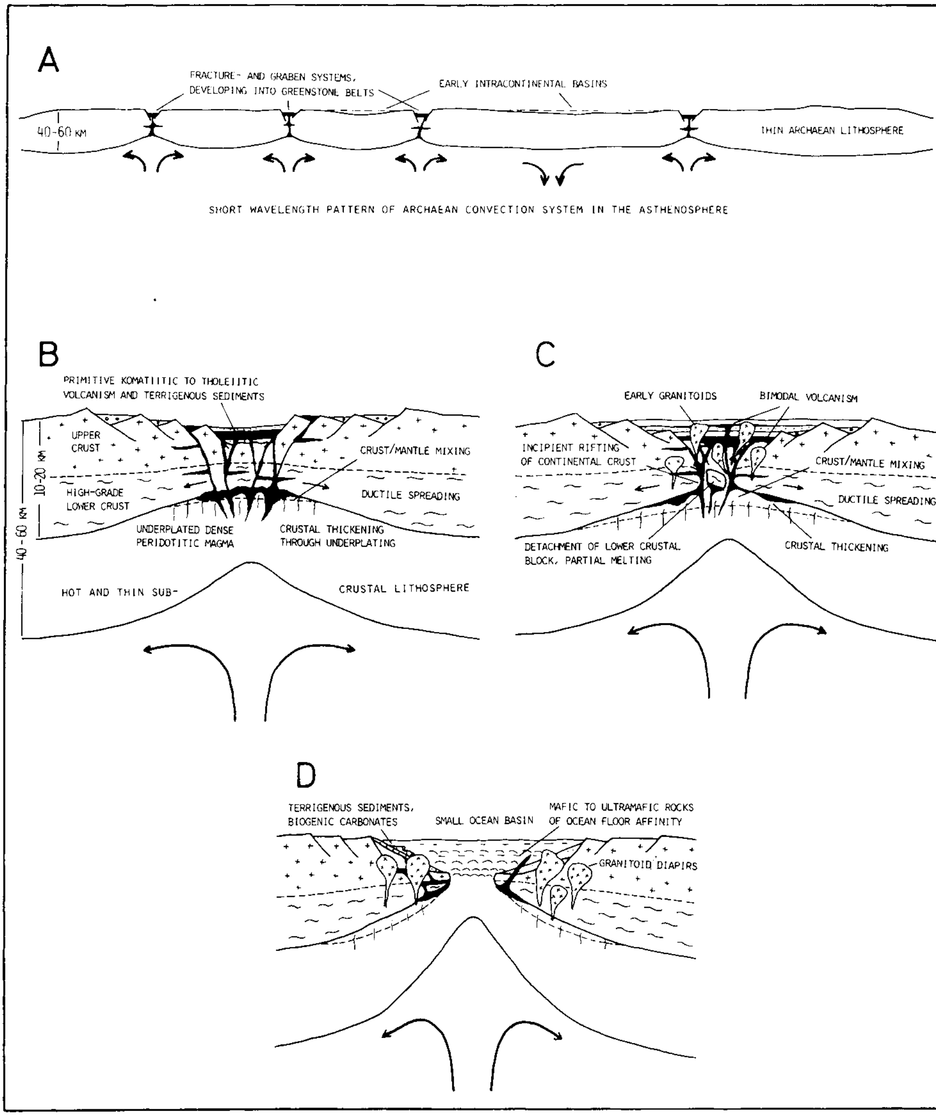

Fig. 3-1. Schematic cross-sections showing suggested evolution of thin Archaean lithosphere in response to small-scale convection pattern in the mantle. For explanation see text. Modified after Kröner (1979a).

crust, structurally unlike that of modern oceans, and would have to be founded on a subcrustal lithospheric residue, presumably a massive peridotite of some sort. Such mantle peridotite has not yet been found at the base of any greenstone belt but there are many belts whose base is not exposed. It is also possible that closing of such oceanic basins would result in detachment of the dense lower parts, if they were present, or anatectic melting would conceal the original relationships as in the Barberton belt of South Africa (Anhaeusser, this volume, Chapter 6).

The model described here is attractive in that it explains the great diversity of granite-greenstone relationships and greenstone lithologies as a consequence of different degrees of subcrustal thermal activity which may lead to different stages of crustal breakup. In cases of incipient rifting such as in the Pongola basin of South Africa (Hegner et al., 1981) early ultramafic melts did not reach the surface and only lighter and more differentiated volcanics were extruded. Elsewhere, however, deep fracturing enabled variable amounts of high-Mg magma to reach the surface where they became interlayered with more evolved volcanic flows. The existence of an older sialic crust is an essential part of the model and is not only supported by numerous gneiss/greenstone unconformities but also by the suggestion that the c. 2.7 Ga old greenstone belts of Zimbabwe and Western Australia are remnants of once continuous stratigraphic sequences with a lateral extent of several hundred km (Wilson et al., 1978; Gee, 1979; Hickmann, 1980).

Even the Barberton belt of South Africa, frequently cited as a prime example of ensimatic Archaean plate interaction (Glikson, 1976; De Wit and Stern, 1980), must have evolved on or at least near an older continental crust as suggested by the apparently pre-Onverwacht ages of some of the granitoid clasts in the upper greenstone Moodies conglomerate (Van Niekerk and Burger, 1978). Also, the available radiometric ages indicate that deposition in the Barberton basin took place during more that 200 Ma, an unlikely long time span if compared to contemporary ensimatic arc evolution and, in my opinion, more in line with development of the belt from a slowly opening rift structure. This view is also supported by the shallow-water nature of deposition during Onverwacht times (Lowe and Knauth, 1977) and by sedimentary structures in the upper strata of this belt (Eriksson, 1979).

Windley (this volume, Chapter 1) considers granite-greenstone terrains as products of extensive rifting in modern-type marginal basins and of calc-alkaline magmatism due to subduction of oceanic crust. Apart from the difficulty in explaining the remarkably continuous volcanic stratigraphy in some regions as mentioned above, this model relies on the operation of present-day convection and subduction processes in the mantle, a prerequisite for the formation of marginal basins (Toksöz and Bird, 1977a; Judy, 1979), but considered unlikely for the Archaean on thermal grounds (Hargraves, this volume, Chapter 2; Lambert, this volume, Chapter 18).

Even in cases of complete plate separation and ensimatic greenstone generation the high oceanic geotherm at that time must have kept the newly formed crust buoyant. Hargraves (this volume, Chapter 2) has shown that subduction of oceanic crust by asthenospheric viscous drag could only occur if such crust was thinner than about 4 km. More likely, the subcrustal oceanic lithosphere would decouple from its top and sink down while the "floating lid" would be piled up and intensely folded between the advancing continental blocks and rising granitoid diapirs (Goodwin, this volume, Chapter 5).

The Archaean high-grade terrains represent the pre-greenstone sialic crust together with predominantly sedimentary (i.e. ensialic) deposits and granitoids, produced during various stages of incomplete pre-greenstone fissuring and subsidence. Greenstone evolution (i.e. overplating) and/or crustal underthrusting and interstacking moves much of this material into lower crustal levels with transformation of granitoids and supracrustal rocks into migmatites, banded gneisses, granulites and grey tonalitic gneisses. It is suggested that this type of lower crust underlies most, if not all, granite-greenstone terrains in the ancient shields and was not produced during distinct "orogenic" periods.

By the end of the Archaean large crustal segments of at least subcontinental proportions were in existence (Kröner, 1979a), and one might speculate that the entire Archaean continental crust was assembled in one large supercontinent (Goodwin, 1974) such as the later Pangaea, thus leaving a globe-encircling primeval ocean in which innumerable volcanic chains formed lines of hot-spots — the forerunners of modern ridge systems — through which the earth outgassed and lost much of its heat.

The worldwide production of granitoid rocks at about 2500—2700 Ma ago may not only have increased the size and thickness of the early crust but must also have led to substantial cooling in the mantle, thereby changing the convective pattern to larger, more regular convection cells (Runcorn, 1965; Fyfe, 1978) and probably also leading to thicker lithospheric segments. Palaeomagnetic data suggest that some form of continental drift must have operated by that time (Irving and McGlynn, this volume, Chapter 23) and this may indicate that the "disorganized" Archaean hot-spot spreading gave way to a more regular pattern of linear volcanic activity along primitive ocean ridge systems, determined by the orientation of the larger convective mantle currents. McCulloch and Wasserburg (1978) speculated from Sm-Nd and Rb-Sr systematics of Precambrian rocks that the dramatic decrease in the volume of continental crust produced since the end of the major crust-forming event at c. 2.5 Ga ago may be ascribed to a global change in crust-forming mechanism rather than only in the rate of crust formation.

There remains the question whether the major crust-forming episode around 2.7—2.5 Ga ago was related to a dramatic climax in the thermal evolution of the early earth (Lambert, in press), or whether we overestimate this time period simply because many Archaean cratons reflecting ages in the above range are not deeply eroded so that we always date the late Archaean upper crustal events. More precise isotopic data on the high-grade gneisses of Archaean lower crust already indicate that equally important crust-forming episodes occurred at earlier times (e.g. Mueller and Wooden, 1979), but available data are not sufficient so far to recognize significant worldwide clustering of ages, although the 3.5 Ga event appears to be a good candidate.

First large intracontinental basins with mixed volcanic-sedimentary infill developed towards the end of the Archaean on greatly stabilized and thickened cratons such as the Kaapvaal. The geochemical similarity between the greenstone volcanics and some of the basin lavas such as the c. 2.9 Ga old Pongola (Hegner et al., 1981) may indicate that the latter are failed greenstone belts where attempts at crustal rupture remained unsuccessful.

THE LOWER PROTEROZOIC (2.5—1.2 Ga)

While the literature abounds with speculations on Archaean tectonics there are much less data on crustal evolution during the Lower Proterozoic between c. 2.5 Ga and c. 1.2 Ga ago. Particularly the time period between 2.5 Ga and 2.2 Ga ago apparently lacks worldwide tectono-thermal activity or orogenesis and appears to be characterized by the formation of large sedimentary basins, some of which evolved on stable continental crust (e.g. Witwatersrand—Ventersdorp basin of the Kaapvaal Craton, Fortescue—Hamersley basin of Western Australia) while others begin to bear resemblance with Phanerozoic plate margins or "geosynclinal" successions (e.g. Birrimian of West Africa, Bessoles, 1977; Huronian of the Canadian Shield; Svecokarelian of the Baltic Shield).

Some of the late Archaean to early Proterozoic mixed volcanic-sedimentary sequences resemble greenstone belts although the proportion of sedimentary rocks is much higher and komatiitic lavas are only rarely found, as exemplified by the Dharwar belts of Peninsular India (Naqvi, 1976). All these sequences rest on older granitoid-gneiss-greenstone complexes and, as postulated for the Pongola basin of South Africa, they may represent abortive attempts of the stabilized continental crust to break apart. The same applies to the Great Dyke of Zimbabwe (Rhodesia), a prime example of incipient rifting of rigid crust 2500 Ma ago (Van Biljon, 1976).

From a crustal evolution point of view, the Archaean—Proterozoic boundary represents a transition (Cloud, 1976), characterized by the diachronous successive stabilization of cratonic blocks during more than 500 Ma from about 3 Ga to about 2.4 Ga ago with a peak at about 2.6—2.7 Ga ago, identified as one of the major crust-forming events in earth history (Moorbath, 1977; McCulloch and Wasserburg, 1978).

If, as speculated above and perhaps supported by the formation of early Proterozoic cratonic basins, the mantle convection system changed to larger, more regular convection cells, large rigid plates drifted independently of each other (Irving and McGlynn, this volume, Chapter 23; McWilliams, this volume, Chapter 26), and the likelihood of foundering of the thickened lithosphere increased due to the decline of the average heat flow (Hargraves, this volume, Chapter 2), did this signify the onset of conventional Wilson-cycle plate tectonics?

Many geologists working in Precambrian terrains would say no, since there is not one single documented case where the important and decisive criteria for seafloor-spreading and B-(Benioff) subduction such as ophiolites with sheeted dykes and blueschist associations have been observed in the Lower Proterozoic rock record. Therefore, "ensialic" orogenesis not involving plate separation has been proposed. For example, Dimroth (1972, and this volume, Chapter 13) has provided convincing evidence from a detailed analysis of the mid-Proterozoic Labrador geosyncline that this basin evolved into a mobile belt without plate separation and collision but by horizontal shortening through basement reactivation and ensialic piling up of crustal wedges and nappes.

Similar explanations have been offered for several Australian (Rutland, 1976) and African (Kröner, 1977b) Proterozoic belts in which volcanic rocks of tholeiitic or calc-alkaline affinity are conspicuously absent and in which the existence of continuous pre-orogenic sialic basement floor can be demonstrated (e.g. Median belt of Western Australia, Horwitz and Smith, 1978; Zambezi and Irumide belts of east-central Africa, Kröner, 1977b).

Palaeomagnetic data seem to support such models or at least rule out extensive relative motion between the crustal blocks bordering such belts (Irving and McGlynn, 1976 and this volume, Chapter 23); McElhinny and McWilliams, 1977; McWilliams, this volume, Chapter 26; Poorter, this volume, Chapter 24), but the early to mid-Proterozoic apparent polar wander paths for many cratonic segments of the present continents are still poorly known and leave room for wide speculation (e.g. Burke et al., 1976). It must also be remembered, as pointed out by Roy and Robertson (1979), that a palaeomagnetic pole obtained from a rock unit is only a rough approximation of the true pole location, and the pole derived from the sites selected does not necessarily represent the location of the entire rock unit pole as accurately as the statistics indicate. McWilliams (this volume, Chapter 26) and Irving and McGlynn (this volume, Chapter 23) discuss further limitations of palaeomagnetic data in determining past-ocean opening and closing events on the basis of presently available information.

In spite of all these serious objections to the application of actualistic principles to explain Proterozoic tectonics there are a number of well-studied terrains, particularly in the age range 1.7—2.1 Ga, which exhibit geological features closely resembling Phanerozoic orogenic belts. Many of these have well-defined miogeosynclinal and eugeosynclinal basins. Their deformation by horizontal compression and the evolution of calc-alkaline assemblages with primitive isotopic systematics (e.g. Reid, 1979) somehow suggest that Lower Proterozoic tectonics must have been controlled by some form of plate tectonics.

Hoffmann (1980) has interpreted the mid-Proterozoic Wopmay orogen of the Canadian Shield in terms of Wilson cycle evolution but admitted that there seems to be no suture and no trace of obducted ocean floor. Likewise, Baragar and Scoates (this volume, Chapter 12) have interpreted the Lake Superior Association along the southern margin of the Archaean Superior Craton as a result of ocean consumption and continental collision, but fail to identify a suture which they suspect to be obliterated by the later Grenville belt. At the same time, however, these authors recognize contemporaneous ensialic assemblages elsewhere in the Circum-Superior belt which they ascribe to incipient continental rifting.

Other interpretations of early to mid-Proterozoic mobile belts in terms of the Wilson cycle are frequently based on apparent systematic changes in certain geochemical patterns rather than on rock relationships (e.g. Hietanen, 1975), and such models must be regarded with caution for reasons discussed before (p. 63). For instance, the supposedly simple relationship between rock composition and depth to subduction zones in modern arc systems has been applied to many Proterozoic settings and led to numerous plate-consumption models. However, even in recent arc systems the variations along arcs are often as great as, if not greater than, those across them (Gass, this volume, Chapter 15). It seems unjustified, therefore, to apply poorly understood geochemical and petrological patterns from modern volcanic belts to ancient settings, the more so since a detailed understanding of the mechanism which produces igneous rocks in contemporary island arcs is still lacking (Turcotte, 1979).

In summary, there are indications for Wilson-cycle evolution in the Lower Proterozoic in a few cases, but the majority of mobile belts in the age range 2.3 Ga to about 1.2 Ga exhibit features which are more compatible with no or only limited ocean opening and virtually no consumption of ocean floor but substantial crustal shortening involving both cover and basement (Kröner, 1977a, b) and/or intraplate horizontal displacements along major shear-zones and shear-belts (Sutton, 1977; Plumb, 1979).

In 1977 I suggested that the apparently ensialic belts were formed by the same subcrustal forces which operate during present-day plate separation, ocean opening and continental collision (Kröner, 1977b). If these forces are not strong enough to cause complete continental rupture, linear zones of weakness and intraplate graben systems, aulacogens or geosynclinal basins develop in the crust. The orogenic cycle is set in motion through compressive stress in the lithosphere, gravitational instability, and increased heat flow through the fractured crust in the zone of weakness.

All transitional stages from non-orogenic intraplate graben systems through ensialic mobile belts to collision belts following plate separation and ocean closure should therefore exist. This model tried to reconcile the conflicting views on Proterozoic tectonics but left many supporters of the plate-tectonic concept unconvinced. "Ensialic orogeny" remained an idiosyncratic, non-uniformitarian mechanism (Windley, this volume, Chapter 1).

Later I suggested that limited intracrustal subduction may have been the main mechanism of Proterozoic belt formation (Kröner, 1979b) since geophysical data from several Precambrian shields supported the presence of major listric fault systems along the margin of mobile zones which reach down far into the lower crust (e.g. Kratz, 1978; Smithson et al., 1978; Kaila et al., 1979).

Behr (1978) and Weber (1978) reached similar conclusions from structural and petrological data and interpreted part of the late Palaeozoic Hercynian belt of central Europe in terms of subfluence processes in the lower crust, caused by viscous drag in the underlying mantle.

Following a Penrose Conference discussion in 1978 where broadly similar mechanisms were proposed for the evolution of the Alpine belt (Hsü, 1979) and the term A-subduction was introduced for the underthrusting of continental lithosphere in honour of O. Ampferer who first suggested this process in 1906, even plate-tectonic devotees began to accept the feasibility of sialic underplating (Bally, 1980). The major objection to this assumption — the non-subductibility of continental lithosphere — had already been removed by the demonstration of intracontinental subduction in the Himalayas by Toksöz and Bird (1977b) and Bird (1978) and by the calculations of Molnar and Gray (1979), which showed that significant fractions of the continental crust up to several hundred km in length may be subducted if they could be delaminated from their upper part. These authors and Hargraves (this volume, Chapter 2) suggest that the gravitational force acting on the descending old and dense oceanic lithosphere may pull parts of continental lithosphere with it until this pull is counterbalanced by the high buoyancy of the subducted continental crust. Hargraves (this volume, Chapter 2) also calculated that shear stresses at the Moho related to the differential buoyancy of sialic crust and the underlying mantle lithosphere may well have been sufficient in the Proterozoic to cause their decoupling. All that is needed are fractures in the lithosphere to free the subducting segment (Bird, 1979).

It is of some comfort to know that A-subduction, which for some time was repulsive to theoretical plate tectonicians, has now been accepted in principle (Bally, 1980), and we may now proceed to analyze Precambrian orogenesis without the need to invoke non-plate-tectonic mechanisms.

Lower Proterozoic plate tectonics — dominance of A-subduction and intraplate shearing

Massive crustal growth, lithospheric thickening, decrease in heat flow and probable change in the mantle-convection pattern at the end of the Archaean all had a profound influence on the behaviour of continental lithosphere in the following time period of the early to middle Proterozoic; it may have consisted of only a few large supercontinents (Piper, 1976; McElhinny and McWilliams, 1977). It is noteworthy that Lower Proterozoic strata contain only few of the primitive volcanic and granitoid rock associations so widespread in the Archaean; mature stable shelf sequences strongly predominate. Furthermore, the initial $^{87}Sr/^{86}Sr$ ratios of granitoid rocks of that time show a marked increase as compared with the low values during the Archaean (e.g. Glikson, 1979, table 2), suggesting pervasive remobilization during intracratonic anatectic processes which "softened" the crust in zones of high strain.

Two decisive factors seem to have controlled Proterozoic tectonics in the supercontinents:

(1) From the end of the Archaean the thickness of continental lithosphere was much greater than those of oceanic lithosphere and values up to 200 km have been suggested by Davies (1979). Most of the earth's heat therefore emerged through the oceanic lithosphere which remained comparatively warm and buoyant.

(2) The resultant lower heat flux from the sub-continental mantle, coupled with a higher production of radiogenic heat from the upper crust as a result of late Archaean differentiation processes, produced a positively buoyant crust underlain by dense, thick, negatively buoyant mantle lithosphere.

It is therefore unlikely that subduction of oceanic crust could be initiated (Green, 1975; Baer, 1977), but this situation need not have prohibited plate tectonics. Since oceanic crust is considered to be more rigid than continental crust (McKenzie, 1969; Jordan, 1979) some relative motion between strong, rigid oceanic plates might have been possible with the intervening weaker continental plates taking up most of the necessary deformation through internal distortion or along large shear-zones. This process led to the generation of remarkably linear belts (Sutton, 1977) and segments of the lower crust were uplifted and overthrust on upper crustal blocks (Fig. 3-2a).

Internal distortion and crustal "softening" also facilitated the formation of linear fracture zones (e.g. the Great Dyke of Zimbabwe), and if these were located above asthenospheric diapirs or hot rise systems intracontinental troughs or basins could develop, which subsided slowly and collected considerable amounts of terrigenous sediments, cut by mafic dykes.

Further attenuation of the crust, mafic underplating through dense primitive mantle melts and their subsequent rise as differentiated magmas (Fig. 3-2b) now provide the necessary conditions for lower crustal or subcrustal delamination (Fig. 3-3b) to occur, and ensialic orogeny following A-subduction ensues (Fig. 3-2c). The absence of Phanerozoic-type ophiolites and blueschist belts in Lower Proterozoic mobile zones suggests that complete continental breakup, formation of marine basins floored by *extensive* oceanic crust and subsequent ocean closure through B-subduction did not occur. Instead, part of the attenuated crust flooring the ensialic "geosynclinal" basin was dragged down and the original crustal thickness was eventually restored largely by horizontal compression of the geosynclinal fill and subfluence, underthrusting or interstacking processes in the lower crust (Fig. 3-2c). The pre-geosynclinal crystalline basement did not behave as a rigid entity during the orogenic process but was incorporated in thrust sheets

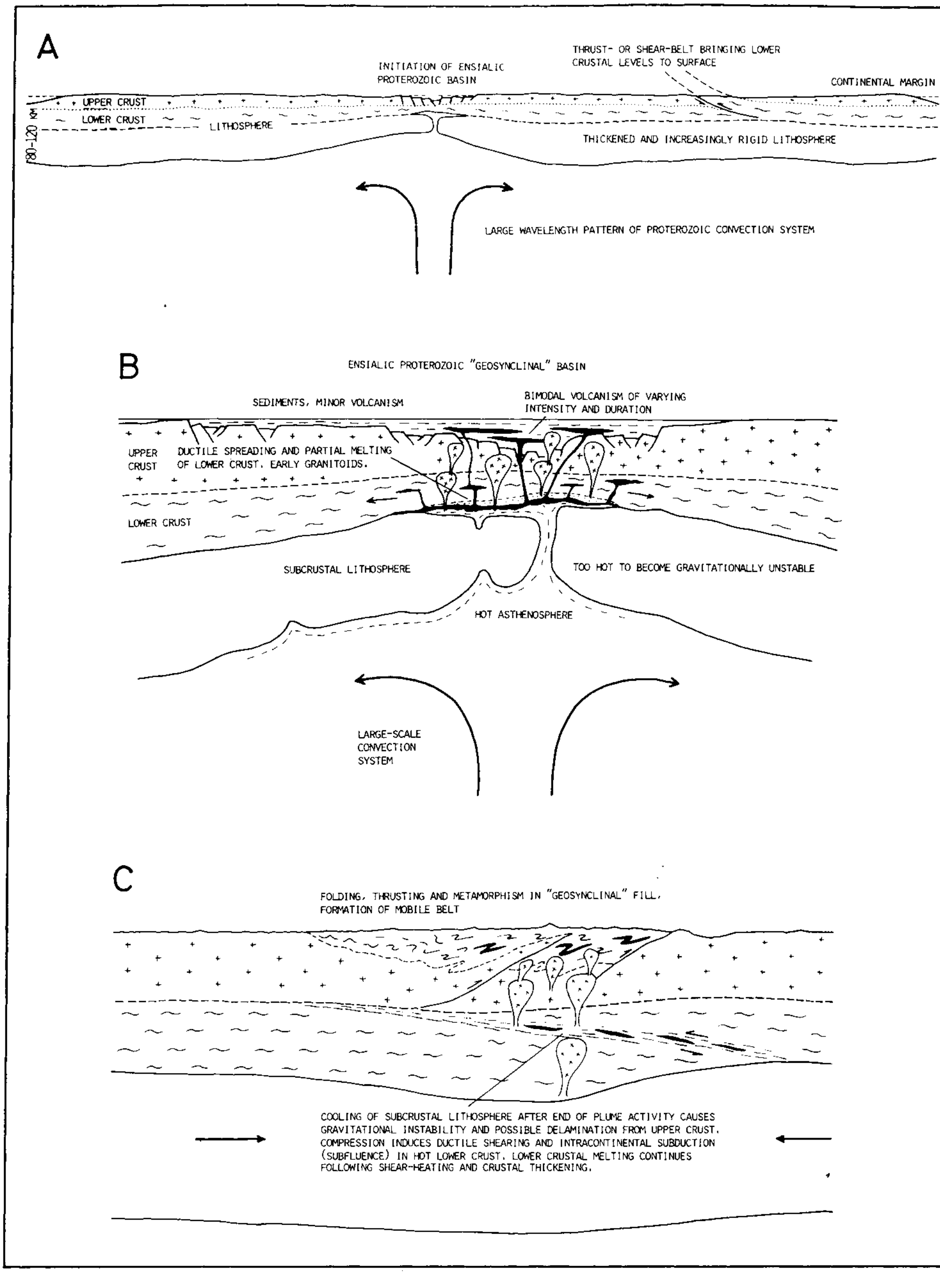

Fig. 3-2. Simplified and schematic sections showing suggested evolution of Proterozoic ensialic mobile belts. For explanation see text. Modified after Kröner (1979a).

and anatectic mobilization, thereby often concealing the original basement-cover relationships and generating polymetamorphic assemblages.

The model discussed above has the advantage that all observed Lower Proterozoic continental settings can be accommodated in a global scenario reflecting variable degrees of crustal stretching and intracrustal distortion. Its admitted weakness is that nothing is said about the oceans which must have existed at that time and the oceanic crust of which must have been destroyed completely by mechanisms which we are as yet unable to decipher from the rock record.

THE UPPER PROTEROZOIC (1.2—0.6 Ga)

Towards the end of the Precambrian another major change in global tectonics may have occurred, perhaps related to further convective cooling, changes in the convective system as well as thickening and increased rigidity of the lithosphere. The recognition of rock assemblages characterizing both passive and active continental margins as well as island-arc environments, together with the identification of true ophiolites and high-pressure mineral facies strongly suggests operation of modern-type Wilson-cycle processes (Kröner, 1979b; Strong, 1979; Gass, this volume, Chapter 15; Leblanc, this volume, Chapter 17). The transition to the contemporary plate-tectonic regime was caused by the onset of subduction of oceanic crust, by this time sufficiently thick and cool at boundaries with continental plates to become negatively buoyant and subside (Baer, 1977; Lambert, this volume, Chapter 18). Initiation of B-subduction at continental margins may have been facilitated by the formation of thick sedimentary wedges.

Gravitational pull of the subducting oceanic lithosphere and generation of new oceanic crust at spreading ridges greatly increase the mobility of rigid lithospheric plates which now follow a large-scale convection pattern in the mantle and whose movement can be portrayed by present-day plate geometry (Dewey, 1975) and reasonably well known apparent polar wander paths (McWilliams, 1981). The operation of Wilson-cycle processes is documented by several suture-bounded Upper Proterozoic ophiolites, some of which closely resemble Phanerozoic layered oceanic crust (Leblanc, this volume, Chapter 17; Strong, 1979; Gass, this volume, Chapter 15; El-Sharkawi and El-Bayoumi, 1979), the presence of island-arc systems (Gass, this volume, Chapter 15) and the isolated occurrence of possible high-pressure metamorphic assemblages containing glaucophane (Kröner, 1974).

However, not all Upper Proterozoic to Lower Palaeozoic orogenic belts fit the pattern of Wilson-cycle evolution (Wells et al., 1970; Kröner, 1977b, 1979a), and many are better explained by the model of ensialic orogeny and A-subduction as detailed above. For example, the Damara, Katanga and West Congo belts of Africa all contain thick miogeosynclinal and eugeosynclinal sediments but no ophiolites, calc-alkaline volcanics and blueschists.

Their evolution has been explained by various degrees of continental fissuring and subsequent horizontal compression during the Pan African tectonothermal event (Kröner, 1977b, 1979b; Martin and Porada, 1977). An ensialic evolution is particularly convincing for the West Congo belt whose northern end is underlain by continuous autochthonous older basement. Palaeomagnetic data support such interpretations (McWilliams et al., in prep.).

The difference between these ensialic Pan-African belts and older Lower Proterozoic mobile zones is evident in the fact that the former contain well developed thick geosynclinal sequences, signifying the formation of mature passive continental margins within wide basins bordering stable cratons, and sometimes even indicate limited crustal rupture with the emplacement of primitive tholeiitic lavas and mafic to ultramafic complexes (Kröner, 1979c; Strong, 1979). Also, the intensity of deformation, basement reactivation and thrust and nappe tectonics indicate greater horizontal shortening than in Lower Proterozoic belts and all these features are in accord with greater lithospheric mobility during Upper Proterozoic times.

Upper Proterozoic tectonics — transition to contemporary plate interaction

The large-scale generation and consumption of oceanic crust that characterizes modern global tectonics began in the late Precambrian and, by the beginning of the Phanerozoic, had transferred orogenic evolution from predominantly intraplate settings to plate-margin activity. The transition took place during a time period of almost 600 Ma from about 1100 Ma ago to about 500 Ma ago, and during this time all variations from Lower Proterozoic-type orogenic development to Wilson-cycle tectonics occurred. (Kröner, 1980b). However, it would appear that incipient crustal breakup and A-subduction still dominated the scene, and the following typical developments are suggested by the available rock record.

(1) Lithospheric stretching and crustal fissuring as detailed before form grabens which may evolve into aulacogen-like settings (Martin and Porada, 1977)(Fig. 3-3a). This weakening process causes spontaneous gravitational instability and detachment of dense mantle lithosphere. The model of Toksöz and Bird (1977b) and Bird (1978) provides a thermal and mechanical scenario which leads to insertion and "underplating" of less viscous and lighter asthenosphere below the delaminated crust (Fig. 3-3b). This causes significant heating, thereby facilitating further crustal attenuation, differentiation and ascent of magmas from the underplated primitive mantle material and crust-mantle mixing. This process may continue until a deep "geosynclinal" basin has formed and, in the extreme case, it may lead to continental breakup, initiation of sea-floor spreading and generation of Red Sea-type oceans floored by layered oceanic crust. This is not unlike the early evolution of the Alpine basin (Hsü, 1979; Trümpy, 1969).

Further downward pull of the delaminated lithospheric slab will then

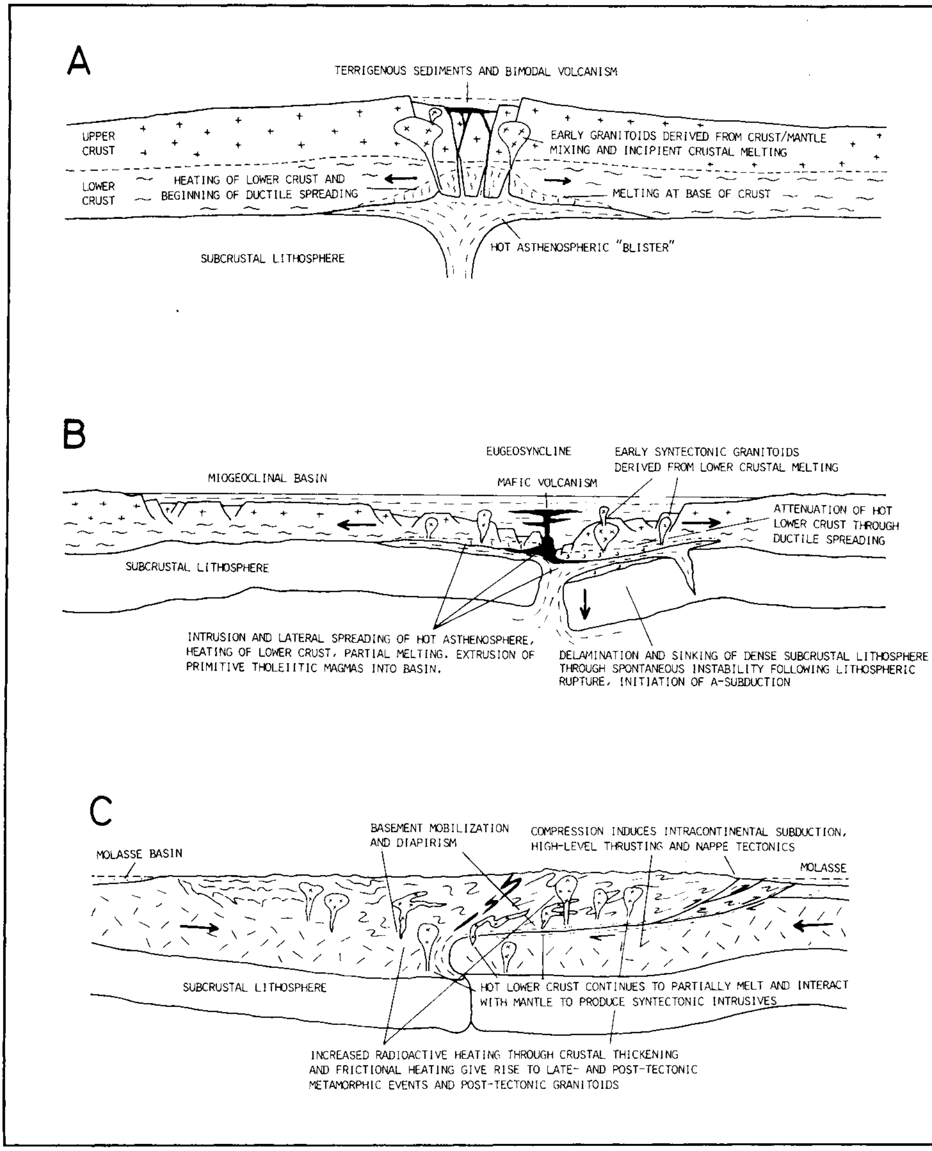

Fig. 3-3. Simplified and schematic sections showing suggested evolution of late Precambrian belts through crustal thinning, delamination and intracrustal subduction. For explanation see text. Modified after Kröner (1979a).

cause continental convergence so that the greatly attenuated crust is thickened again by A-subduction. Crustal shortening results in folding of the geosynclinal fill and its underlying "softened" and partly mobilized basement. The whole stack is then intruded by syn- and post-tectonic granitoids and uplifted so that molasse sequences form in marginal cratonic basins (Fig. 3-3c).

It is also possible that shallow A-subduction or thin-skinned tectonics (Hsü, 1979) adds to the crustal shortening and produces typical collision settings such as large thrusts and nappes and high-pressure assemblages (Bally, 1981).

In cases where continental separation did not succeed, no ophiolite assemblages have formed in the ensialic basins. Since continental convergence was not accompanied by consumption of oceanic lithosphere, the typical island-arc volcanism could not develop. Examples of such environments are the Pan-African Damara, Katanga, West Congo and Dahomey belts (Kröner, 1977b, 1979b, 1980b).

(2) Intracontinental distortion such as envisaged for the Lower Proterozoic may have continued well into the Upper Precambrian and, in analogy with the Recent intraplate tectonics of Asia, may be interpreted as a result of continental collision or the reaction of large rigid plates to rapid intermittent drift. Such cases have been described from northeast Africa (Kröner, 1979c; Ball, 1980) and Australia (Duff and Langworth, 1974), where the formation of the Amadeus basin with its well documented entirely ensialic evolution may be a particularly convincing example of this type of intracontinental thrust or shear belt (Wells et al., 1970).

(3) Ensimatic island-arc evolution and extensive lateral continental growth through subduction in the late Proterozoic has so far been postulated only in the Arabian—Nubian Shield (Greenwood et al., 1976; Gass, this volume, Chapter 15) although similar speculations have been made for the late Precambrian development of the western margin of the Canadian Shield (Badham, 1978).

For Arabia all investigators agree that large volumes of predominantly calc-alkaline material of juvenile origin have been generated during the time period $\sim$1200 Ma to $\sim$600 Ma ago although proposed models of how this crust grew differ considerably in detail. Gass (this volume, Chapter 15) summarizes the interpretation that favours intra-oceanic growth and eventual emergence of several arc systems which collided along ophiolite-decorated sutures and were "welded" onto the African continent during a Pan-African cratonization event. Others have compared the Arabian evolution with an Andean or Cordilleran situation and proposed a continuous eastward growth of the Arabian—Nubian Shield by accretion resulting from westward subduction of oceanic lithosphere (Schmidt et al., 1978; Garson and Shalaby, 1976; Kröner, 1979c).

Stern (1979), however, rejects these concepts on the basis of work in Egypt in arguing that no paired metamorphic belts with blueschists are developed, that granitic plutons are scattered randomly across the Shield and that stratigraphic and structural data argue against a direct analogue with modern island-arc and marginal basin systematics. He favours a Red Sea-type ocean opening and closing model but supports the contention that some form of Wilson-cycle tectonics was in operation. The recognition of an early Pan-African continent-derived sedimentary sequence in the southern Arabian Shield (BaSahel, Jackson, Kröner and Ramsey, unpubl. data) may indicate the presence of at least some pre-Pan-African sialic crust, and lateral accretion models may therefore not be the only alternative in explaining the evolution of northeast Africa and Arabia.

Whatever the detailed mechanism, the Arabian Shield offers a unique possibility to study the onset of Benioff subduction-related continental growth and Upper Proterozoic ophiolite obduction and provides clear evidence that the Wilson cycle was fully established at the end of the Precambrian.

CONCLUSIONS

There is little doubt that global crustal evolution was non-uniformitarian. Major changes in the tectonic development of the crust at the end of the Archaean (~ 2700—2500 Ma ago) and at the end of the Proterozoic (~ 1200—600 Ma ago) are apparent and may be related to slow cooling in the mantle. This was accompanied by a change from small-scale turbulent convection to large-scale organized convection whereby the early hot-spot regime gradually developed to a stage where ocean-ridge systems and rigid plate margins could form. At the same time lithospheric thickness increased and modern-type plate-tectonic processes may have been triggered by gravitational drag and delamination of dense lithosphere from overlying crust.

It is speculated that extensive global motion of large plates was impossible during the early Precambrian since the asthenospheric conveyor belt system was too small and too irregular to cause movement of large lithospheric slabs. The change from relatively small Archaean greenstone belts to large Proterozoic mobile belts may be related to size changes in the convective system and to an increased rigidity of the thickening lithosphere.

The motion of large and rigid crustal plates since the Proterozoic must have had an important consequence for the distribution of intraplate stresses (Oxburgh and Turcotte, 1974), particularly during times when the direction of plate movement and with that the angular velocity of individual points changed rapidly. The resulting limited horizontal displacements between parts of the same continental mass (transcurrent faulting), in conjunction with areally extensive vertical movements, are seen as causes of Precambrian tectogenesis, and Proterozoic mobile belts show that variable amounts of crustal spreading were also involved.

Dewey (1975) has shown that transform faulting, crustal distension and compression can all be related to changes in the position of finite motion poles around which plates rotate during their continuous global rearrangement. The resulting geological relationships are complex and, for instance, allow the evolution of an intraplate transform fault into a wedge-shaped aulacogen and finally into a continental margin.

For the late Precambrian (Upper Proterozoic) period the available data from all continents may be interpreted in terms of repeated attempts of the crust to break apart as exemplified by the African plate (Kröner, 1979b): In some areas this process failed (West Congo and Zambezi belts), in others it nearly succeeded (Damara and Dahomey belts) while in still others small Red Sea type oceans may have formed (Caby et al., this volume, Chapter 16; Leblanc, this volume, Chapter 17) or the full Wilson cycle came into operation (Arabian Shield, Gass, this volume, Chapter 15).

It is not clear whether the unusually high rate of apparent polar wander for Gondwana segments during the Pan-African period (at least 1°/Ma as compared to about 0.3°/Ma for earlier times, McElhinny et al., 1974) is the result of increased plate motion due to the onset of contemporary Wilson-cycle development or whether this mobility was only a contributing factor to those listed above in triggering modern large-scale consumption of oceanic crust.

Contemporary plate-tectonic style reflects the *present* response of the crust to mantle convection, and I feel that the New Global Tectonic Theory is flexible enough to incorporate the changing tectonic processes as reflected by the Precambrian rock record.

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Chapter 4

UNIFORMITARIAN ASSUMPTIONS, PLATE TECTONICS AND THE PRECAMBRIAN EARTH

A. Y. GLIKSON

ABSTRACT

The scientific methodology which has underlain the development of continental drift and plate tectonic concepts argues against their extension into earlier chapters of earth history as a uniformitarian assumption. By contrast, multidisciplinary synthesis of the Precambrian data themselves, following a parsimony principle, is capable of determining the distribution of rocks in place and time and of defining the nature of their source and differentiation history. Geochemical and isotopic data have been widely applied to the identification of magma source regions, but only palaeomagnetic measurements offer definitive indication regarding the existence of plate-tectonic regimes. From present evidence, pending the extension of palaeomagnetic work into Archaean terrains, operation of plate-tectonic (Wilson-cycle) processes during this era is consistent with the isotopic and geochemical evidence for two-stage mantle melting. However, there is no intrinsic reason in these data to suggest modern-type tectonic environments, and many features of the Archaean rocks suggest temporally unique conditions. The predominantly ensialic record and joint apparent polar wander paths (APWPs) observed for the 2.6—1.2 Ga era render operation of present-day plate-tectonic processes during this time unlikely, possibly with local small-scale exceptions. The appearance of ophiolites and divergence of APWPs from about 1.2 Ga ago signifies an opening of considerable oceanic crustal gaps. Temporally unique features of the Archaean earth include evidence for high-temperature melting, distinct geochemical characteristics, the indication of anomalously thick crustal segments and the dominance of vertical crustal movements. Palaeomagnetic, geological and geochemical data for the 2.6—1.2 Ga interval give rise to a major enigma with regard to the earth surface dimensions and consequently its radius in pre-Late Proterozoic time.

THE UNIFORMITARIAN DOCTRINE

Hutton in 1788, observing the great thickness of sedimentary successions, deduced the long-term operation of erosion and deposition and the vast expanse of geological time. This concept, philosophically related to Darwin's evolutionary theory, has since progressively replaced catastrophism, the theory that violent short-lived events outside our present experience have drastically effected the earth. The new trend culminated with Lyell's uniformitarian doctrine, suggesting that natural laws and geological processes applicable at present have operated in the past in the same regular manner and essentially the same intensity, as summed up by the dictum: "the present is the key to the past". According to Hutton's principle: "no powers are to be employed that are not natural to the globe, no action to be admitted except those of which we know the principle" (cited in Holmes, 1965, p. 44). This philosophy has unfortunately become a restrictive dogma at a later stage. Holmes (1965) recognized this problem and regarded uniformitarianism as an unhappy word liable to be taken too literally, stating: "Lyell's term inevitably suggests a uniformity of rate, whereas what is meant is a uniformity of natural laws". The question is of both rate and scale, namely, while fundamental processes such as volcanic activity and sedimentary deposition obviously persisted throughout earth history, the geotectonic framework, rates of vertical and horizontal movements, geothermal gradients and nature of igneous processes have demonstrably varied temporally. Neither can the role of catastrophic events such as meteorite impacts be neglected. Frequency distribution diagrams of isotopic ages plotted against time point to the strongly episodic nature of tectonic-thermal history (Dearnley, 1966; Stockwell, 1973). Temporally unique features in the geological record may easily be overlooked where uniformitarian assumptions dominate.

A principal consequence of the uniformitarian doctrine was that geological theory became constrained by the state of knowledge of physics in the 19th century. More recently, as Newtonian mechanics were superseded by modern astronomical and physical concepts, geological thinking has remained in this bind. Thus, major observations such as continental drift (Wegener, 1929; Du Toit, 1937) were ignored for many years on account of the incomprehensibility of their underlying physical factors. Although plate-tectonic theory is now widely accepted, the philosophical and methodological weaknesses which prevented an earlier acceptance of this theory remain in the way of further advances.

However, Hutton has also insisted that the way and means of nature could be discovered only by observation. The empirical approach was emphasized by Bullard, who stated (1964, p. 19): "it is usually best to decide on the existence of a phenomenon in the light of the known facts, or of a reasonable explanation of them, and then to look for a mechanism or a physical theory". Unfortunately, a reverse procedure commonly dominates current geological thinking: e.g. it is erroneously assumed that the physics of the lower mantle and core are basically understood and thus constrain conclusions derived by direct observation of crustal and upper mantle rocks. Furthermore, with few exceptions (e.g. Runcorn, 1962), a constancy of deep-seated processes throughout earth history is commonly taken for granted. Had similar constraints applied in astronomy, for instance, the observation neither of universal expansion from redshift spectral effect nor of black holes from neutron star rotation periods would have been possible, as the physics underlying these phenomena are still only vaguely perceived. If the potential of empirical observation in earth science to hint at unforeseeable physical processes is to be maintained, the data must be considered on their merit. The methodology of relevant multidisciplinary synthesis is considered below.

METHODOLOGICAL PROBLEMS IN EARTH SCIENCE

Central to the correct interpretation of empirical observations is the application of parsimony, namely the explanation of any given set of data in terms of the minimum possible number of unknown variables. In this manner scientific theory is systematically constructed from the data base and can be progressively modified, improved or retracted as more information comes to light. Where additional factors not essential to an explanation are invoked, a proliferation of uncontrolled models ensues. Where more than one explanation exists for a set of observations, which is the usual case in earth sciences, the one with the least number of unknowns should be tentatively preferred. The procedure involves the development of transient working hypotheses against which the data are repeatedly tested. By contrast, purely conceptual models and theoretical considerations can be highly misleading, unless used to formulate questions related to the data base, rather than provide hypothetical answers unrelated to the data. It is an alarming feature of Precambrian studies that this class of models is still propagated and that little consensus exists regarding methodology — a situation which has no parallel in physics and chemistry, where logical formulations and repeatable experiments are prerequisites.

Mitroff (1974) investigated the methodological questions associated with NASA's lunar rocks study and made the observation that, in multidisciplinary studies of this type, a proliferation of models is an inherently necessary development since usually no single mind does objectively master, assess and verify all the evidence for several contradictory hypotheses. Workers supporting what are eventually shown to be mistaken concepts thus render as important a service to science as those who prove correct, as in this way the truth should come out in the wash. Where the conditions for debate do not exist and basic scientific ethics are not observed, dominant dogmas take over scientific progress.

Any investigation of complex natural phenomena requires that both the limitations of individual methods and their implications to one another are recognized. It is common practice for workers to extrapolate observations from a single terrain or a single method into other areas or subjects. In the following some limitations and weaknesses of methods applied to the spatial and temporal definition of Precambrian rock units are considered.

The order of superposition of structural elements, commonly used for deciphering tectonic evolution within rheologically homogeneous domains, are not readily applicable to inter-domain correlations. It can be shown that differences in the rheological characteristics and structural behaviour of distinct rock types in separate blocks render such correlations tenuous

(Glikson, 1978). In the general absence of fossils in Precambrian strata, lithological similarities of sedimentary and volcanic rocks have been widely used for unit-time correlations. However, except where lithostratigraphic units can be continuously traced or closely analogous sequences demonstrated in adjacent blocks, such correlations are uncertain. The use of dykes as reference markers defining older (intruded) and younger (unintruded) terrains (McGregor, 1973) is also doubtful, due to controls of jointing and dyke emplacement by the differential rheological properties of the country rocks.

The geological significance of thermal events recorded by isotopic systems is not always evident. Secondary reset of Rb-Sr and K-Ar whole-rock and mineral systematics during low-temperature metamorphic episodes may or may not be identified by mineral replacements and structural imprints and thus often remains geologically cryptic. U-Pb zircon and apatite ages may represent igneous events or inherited source ages, a distinction not always evident. Initial $^{87}Sr/^{86}Sr$ ratios (Ri) may provide constraints on the crustal prehistory of the measured rocks only if the Rb/Sr ratio of their precursor materials can be estimated with confidence. Low-Ri silicic igneous rocks are commonly interpreted in terms of their derivation from basic mantle-type parents (Moorbath, 1977). However, Ri values alone can not discriminate between this possibility and that of anatectic melting of low Rb/Sr silicic granulites, tonalites or trondhjemites. An effective discrimination between the two possibilities is obtained from trace element and, particularly, rare earth element (REE) data (Glikson, 1979a). Initial $^{207}Pb/^{206}Pb$ ratios are capable of defining the μ value ($^{238}U/^{204}Pb$) of parental materials and the time of U/Pb fractionation but have little diagnostic value with regard to the overall petrological and geochemical nature of the source, i.e. diverse geological materials may have similar μ values or similar rocks may have highly different μ values.

Attempted correlations between trace-element abundances and the tectonic setting of modern volcanic suites are beset by severe difficulties (Jaques et al., 1978) which pertain all the more to ancient environments. Nevertheless, when combined, the isotopic and geochemical data yield key information on the nature of source regions, partial melting and magmatic differentiation.

Measurements of remanent magnetic fields, when combined with precise age determinations (McElhinny, 1973), constitute the only method capable of providing definitive information on the relative location and possible motions of crustal plates. Many Precambrian geologists, geochemists and geochronologists have discussed Precambrian plate tectonics with little or no reference to the palaeomagnetic information available for Proterozoic terrains. One reason for this neglect are conflicting interpretations of the palaeomagnetic poles, as exemplified by variable selections of north and south poles as a basis for apparent polar wander paths (APWPs). Another question relates to the uncertainties of isotopic ages in relation to the timing of magnetization. The errors in pole measurement and age tolerance may together obliterate the significance of an APWP; nevertheless, the observed alignment of poles in accordance with sequential ages on APWPs cannot be dismissed as coincidental and must reflect a close relation between thermal events and resetting of major magnetic components. Observed systematic pole-age alignments along similarly shaped APWPs which characterizes craton clusters within the African, Laurentian and Australian Precambrian shields (Irving and Park, 1972; McElhinny and Embleton, 1976; McElhinny and McWilliams, 1977) is statistically unlikely to be accidental. This evidence must be considered as valid as, for example, systematic sea-floor magnetic lineaments or isotopic isochrons. Palaeomagnetic implications for the question of Proterozoic geotectonics are discussed later below. The extension of palaeomagnetic studies into Archaean terrains provides the best hope for elucidation of the possibility of plate motions during this era.

Despite their limitations the above methods, when combined, are capable of solving the major questions of Precambrian evolution. The current proliferation of uncontrolled models clearly reflect a failure to observe basic methodological principles and a failure to study the implications of each method on the others by parsimonious multidisciplinary analysis.

POSSIBLE RELEVANCE OF PLATE TECTONICS TO PRECAMBRIAN HISTORY

Following the wide acceptance of plate-tectonic theory it has become fashionable to interpret a variety of structural lineaments, faults, shears, elongated intrusions, volcanic belts, metamorphic isograds, linear geochemical anomalies and magnetic and gravity lineaments in Precambrian terrains as ancient plate boundaries (Katz, 1972; Hoffman, 1973; Dewey and Spall, 1975; Walker, 1976; Burke et al., 1976). In many instances the occurrence of volcanic rocks is being taken as evidence in its own right for underlying Benioff zones. The assumption of a plate-tectonic model is also implied by comparisons between Precambrian terrains and arc-trench, back-arc spreading and Cordillera systems (Tarney et al., 1976; Windley and Smith, 1976; Tarney and Windley, 1977). This approach has commonly ensued in a deus-ex-machina solution for all Precambrian problems; however, it is often only loosely related to the data base and has come at the expense of detailed considerations of the Precambrian records themselves.

Arising from any theory are predictions whose confirmation or otherwise serves to test its validity. Had horizontal sial—sima plate interactions been widespread during the Precambrian, geochemical and isotopic signatures similar to those associated with modern consuming plate margins should be extensively retained in the geological record. Further, had the sialic crust grown by means of accretion of island-arc and Cordillera chains, concentric age zonations should be commonly manifest in Precambrian terrains. Independent plate migrations would have been registered by divergent and convergent palaeomagnetic APWPs. The applicability of these criteria is considered below.

Ophiolites

The occurrence of oceanic crustal segments in Phanerozoic orogenic belts offers a confident criterion for sima—sial interaction. Had oceanic crust existed throughout the Precambrian, it could be expected to be locally preserved. Stratigraphically low komatiite-tholeiite volcanic assemblages which constitute the oldest units in Archaean granite-greenstone terrains are considered to be derived from an early Archaean ultramafic-mafic crust (Glikson, 1971; 1976). However, 2.6—1.2 Ga old terrains are remarkably free of proven simatic crustal relics, with minor exceptions such as in the Churchill Province in Manitoba, although it is possible that some mobile belts originally formed above intercratonic simatic gaps. True ophiolites are found in late Proterozoic post-1.2 Ga old mobile zones such as the Red Sea and Namibian belts (Greenwood et al., 1976; Kröner, 1979; *see also Gass, this volume, Chapter 15, Leblanc, this volume, Chapter 18, ed.*).

Two-stage mantle melting products

Inherent in plate-tectonic processes is subduction and partial melting of large volumes of oceanic crust and production of basic magma in overlying mantle wedges, giving rise to belts of andesite, sodic dacite, sodic rhyolite and minor shoshonite and their dioritic, tonalitic and trondhjemitic plutonic counterparts. This class of silicic igneous rocks, characterized by low Ri values, heavy REE-depleted and light REE-enriched patterns, high Na/K and low LIL (Large ion lithophile) elements, is dominant in the Archaean (Glikson, 1979a). Only a minor proportion of pre-2.6 Ga old rocks consist of high-K and alkaline volcanics. By contrast, early and middle Proterozoic terrains are commonly dominated by eutectic K-rich granites, adamellite, granodiorite and their extrusive equivalents, characterized by intermediate to high Ri values, high LIL-element levels and negative europium anomalies (Glikson, 1980). Late Proterozoic Pan-African mobile belts are marked by granodiorite and diorite of commonly, but not always, low Ri values. Andesites are common in some Archaean greenstone belts (i.e. Superior Province, Midlands belt in Zimbabwe (Rhodesia), Marda belt in Western Australia) and, less common though not absent, in Proterozoic terrains.

Lateral accretion

The broad age zonation of Precambrian shields consists of Archaean cratonic nuclei surrounded and intersected by successively younger Proterozoic mobile belt networks, a pattern originally interpreted in terms of continental growth (Engel, 1963). It is now established that many mobile belts overlie and contain blocks of reworked Archaean sial and thus cannot be equated with lateral continental accretion of the type developed along the western Pacific margins, for example. Had sialic crust accreted along consuming plate margins, as suggested by island-arc models of continental evolution (Jakes, 1973), there would have ensued a concentric age zonation of volcanic-sedimentary belts formed by two-stage mantle melting. This is observed neither in Archaean nor in Proterozoic terrains although age zonation is suggested for the Baltic Shield (A. Berthelsen, pers. commun., 1980). It appears that the Archaean crust evolved by progressive diachronous nucleation of batholiths rather than by lateral accretion, although available isotopic age data are far from sufficient to demonstrate this point unequivocally.

The above summary suggests that the petrological, geochemical and isotopic characteristics of Archaean granite-greenstone terrains are consistent with plate-tectonic processes, whereas the bulk of observed early and middle Proterozoic igneous processes have occurred predominantly in ensialic environments. However, to date there is no structural, palaeomagnetic or age zonation evidence to indicate whether the Archaean sima-to-sial transformation process has taken place in ancient analogues of modern arc-trench domains or in temporally unique tectonic regimes. There is no intrinsic reason to assume actualistic models in this regard, and it is likely that the higher heat production and more active mantle convection systems (Lambert, 1976) resulted in temporally unique tectonic crustal patterns (*see also, Hargraves, this volume, Chapter 2; Kröner, this volume, Chapter 3; Lambert, this volume, Chapter 18, ed.*). This concept is consistent with observations of the structural style of granite-greenstone terrains, i.e. the dominance of vertical tectonics (Anhaeusser et al., 1969; Hickman, 1975). The palaeomagnetic constraints on horizontal plate movements during the early and middle Proterozoic (see below) may well hint at a lesser role of plate tectonics during the Archaean, though this remains unknown pending the extension of palaeomagnetic work into Archaean terrains.

PALAEOMAGNETIC EVIDENCE AND PRECAMBRIAN GEOTECTONICS

Palaeomagnetic studies of Proterozoic terrains in North America, Africa and Australia demonstrate a maintenance of relative spatial positions of cratons within each of these continents during the period 2.6—1.0 Ga ago (Irving and Park, 1972; Piper et al., 1973; McElhinny and Embleton, 1976). Within each of these shields pole data from the various cratons allow the delineation of a single APWP. This observation was questioned by Burke et al. (1976) who suggested that the pole paths of each of the cratons differed in its detailed shape. In replying to this criticism McElhinny and McWilliams (1977) have shown that the differences arise from gaps in the available data and thereby also in the continuity of the APWPs. They argued that, as the data from all the cratons within each of the shields conform to a single path, relative horizontal plate displacement cannot be invoked. The tolerance of the APWP (15°) allows for separation of about 1000 km but plates would have to return to orginal positions to maintain the observed path continuities.

McElhinny and McWilliams (1977) noted differences between the APWPs for North America, Africa and Australia and left the question of possible horizontal displacements between these shields unresolved. Very recently Embleton and Schmidt (1979) have made the surprising observation that, when these continents and also Greenland are maintained in their present-day angular distances from each other, their APWPs for the time interval 2.3—1.6 Ga ago overlap closely. This coincidence not only rules out relative plate migrations of these shields during the above period but gives rise to major questions regarding the continent—ocean distribution pattern during the early and middle Proterozoic and the radius of the Precambrian earth, as discussed below.

TEMPORALLY UNIQUE PRECAMBRIAN FEATURES

The Precambrian crustal record contains abundant evidence for temporally unique features which are unrecognized, or only partly recognized, in Phanerozoic terrains. Archaean examples include:

(1) Occurrence of peridotitic komatiites (above 20% MgO) which are not known from post-2.6 Ga old terrains; abundance of basaltic komatiites (high-Mg basalts).

(2) Occurrence of near-liquidus to superheated tonalite/trondhjemite magmas as testified by the assimilation of large volumes of mafic-ultramafic greenstone materials by the Archaean batholiths (Glikson, 1979a; Anhaeusser, this volume, Chapter 6).

(3) Dominance of a bimodal low-K-tholeiite/Na-dacite volcanic suite and rarity of andesites.

(4) A near-absence of alkaline igneous rocks, except as minor occurrences at the top of sedimentary late Archaean sequences.

(5) Little evidence for ensialic anatectic processes in the period from 3.9 to 2.6 Ga ago, as evidenced by low Ri values and generally low LIL element abundances of Archaean silicic igneous rocks.

(6) Diachronous diapiric rise of Na-rich plutons and paucity of evidence for modern-type lateral sialic accretion.

(7) Evidence for anomalously thick sialic crustal regions (40—80 km) from metamorphic mineralogy (O'Hara, 1977; Tarney and Windley, 1977).

(8) Evidence for high-pressure magmatic fractionation of garnet and possibly eclogite from heavy REE-depleted patterns. Modern arc-trench volcanics generally display lesser REE fractionation.

(9) High abundances of transition metals (Ni, Cr, Co) in Archaean andesites and some dacites as compared to modern equivalents.

(10) Evidence for high Fe/Mg ratios in the Archaean mantle (Glikson, 1979b).

Some temporally distinct features of Proterozoic terrains include:

(11) Polygonal intrasialic craton—mobile belt development not recognized during the Mesozoic and Cainozoic (*e.g. Baragar and Scoates, this volume, Chapter 12; Dimroth, this volume, Chapter 13, ed.*).

(12) Extensive sialic crustal reworking along mobile loci and throughout large regions of the crust where silicic granulites, Rapakivi granites and anorthosites abound (Bridgwater and Windley, 1973).

(13) Lack of evidence for existence of extensive simatic crustal regimes during the early and middle Proterozoic (2.6—1.2 Ga ago) (Glikson, 1979c, 1980; Kröner, this volume, Chapter 3).

(14) Appearance of ophiolites and two-stage mantle melting products from about 1.2 Ga ago in some Pan-African mobile belts (Greenwood et al., 1976; Kröner, 1979).

Other features on a variety of scales can be pointed out to demonstrate the distinction between the tectonic, petrological and geochemical characteristics of every stage of earth history, in contradiction of the uniformitarian assumption. Moorbath (1977) used the isotopic and geochemical evidence for the operation of two-stage mantle melting processes as a basis for a new doctrine of "episodic uniformitarianism", namely a view of crustal evolution as a series of major accretion—differentiation episodes taking place in environments akin to modern circum-Pacific domains. In this model juvenile mantle-derived additions to the sial dominated over sialic reworking. However, as the geochemical evidence indicates, ensialic anatectic silicic rocks prevail in many Proterozoic terrains. Further, the operation of two-stage mantle melting does not in itself constitute evidence for a modern tectonic setting, as discussed above. Although many similarities exist between Archaean granite-greenstone terrains, Sierra Nevada Palaeozoic batholiths (Hietanen, 1975) and Chilean Mesozoic granodiorite-greenstone back-arc terrains (Tarney et al., 1976), there is no evidence for the existence in the Archaean of large continents along which the Cordilleran orogenic belts have developed.

In attempting to understand Precambrian crustal evolution, major questions arising from the Precambrian data themselves include the following:

(1) It is known that the moon and the terrestrial planets were bombarded by major meteorite swarms before 3.8 Ga ago (Schmitt, 1975). Although terrestrial rocks of this age are known (Moorbath, 1977), no direct evidence for such impacts has as yet been observed. It is possible that early ultramafic-

mafic volcanics which occur at the base of greenstone belts reflect long-term mantle diapirism initially triggered by such events (Green, 1972).

(2) The geotectonic significance of the 2.7—2.6 Ga isotopic age peak is not clear. The major thermal events which mark the end of the Archaean signify a transition from an era dominated by sima-to-sial transformation to an era dominated by ensialic processes during which there is little evidence for existence of simatic crust. It is considered likely that the Archaean—Proterozoic boundary represents a culmination of Archaean processes in global cratonization (see below).

(3) The possibility of global cratonization at the end of the Archaean gives rise to a major enigma regarding the surface dimensions of the Precambrian earth (Glikson, 1979c, 1980) as discussed below.

The total surface area of the Precambrian crust is about 80% of the total area of the present continents (Goodwin, 1974), i.e. less than one quarter of the present surface of the earth. It is impossible for the sialic crust, therefore, to have ever extended over most of the earth's surface if a constant radius is assumed, as this is contrary to volumetric constraints, estimates of Precambrian crustal thicknesses, common preservation of original structural features and palaeomagnetic data (Glikson, 1979c). Thus, any constant-radius tectonic models for the Proterozoic earth must allow for a simatic composition for at least three quarters of its surface. Since the coincidence of the 2.3—1.6 Ga old APWPs of North America, Greenland, Africa and Australia demonstrated by Embleton and Schmidt (1979) can hardly be accidental, it would appear that, on a constant-radius earth, major oceanic crustal regimes coincided in the early to middle Proterozoic with the modern location of the Atlantic, Indian and Pacific ocean basins. However, this is extremely unlikely for the following reasons:

(1) The destruction of Proterozoic oceanic crust along consuming plate margins around the three ocean basins would have resulted in circum-Pacific-type belts. Such are missing around the Atlantic and Indian Oceans. Proterozoic belts around these basins contain little evidence for extensive long-term two-stage mantle melting processes. Further, there is no evidence for destruction of Precambrian sima around the Pacific Ocean. Since higher heat flow and consequently more active sea-floor spreading can be reasonably assumed in Precambrian oceanic regimes, major volume problems arise if accretion occurs without accompanying subduction (Glikson, 1980). Considerations pertaining to the Precambrian thermal regime do not allow subduction of oceanic crust unaccompanied by partial melting and production of two-stage mantle melting materials, which are scarce in Proterozoic terrains.

(2) In the light of the geochemical and isotopic evidence for the origin of Archaean granite-greenstone terrains by sima-to-sial transformation, had the bulk of the crust by the end of the Archaean consisted of sima, it is not possible to explain why the evolution of granite-greenstone systems

has ceased about 2.6 Ga ago. A possible answer is that all sima had been consumed. However, volume limits of the sial do not allow existence of a global continental crust on a modern-radius earth (Glikson, 1979c, 1980).

(3) Reconstructions of Precambrian craton—mobile belt patterns across the Atlantic Ocean (Hurley et al., 1967) and the Pacific Ocean (Sears and Price, 1978) suggest a continuity of middle Proterozoic mobile belts between Brazil and West Africa and between Siberia and the western part of North America. Further, the tectonic grain of greenstone belts is aligned in a parallel array on Pangaea reconstructions (Engel and Kelm, 1972), suggesting crustal continuity. Further evidence is needed to establish the integrity of the Proterozoic sial beyond doubt.

The contradictions arising from attempts to reconcile the Proterozoic geological record with a present-radius earth (Glikson, 1979c, 1980; Embleton and Schmidt, 1979) defy uniformitarian explanations. Neither the geochemical nor the palaeomagnetic enigma arise on a globe of radius about half that of the present earth. In reflecting on the immense physical unknowns inherent in an eight-fold increase in the earth's volume the original philosophical and methodological questions discussed at the outset of this paper must be borne in mind, namely, is earth science to limit its conclusions to those commensurate with the physics of the day, or to advance empirical observations from the geological data themselves and thus possibly hint at yet unexplained physical processes? Perhaps the best way to ponder this question is in the context of the history of plate tectonic theory. As summarized by Wyllie (1971, p. 256): "The theory (continental drift) suffered from a lack of quantitative data, and the type of evidence put forward was perhaps psychologically unacceptable to many geologists. Opponents of drift argued against the theory partly on the grounds that there was no satisfactory explanation as to why the continents had drifted and, furthermore, that the *known* physical properties of the earth were such that the proposed lateral migration of the continents was impossible. Proponents of drift, on the other hand, argued that geological facts should not be ignored simply because there was no explanation available for them. Then other experts disputed the geological "facts". The arguments continued in this indeterminate vein like some medieval philosophical controversy until a stalemate was reached in the 1940's. . . The controversy was revived in the 1950's by the work of P.S.M. Blackett and S. K. Runcorn on palaeomagnetism. The new evidence led many geophysicists to consider the theory of continental drift seriously, while many geologists remained unimpressed by, and suspicious of, this new approach." The present situation with regard to the expanding earth theory is strikingly similar to the state of continental drift theory in the 1950's.

The "true story" of Precambrian crustal evolution (Moorbath, 1977, p. 181) will not arise from the study of any single terrain or by any single method but as a result of multidisciplinary synthesis combined with open discussion of the complex issues involved. The lesson for Precambrian geologists from the philosophical and methodological problems inherent in the history of plate tectonic theory is evident. The demonstration of the theory for Mesozoic and Cainozoic systems has been a triumph of empirical observation over assumed theoretical constraints. Major breakthroughs should take place in Precambrian research when the same approach is applied.

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2. Archaean tectonics

Chapter 5

ARCHAEAN PLATES AND GREENSTONE BELTS

ALAN M. GOODWIN

ABSTRACT

Archaean greenstone belts lie in granite-greenstone terrains, contain mafic volcanic rocks and feature low- to medium-grade metamorphic assemblages, yet otherwise display great diversity in size, form, lithology, stratigraphy, age, basement-cover relations and structural deformation as illustrated by consideration of four main granite-greenstone terrains respectively in Peninsular India, southern Africa, western Australia and the southern Canadian Shield.

Archaean greenstone belt diversity demands equally flexible tectonic development. Consideration of Archaean relations and constraints on a global scale leads to a preferred rift-and-sag model involving repeated attenuation and fissuring of pre-existing sialic crust within a time span of at least 1200 Ma with both ensialic and ensimatic accumulation of assorted greenstone assemblages.

Greenstone belt development is attributed to an Archaean plate-tectonic process that involved limited movement of numerous small plates operating under a high geothermal gradient and thereby lacking conventional subduction (Benioff) zones. Thus the Archaean plate-tectonic process differed in significant respects from modern plate tectonics, yet produced similar volcanic and plutonic products. It is contended that during earth history the plate-tectonic process has taken several forms, all involving at least some moving plates, of which the current is but the latest.

INTRODUCTION

Archaean greenstone belts are present in all Precambrian shields of the world. An old field term applied primarily to any compact dark-green, altered mafic to ultramafic igneous rock owing its colour to secondary minerals, the greenstone belt is a deformed elongate low- to medium-grade metavolcanic and metasedimentary unit dispersed in predominantly granitic rocks, the assemblage collectively termed a granite-greenstone terrain. Greenstone belts vary greatly in size, shape, age, lithology, stratigraphy and degree of metamorphic alteration and structural deformation. They represent preserved relicts of larger original supracrustal entities which were variably reduced by processes of erosion, deformation and igneous intrusion.

Archaean greenstone belts are products of once common crustal environments since largely superseded in a constantly evolving earth. Limited in size and number, they are by far the predominant supracrustal survivors of earth's early growth. The modern plate-tectonic paradigm involving interplay of large rigid plates — the global jigsaw — has profoundly influenced geologic thought. The Archaean tectonic analogue and, by inference, creator of Archaean greenstone belts, also apparently involved horizontally moving albeit comparatively small crustal plates. Such an Archaean analogue would then qualify as a plate-tectonic process despite the apparent absence of Archaean subduction (Benioff) zones, a key feature in modern plate tectonics. It is contended, then, that plate tectonics has taken several forms, all involving moving plates, of which the current is but the latest.

Greenstone belts of the world, as presently loosely defined, cover a broad spectrum of types and associations not necessarily restricted to Archaean or even Precambrian times. However, certain types of greenstone belts as integral parts of granite-greenstone terrains are characteristic of earth's earlier crust as considered below.

Archaean greenstone belts have two points in common, the first a matter of definition and observation and the second, of interpretation as herein developed: (1) all contain mafic volcanic rocks; and (2) they are genetically attributed to plate movements.

Four granite-greenstone terrains are considered in illustration: Peninsular India, southern Africa, western Australia, and southern Superior Province of the Canadian Shield. Abbreviated descriptions only are provided for the most part in view of extensive treatment elsewhere, including this volume.

GREENSTONE BELTS

Peninsular India

Setting

The Archaean granite-greenstone terrain of south-central Peninsular India (Fig. 5-1.1) is an irregular square 205,010 km^2 in area in which granite: greenstone $\simeq$ 4:1. It is overlain by Deccan Traps to the north; mainly by Proterozoic platforms of the Cuddapah—Vindhyan basins to the east; by a narrow discontinuous Phanerozoic strip bordering the Arabian Sea to the west; and to the south by the Charnockite Mobile Belt of Tamil Nadu—Kerala that may include highly metamorphosed greenstone belt equivalents.

The terrain includes approximately 23 NNW-trending greenstone belts distributed unevenly in the Peninsular Gneiss, a migmatitic gneiss complex of tonalite-granodiorite composition (Balasubrahmanyan, 1977) with accompanying granitoid intrusive suite. The greenstone belts range from small units 10 km long to the 400 km long Chitradurga belt in central Karnataka. Commonly, length : width $\simeq$ 6 or 10 : 1.

The terrain is partially divided into western and eastern parts by the north-trending Closepet Granite, 340 km long by 8—30 km wide. Three distinct greenstone types are present. Those to the west, referred to as Dharwar-type greenstone belts (Ramakrishna et al., 1976), are commonly long, irregular and open units. Those to the east are typically small, narrow linear units previously referred to as Keewatin-type (Ramakrishna et al., 1976) or "true greenstone belts" (Naqvi et al., 1978), but here called Kolar-type greenstone belts. Finally, numerous small schist remnants widely distributed in the Peninsular gneiss are called Sargur schist belts (Nath et al., 1976), features two cycles, the older Bababudan Group (1200 m) and the belts are uncertain (Janardhan et al., 1979).

Dharwar-type greenstone belts locally unconformably overlie Peninsular gneiss in well-developed basement—cover relations; elsewhere, however, the greenstone belt contacts are intrusive and migmatized. In contrast, no basement has been directly observed to the Kolar-type and Sargur belts where contacts are consistently intrusive or migmatized.

Lithostratigraphy

Dharwar-type greenstone belts (*Dharwar Supergroup*) are rich in metasediments, resembling in this respect some late Archaean to early Proterozoic basinal and geosynclinal fold belts such as the Huronian belt and Labrador trough of Canada and Witwatersrand triad of South Africa (Nath et al., 1976). The belts are curvilinear and elongate as well as including some broad open types. They feature shallow-water "shelf facies" sediments at the base which develop into volcanic-rich eugeosynclinal piles above.

The common Dharwar succession, c. 8200 m thick (Ramakrishna et al., 1976), features two cycles, the older Bababudan Group (1200 m) and the younger Chitradurga Group (7000 m), separated by an unconformity.

Bababudan platform sediments unconformably overlie the basement gneiss with well-exposed unconformable relationships in at least six localities. Prominent lithologic features include basal oligomict, locally uraniferous conglomerate (Chikmagalur) and orthoquartzite with ripple marks and cross-beds (Viswanatha, 1969). The overlying members include metapelites, BIF, argillite, and mafic to felsic volcanic rocks.

The upper Chitradurga Group, in turn, contains a basal oligomict conglomerate (Talya) which locally onlaps to the basement gneiss with excellent angular unconformity (Ramakrishna et al., 1976). It is overlain by BIF manganiferous cherts, carbonates and orthoquartzite of the platformal suite, in turn overlain by the thick greywacke-volcanic suite, itself containing local BIF. This volcanism is of calc-alkaline nature ranging from olivine tholeiite to andesite with minor rhyolite (Naqvi et al., 1978). At places potassic granitoid batholiths, 2.6—2.2 Ga in age (Crawford, 1969) and carrying amphibolite and metabasalt xenoliths, are found within the belts.

However, previous Dharwar stratigraphic interpretations have placed the volcanic rocks at or close to the base of the sequence followed by a cycle of sedimentation involving the accumulation of gravels and pebbles, followed by deposition of sand, clay, carbonate sediment and ferruginous cherts (Pichamuthu, 1967).

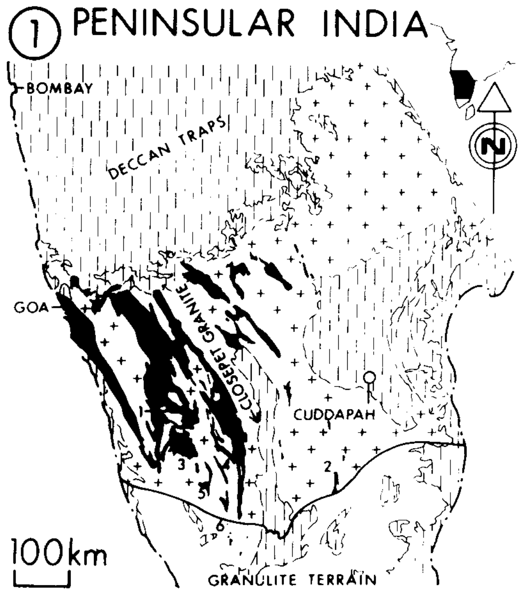
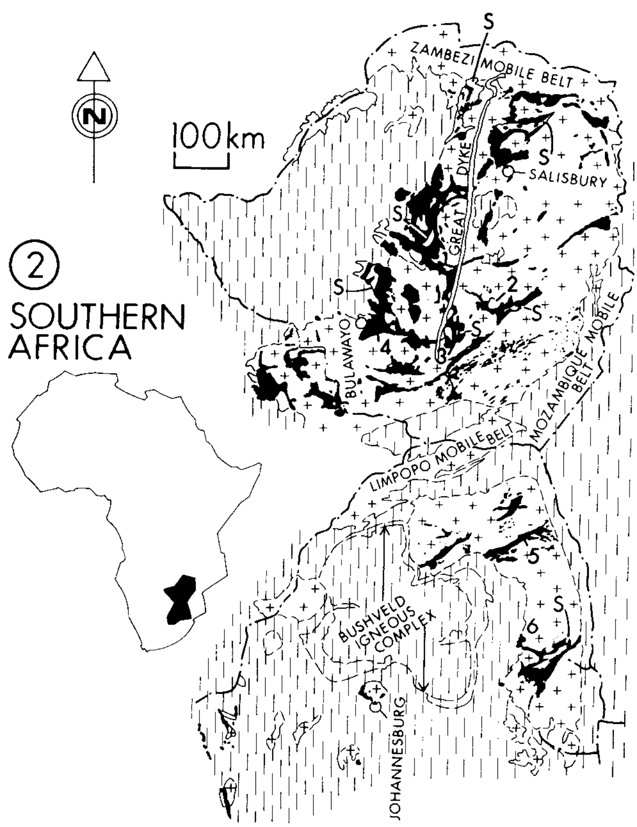


Fig. 5-1. Granite-greenstone terrains of (1) Peninsular India, (2) southern Africa (3) Yilgarn subprovince, Western Australia and (4) southern Superior Province, Canadian Shield. All are drawn to the same scale as are the inset maps for purposes of comparison. The greenstone belts are predominantly volcanic; some sediment-rich parts, especially in the Yilgarn subprovince, are shown separately. Data sources: (1) Peninsular India — fig. 1 in Naqvi et al., 1978; (2) Southern African — fig. 1 in Anhaeusser, 1976; (3) Yilgarn subprovince — Tectonic map of Australia and New Guinea, scale 1:5,000,000, Geological Society of Australia, Sydney; and (4) Geological map of Canada, 1968, map no. 1250A, scale 1:5000,000, Geological Survey of Canada (with modifications and additions).

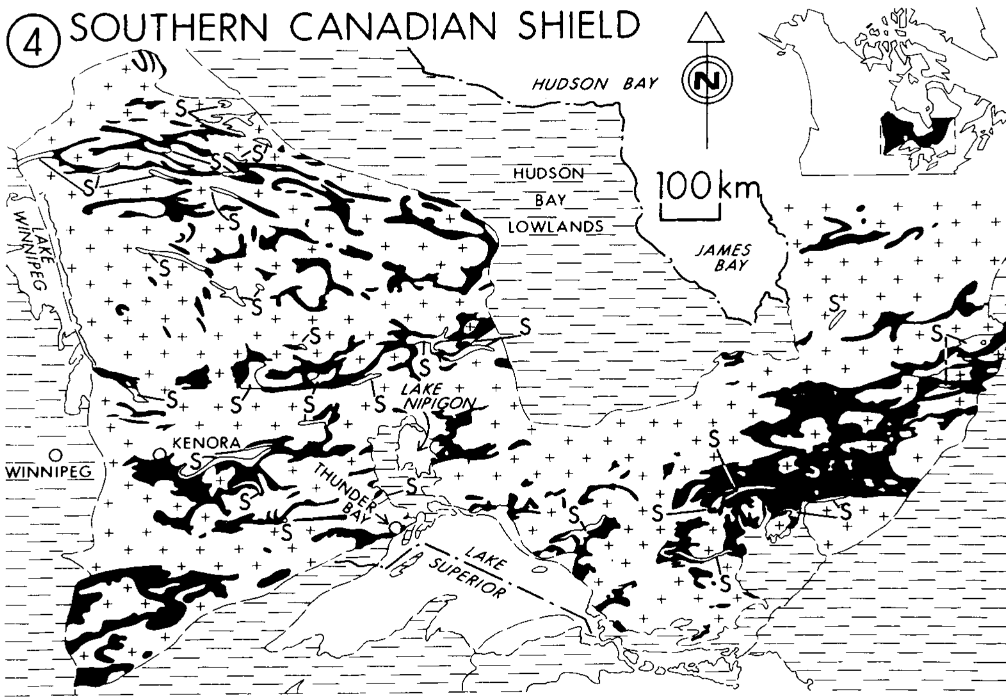


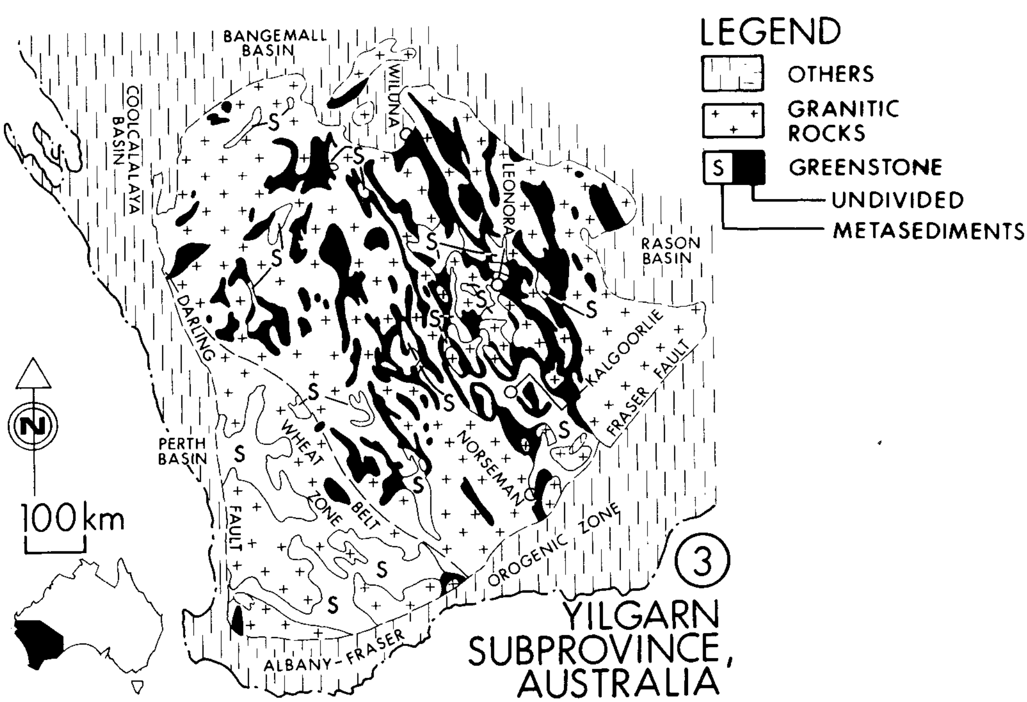
(1) Peninsular India:

1 = Shimoga,
2 = Kolar,
3 = Bababudan,
4 = Chitradurga, and
5 = Holenarsipur belts;

(2) Southern Africa:

1 = Selukwe-Gwelo;
2 = Fort Victoria-Mashaba;
3 = Belingwe;
4 = Bulawayo-Bubi;
5 = Murchison and
6 = Barberton belts.


4
SOUTHERN CANADIAN SHIELD
N
100 km
HUDSON BAY
HUDSON BAY LOWLANDS
JAMES BAY
LAKE WINNIPEG
LAKE NIPIGON
KENORA
WINNIPEG
THUNDER BAY
LAKE SUPERIOR
S


LEGEND
OTHERS
GRANITIC ROCKS
GREENSTONE
UNDIVIDED
METASEDIMENTS
BANGEMALL BASIN
COOLCALALAYA BASIN
WILUNA
LEONORA
RASON BASIN
DARLING
KALGOORLIE
FRASER FAULT
PERTH BASIN
WHEAT BELT
ZONE
NORSEMAN
FAULT
OROGENIC ZONE
ALBANY-FRASER
N
100 km
S
3
YILGARN SUBPROVINCE, AUSTRALIA

Kolar-type greenstone belts, e.g. Kolar, Sandur, Hutti, Nellore and Ramagiri belts, are predominantly ultramafic-mafic volcanic assemblages with minor sediments. They are characterized by an ultramafic-mafic assemblage at the base, stratigraphically overlain by fine-grained clastics and chemical sediments. The main components are metamorphosed peridotite, dunite, gabbro and anorthosite, talc-tremolite-actinolite schist, garnetiferous hornblendic schist, quartzite and amphibolite, kyanite-staurolite schist and magnetite quartzite (Naqvi et al., 1978).

The belts are comparatively small, narrow and pronouncedly linear; many are approximately 10—50 km long by 1—5 km wide, but some are 100 by 10 km in plan. The greenstone assemblage is extensively intruded by sodic tonalite-granodiorite-trondhjemite gneiss and migmatite. This is especially so at the margins of the belts. As a result no basement has been observed, so the nature of the original floor of the volcanic rocks cannot be established by direct observation (Naqvi et al., 1978).

Sargur schist belts comprise numerous small high-grade volcano sedimentary units scattered throughout the gneiss. Individual belts occur as parallel synclinal keels and antiformal units (Nath et al., 1976). Most are only a few tens of kilometres long and 1—5 km wide. They are characterized by high-grade mineral assemblages, intimate association of ultramafic rocks and ubiquitous migmatization. They display a consistent stratigraphic succession with basal quartzite-carbonate-aluminous sediments (shelf facies), overlain by and intercalated with mafic-ultramafic assemblages (serpentinized dunite, peridotite, pyroxenite, amphibolite and anorthosite) and banded ironstones (Viswanatha, 1977).

The high-grade Sargur belts are considered to have been deposited upon a sialic basement because of their high pelitic content. But no positive evidence of an unconformity has been observed (Nath et al., 1976). Instead the margins of the belts show gradational "mobilized" contacts marked by several "screens" of schistose rocks within the gneiss.

Geochronology

Pebbles of tonalite-trondhjemite composition from the Kaldurga conglomerate of the Bababudan Group (Dharwar Supergroup) and probably derived from the Dharwar basement have yielded a 3-point Rb-Sr whole-rock isochron age of 3160 ± 150 Ma with $I = 0.702 \pm 0.003$ (Venketasubramian and Narayanaswamy, 1974). Cordierite gneiss interpreted as supracrustal remnants in the Peninsular gneiss complex gave a reliable Rb-Sr whole-rock isochron age of 2950 ± 90 Ma with $I = 0.701$ (Jayasam et al., 1976).

A preliminary Sm-Nd age exceeding 3 Ga has been obtained for the Nellore greenstone belt (Kolar type) north of Mysore (A. Kröner, pers. commun., 1980). Dharwar-type greenstone belts, on the other hand, have been dated at 2600—2300 Ma (Viswanatha, 1977), thereby indicating a late Archaean—early Proterozoic age. The data thus indicate that Dharwar belts are considerably younger than Kolar-type belts which, in turn, may be younger than the Sargur schist belts (A. Kröner, pers. commun., 1980).

The post-volcanic granitic suite including the Closepet granite, ranges in age from 2.6—2.2 Ga (Crawford, 1969).

Tectonic development

Dharwar-type greenstone belts clearly unconformably overlie gneissic basement, at least locally. The basal sediments are shallow-water products and include conglomerates of granitic provenance. Middle and upper volcanic-bearing parts of the succession, however, appear to be products of deeper-water, downsinking geosynclinal environments, indicating substantial crustal instability. The mainly mafic and reputedly calc-alkalic Dharwar volcanic suite is compositionally similar to some other greenstone belts of the world (Naqvi et al., 1978).

Naqvi et al. (1976, 1978) interpret the data in terms of a thin early to middle Archaean simatic protocrust. Kolar-type greenstone belts are considered by them to have been developed in the cracks of this crust with tonalite-trondhjemites representing the products of partial melting of mafic crust at comparatively shallow depth. This process, it is contended, first formed the continental nuclei of mixed but predominantly mafic composition between which in late Archaean—early Proterozoic time "geosynclines" were formed in which Dharwar-type greenstone piles accumulated.

Thus a clear distinction is drawn between the earlier more primitive Kolar-type greenstone belts and the considerably younger more mature Dharwar belts related to continental geosynclines with both earlier stable and later dominantly unstable parts.

Southern Africa: Rhodesian and Kaapvaal cratons

Setting

The Rhodesian and Kaapvaal cratons. 312,000 km^2 and 585,000 km^2 in area, respectively, but substantially covered by younger formations, are separated by the east-trending Limpopo mobile belt (Fig. 5-1.2). The main exposed Archaean granite-greenstone terrain of the Rhodesian craton is 77 km by 300 km or 290,820 km^2 in area in which granite : greenstone $\simeq$ 4 : 1. That of the Kaapvaal craton is 79,560 km^2 in area in which granite : greenstone $\simeq$ 8 : 1. Both terrains display mainly synclinal greenstone belts occupying spaces between oval "gregarious" batholithic domes, a relationship particularly well exposed in the Rhodesian craton (Macgregor, 1951). The Zimbabwean granite-greenstone terrain is divided into east and west parts by the north-trending Great Dyke dated at 2460 ± 16 Ma (Hamilton, 1977), which provides the minimum age of the granite-greenstone terrain.

There are 24 main greenstone belts in the Rhodesian craton and 7 in the Kaapvaal craton (Anhaeusser, 1976). Zimbabwean belts range from small units 10 km long to the large Midland belt 200 km by 70 km in plan; other important Zimbabwean units are Fort Victoria—Mashaba, Belingwe, Bulawayo—Bubi and Selukwe—Gwelo belts (*see also Nisbet et al., this volume, Chapter 7, ed.*). Important Kaapvaal units are the Barterton and Murchison belts (Fig. 5-1). Low to medium metamorphic grades prevail.

Granitic rocks of the granite-greenstone terrains include a wide variety of biotite- and hornblende-bearing tonalitic granites and gneisses, passing locally into trondhjemitic, dioritic granodioritic and quartz-dioritic equivalents and including a variety of metamorphic xenoliths. They are associated with a batholithic suite of mean granodioritic composition together with younger medium- to coarse-grained, often porphyritic, potassic granite and syenite plutons.

Both cratons are bounded either by high-grade polymetamorphic mobile belts including the Zambezi belt to the north, the Mozambique belt to the east and the Namaqua belt to the south, all considered to include substantial reworked Archaean cratonic material (Anhaeusser, 1973), or by younger craton cover.

Lithostratigraphy

Rhodesian craton: The greenstone assemblages include Sebakwian, Bulawayan and Shamvaian groups, the first occurring in isolated belts and the latter two commonly combined in ascending stratigraphic order (*for details see Nisbet et al., this volume, Chapter 7, ed.*). The Sebakwian is typically exposed in small discontinuous belts and scattered inclusions associated with older gneissic basement. In the Fort Victoria region, relics of Sebakwian schist in the gneissic basement are demonstrably older than the Bulawayan Group in the main Victoria belt (Stowe, 1971) where the Bulawayan succession, about 12 km thick, contains a lower division of mafic pillow lava and banded ironstone, a middle division of andesite-dacite lavas, and an unconformably overlying andesite and agglomerate division. The unconformably overlying Shamvaian sediments comprise poorly sorted arenaceous sediments, mainly quartzite, conglomerate, greywacke and shale together with some andesitic-felsic lavas and pyroclastics.

Contact relations with adjoining rocks at the belt margins are typically intrusive and migmatitic. However, in the Shabani area of the Belingwe belt, an unconformity is exposed between sediments assigned to the Bulawayan Group and tonalitic basement gneiss (see Nisbet et al., this volume, Chapter 7). Bickle et al. (1975) stress that unequivocal evidence from the field indicates that the sedimentary-volcanic succession, including basalt and komatiite, was deposited directly on a granitic crust within a depositional environment that changed from probable beach, to tidal flat, to deeper water conditions.

Kaapvaal craton: The well-studied Barberton belt (Andaeusser et al., 1969; Anhaeusser, 1971, 1975, 1976, this volume, Chapter 6) provides the type succession for this craton. The Swaziland Supergroup contains three major subdivisions that are, in ascending order, the Onverwacht Group characterized by a relative abundance of ultramafic and mafic (komatiitic) volcanic rocks, the Fig Tree argillaceous Group and the Moodies Group, a cyclically repetitive assemblage of conglomerate, quartzite, subgreywacke, sandstone and shale. Lithologic details are provided elsewhere (see Anhaeusser, this volume, Chapter 6).

There is a broad lithologic similarity among Zimbabwean and Kaapvaal greenstone assemblages respectively of Lower Onverwacht to Sebakwian (ultramafic unit), Upper Onverwacht to Bulawayan (mafic to felsic volcanic unit), and Fig Tree and Moodies to Shamvaian assemblages (argillaceous and arenaceous sedimentary unit) without, however, implying a temporal correspondence (Anhaeusser, 1976, table 3).

The estimated class proportions (percent) in the respective greenstone belts are (Anhaeusser, 1976):

	Ultramafic-mafic	Basalt	Andesite	Felsics	Total
Rhodesian craton	9.8	60.2	22.5	7.5	100.0
Barberton belt	24.3	72.0	—	3.7	100.0

On this basis Kaapvaal belts are heavily weighted to the ultramafic to mafic andesite-poor side and many Zimbabwean belts to the andesite-felsic side of the compositional spectrum.

Geochronology

Rhodesian craton: Amongst the oldest reliable dates from the Rhodesian craton is an Rb-Sr isochron date on Shabani gneiss of 3495 ± 120 Ma (Moorbath et al., 1977). This gneiss and the Mushandike granite of similar age post-date older metasediments in the Fort Victoria greenstone belts, thereby suggesting that older than 3500 Ma greenstone belts may exist on the Rhodesian craton (Hawkesworth et al., 1975; Nisbet et al., this volume, Chapter 7). In addition, Mashabe area gneisses, which are typical of the Rhodesian Basement Complex, provided a date of 3504 ± 400 Ma (Hawkesworth et al., 1975).

Bulawayan Group mafic volcanics have been dated at between 2480 ± 280 Ma and 2670 ± 60 Ma, a similar age to that of 2520 ± 140 Ma obtained for the Selkirk volcanics in Botswana, (Hawkesworth et al., 1975; Jahn and Condie, 1976). Thus Bulawayan-type greenstone belts may have accumulated over considerable time, c. 2700—2500 Ma or c. 200 Ma. They clearly post-date at least some of the "basement" gneisses as well as the earliest Sebakwian-type greenstone generation for which a minimum age of 3350 ± 120 Ma is given by the Mont d'Or granite in the Selukwe belt (Moorbath et al., 1976).

Regarding petrogenesis, the low initial $^{87}Sr/^{86}Sr$ or I values of the Bulawayan volcanics, the cross-cutting Sesombi tonalite and the associated migmatites (Gwenoro Dam) which are all in close agreement around 0.701, suggest that the later granitic and andesitic rocks were not produced by remelting of, or contamination with, the ancient basement gneiss (Hawkesworth et al., 1975; see also Moorbath and Taylor, this volume, Chapter 20).

Kaapvaal craton: The majority of rock types in the Barberton Mountain Land exceed 3000 Ma in age *(for details see Anhaeusser, this volume, Chapter 6, ed.)*. The oldest reliable ages yet reported include a Rb-Sr isochron date of 3426 ± 200 Ma for basaltic komatiite from the Tjakastad subgroup of the Onverwacht Group (Anhaeusser, 1978) and a Sm-Nd age of 3540 ± 30 Ma (Hamilton et al., 1979). A Rb-Sr age of 3303 ± 40 Ma was obtained for the Middle Marker of the Onverwacht Group (Hurley et al., 1972). A zircon U-Pb age of 3289 ± 100 Ma (Van Niekerk and Burger, 1969) was obtained from Hooggenoeg lavas in the overlying Geluk Subgroup of the Onverwacht Group. Tonalitic diapirs which have deformed both the Fig Tree and Moodies sediments, were emplaced 3240 ± 40 Ma ago (Oosthuyzen, 1970).

The granitic rocks of the region provide ages ranging between 3320—3150 Ma for the early tonalitic and trondhjemitic gneisses (including those of the Ancient Gneiss Complex) and 2550 Ma for some of the late potassic plutons (Anhaeusser, 1978). Intermediate ages of 3090 Ma were provided by the Nelspruit migmatite and of 3063 ± 30 Ma by the Boesmanskop syenite (Oosthuyzen, 1970). Moodies conglomerate boulders have yielded tentative pre-metamorphic ages considerably older than 3300 Ma (Van Niekerk and Burger, 1978). An Rb-Sr whole-rock isochron age of 3555 Ma is reported from the Swaziland Bimodal Gneiss Suite (Barton et al., 1980).

Thus the actual dates for the lowermost parts of the Onverwacht volcanic-sedimentary Group about equal those obtained for the Ancient Gneiss Complex. Indeed, the two groups of ages are sufficiently close that no clear conclusion concerning their relative ages can yet be drawn. Confirmation of the > 3.3 Ga age of the Moodies pebbles, however, would support the presence of pre-Onverwacht sialic crust without necessarily implying Onverwacht sialic basement.

I ($^{87}Sr/^{86}Sr$) values obtained from basaltic komatiite of the Onverwacht Group are 0.70048 ± 0.00005 (Hawkesworth et al., 1975) and from the Boesmanskop syenite 0.70103 (Anhaeusser, 1978), thereby implying a mantle derivation for both the extrusive and the intrusive rocks.

Tectonic development

The Belingwe greenstone succession of the Rhodesian craton at Shabani provides unequivocal evidence that an Archaean metasedimentary-metavolcanic succession including basaltic and perioditic komatiites was deposited, at least in part, directly on granitic crust. The volcanicity of the Belingwe greenstone belt, according to Bickle et al. (1975), most closely resembles that of modern continental margins or island arcs where andesitic rocks form a significant proportion of the volcanic pile. Wilson (1973) earlier concluded that the depositories of the main greenstone belts evolved on a sialic basement on sites controlled by mantle-tapping fractures. Anhaeusser (*this volume, Chapter 6, ed.*), however, while accepting the evidence for a local sediment-basement unconformity, refutes that this evidence proves an older sialic basement to the older Kaapvaal greenstone belts.

Two striking conclusions are drawn by Hawkesworth et al. (1975) from their studies in Zimbabwe: (1) the development of the greenstone belts, Gwenoro migmatite and Sesombi tonalite as well as the Great Dyke emplacement all appear to have occurred within a period of c. 200 Ma or less; and (2) the later granitic intrusions (2600—2500 Ma) were not produced by melting of the ancient gneiss basement but rather from a more direct mantle source.

In the Barberton belt Anhaeusser (1975) maintains that incipient detrital sedimentation commenced only after "island arc-like" emergence of the volcanic pile. According to Anhaeusser, this sedimentation was supplemented by detritus derived from the erosion of granite, gneiss and migmatite that evolved progressively as a consequence of the foundering and partial melting of primitive, largely volcanically derived, ensimatic crust. The contrary view (Hunter, 1970) would derive the sialic detritus from older pre-volcanic sialic crust.

Thus the Zimbabwean greenstone assemblages provide some evidence of early crustal stability with beach and tidal flat environments at least in the Bulawayan belts, followed by deeper-water, presumably less stable environments with abundant island-arc-type volcanic accumulations. Kaapvaal greenstone belts, in contrast, provide little if any evidence of tectonic stability during the main greenstone accumulation.

Hawkesworth and O'Nions (1977) conclude that the distinctive trace-element geochemistry of the Archaean volcanic rocks precludes direct comparison with modern island-arc assocations. According to these authors, the essentially homogeneous source of the Archaean volcanic rocks and other associated features indicate that the greenstone belts developed in a rifting environment.

Western Australia: Yilgarn subprovince

Setting

The granite-greenstone terrain of the Yilgarn subprovince (terminology following Rutland, 1976), by far the largest Archaean craton in Australia, occupies the southern part of Western Australia (Fig. 5-1.3). It is crudely rectangular, measuring 950 km N—S by 700 km E—W or 614,000 km^2 in area in which granite : greenstone $\simeq$ 4 : 1. It is bounded to the north by Proterozoic orogenic zones including the Ophthalmian belt; to the east by Phanerozoic platform cover; to the south and southeast by a Proterozoic orogenic zone; and to the west by Phanerozoic cover with the intervening Darling Fault.

Yilgarn subprovince is subdivided into the comparatively small Wheat Belt zone to the southwest, characterized by high-grade metamorphic rocks including argillaceous, calcareous and arenaceous sediments; the small Murchison zone to the north (not illustrated); and the much larger Eastern Goldfields zone to the east, a broadly synclinorial region distinguished by the relative abundance of greenstone belts of dominant NNW trend and possible absence of granitic rocks older than 2800 Ma (Arriens, 1971). Metamorphic grade in the supracrustal rocks of the Eastern Goldfields zone is relatively low ranging from prehnite-pumpellyite to lower amphibolite facies within the belts to upper amphibolite at the belt margins (Binns et al., 1976).

Yilgarn subprovince contains 34 main greenstone belts. They range widely in size up to the large NNW-trending, substantially intruded, Wiluna—Leonora—Kalgoorlie—Norseman belt some 675 km long by 275 km wide.

The exposed areas between the greenstone belts are dominated by post-kinematic granitoids and associated granitic gneiss. A prominent rock type is grey tonalitic gneiss intruded by dykes and plutons of foliated leucogranite (Platt et al., 1978). More than one age of gneiss may be present.

Lithostratigraphy

The Eastern Goldfields zone has been subdivided by Williams (1974, 1975) into three north-trending subzones. The central subzone represents an unstable trough or graben-like environment where most of the ultramafic igneous rocks, nickel deposits and clastic sedimentary deposits occur. It is flanked on both sides by subzones representing a more stable tectonic environment characterized by banded iron-formations belonging to the lowest of three cycles (see below).

Mafic and ultramafic volcanic rocks with felsic volcanogenic and clastic associations of the Kalgoorlie Supergroup, the broad supracrustal assemblage of the Eastern Goldfields zone, are interpreted in terms of 3 major cycles of accumulation. Each cycle includes a mafic-ultramafic volcanic succession overlain conformably by felsic volcanic rocks and derived sediments. These cycles are separated from one another by pronounced unconformities and/or stratigraphically consistent chert layers. The total thickness is estimated to be c. 27 km (Williams, 1975) although Rutland (1976) suggests that strike faulting may require downward adjustments of such apparent thicknesses.

To the south, in the Kalgoorlie—Norseman district, the succession is dominated by a 9 km-thick monotonous sequence of mildly metamorphosed tholeiitic pillow basalts showing little vertical or lateral variation (Hallberg, 1972) but including rare felsic volcanic concentrations up to 1.2 km thick.

To the north, in the Agnew area 400 km north of Kalgoorlie, the greenstone succession consists of interlayered metabasalt, differentiated gabbroic sills, ultramafic bodies and black volcanogenic sediments unconformably overlain by granitoid-clast conglomerate and meta-arkose (Platt et al., 1978).

At Jones Creek a conglomerate rests with angular discordance upon lower volcanic rocks (Durney, 1972). The conglomerate is well-stratified and contains rounded granitic clasts and either an arkosic or mafic matrix. It has not yet been established whether the source of the granitic clasts is a possible basement or a younger granite intrusive into the Yilgarn succession.

Glikson (1971) documents the following upward trends in the Kalgoorlie succession:

(1) The ratio of sediments to igneous rocks increases. Low stratigraphic levels are characterized by pelitic metasediments which are transitional at intermediate levels to thicker greywacke-slate associations, themselves grading upwards to predominant greywacke-conglomerate facies.

(2) The variations in type and abundance of volcanic rocks are cyclic. Thus the Coolgardie ultramafic volcanic assemblage of the lower cycle represents a major magmatic event whereas a second mafic cycle with some intermediate-acid components in the Red Lake belt is comparatively minor.

Thus the increasing abundance and grain size of clastic sediments in the Kalgoorlie Supergroup with time was directly proportional both to the decreasing frequency and volume of the volcanic components and to the sequential transitions from ultramafic to mafic to intermediate to felsic compositions.

Geochronology

In summary the greenstone belts of the Eastern Goldfields zone provide radiometric ages of c. 2700 Ma, and associated granitic intrusions are dated at c. 2600 Ma. So far, no older rocks have been dated in this zone. However, high-grade gneiss older than 2900 Ma is reportedly present in the Wheat Belt zone to the southwest (Rutland, 1976).

Cooper et al. (1978) have established the following ages with accompanying $^{87}Sr/^{86}Sr$ initial ratios or I values (in brackets) to be present in the Agnew area: (1) the main volcanic accumulation occurred at or about 2718 ± 50 Ma ($I = 0.7007 \pm 0.0004$); (2) intrusion of voluminous tonalite-granodiorite magmas are represented by the Lawlers Tonalite dated at 2652 ± 20 Ma ($I = 0.70152 \pm 0.00012$) and Mt. Keith Granodiorite at 2632 ± 17 Ma; the Perseverance Granite at 2625 ± 34 Ma is probably an associated phase; and (3) subsequent minor granitic intrusions occurred at 2576 ± 14 Ma ($I = 0.70218 \pm 0.00021$, Lawlers Leucogranite) and 2474 ± 14 Ma ($I = 0.70193 \pm 0.00012$, leucotonalite-aplite intrusive event). About 90% of the material was emplaced 80 Ma after activity began (tonalite-granodiorite stage) and at least 99% had been emplaced after 140 Ma (leucogranite stage).

Tectonic development

It has been inferred that the Wheat Belt zone forms an older basement to the Eastern Goldfields greenstone belts (Rutland, 1976). However, the actual base to the greenstone belts is unknown since the adjoining granite is always intrusive. Rutland notes that the intrusive contacts are generally close to the lowermost level identified in the greenstone succession which suggests "remobilization" of older basement. In contrast, Glikson and Lambert (1973) argue that the Wheat Belt zone represents the coeval deep-crustal zone of the greenstone-bearing Eastern Goldfields zone. They argue, moreover, that the greenstone sequences represent primitive ultramafic oceanic crust, below which the granite rocks and high-grade metamorphic rocks were subsequently generated.

Regarding the lithostratigraphic record, Glikson (1971) interprets the upward coarsening sedimentary trend in the c. 20 km-thick Kalgoorlie succession from pelitic to greywacke-slate to greywacke-conglomerate facies as recording a transition, with time, from a pelagic environment to a basin differentiated into felsic volcanic centres and subsiding sedimentary troughs.

However, the lower pelitic sediments of the Kalgoorlie Supergroup have been more recently interpreted as products of a uniformly shallow-water environment over a wide area (Gemutz and Theron, 1975), deposited on a platform, shelf or shallow basin-like surface (Williams, 1975) and, specifically, in the Agnew area, in a flood plain or alluvial fan environment (Donaldson and Platt, 1975). Thus, although the interpretations at early water depth diverge sharply, there is agreement at least upon the trend from early stable to later predominantly unstable crustal environment.

Rutland (1976) concludes that the greenstone belts were deposited over a wide area in shallow water in analogy with the Bulawayan-type greenstone belts of Zimbabwe (Nisbet et al., this volume, Chapter 7) and that this appears to imply pre-existing "protocontinental" felsic crust despite the evidence of relatively low $^{87}Sr/^{86}Sr$ ratios in the post-volcanic granitic intrusions. Archibald et al. (1978) go so far as to suggest a comparison with continental plateau basalts or with Proterozoic basins such as the Hamersley as relevant at the early stage with later deepening of fault-controlled basins on a restricted scale leading to development of the coarser-grained sediments.

Based on uniformly low $^{87}Sr/^{86}Sr$ initial values and major- and trace-element studies in the Agnew region, Cooper et al. (1978) interpret the main volcanic accumulation to represent massive additions to the crust from the mantle. The subsequent tonalite-granodiorite and later leucotonalite were obtained by partial melting of tholeiitic material with garnet as the controlling mineral phase. The still younger minor leucogranites more likely resulted from remelting of tonalite-granodiorite under similar conditions. According to the authors, these magmatic events constitute a massive addition to the crust from the mantle, lasting about 250 Ma (2700—2450 Ma).

The above examples serve to illustrate the conflicting interpretations regarding basement—cover relations, tectonic environments and source of igneous components in the Yilgarn greenstone belts.

Two tentative conclusions pertinent to any Archaean greenstone belt model are drawn, then, from available evidence: (1) Kalgoorlie supracrustals witnessed a change from early stable (quiet) environment, whether shallow or deep, to substantial crustal instability involving notable basin development with active volcanic centres and subsiding sedimentary troughs; and (2) there was a massive and sudden addition of indicated juvenile igneous material from the mantle to the crust during and following the main greenstone accumulation. A similar petrogenetic conclusion has been drawn in other Archaean granite-greenstone terrains of the world (Moorbath, 1975).

Canadian Shield: southern Superior Province

Setting

The principal Archaean granite-greenstone terrain of the Canadian Shield in southern Superior Province is a large irregular east-trending unit some 1500 km long by 450 to 1000 km wide or 926,000 km^2 in area (Fig. 5-1.4) in which granite : greenstone $\simeq 4:1$. The terrain is bounded to the north by Phanerozoic cover rocks of Hudson Bay Lowland except for extensive Archaean gneiss of the Ungava lobe in the east; to the east by metamorphic rocks of the Grenville Province; to the south by Proterozoic and Phanerozoic cover rocks; to the west by Phanerozoic cover of the Interior Plain; and to the northwest by the cataclastic-intrusive boundary of the neighbouring Churchill Province (Goodwin et al., 1972).

The granite-greenstone terrain (Fig. 5-2) is subdivided into 5 main east-trending province-wide subprovinces or superbelts which are alternately greenstone-rich (Abitibi—Wawa, Wabigoon and Uchi Volcanic belts) and paragneiss-rich (Quetico and English River Gneiss belts). To the northwest, the Sachigo Volcanic belt characterized by smaller, irregular ESE-trending greenstone belts, lies north of the elliptical Berens Plutonic belt. The south-trending Kapuskasing Fault Zone of Proterozoic age transsects the eastern part of the granite-greenstone terrain (Goodwin, 1977a).

The granite-greenstone terrain contains 38 main greenstone belts ranging in size from small schist units up to the unusually large volcanic-rich Abitibi belt in the east some 650 by 225 km or 95,000 km^2 in area. The somewhat smaller, pluton-intruded Lake-of-the Woods—Wabigoon belt in western Wabigoon subprovince includes the type Keewatin volcanic succession (Lawson, 1885). The common cuspate shape, especially of the smaller greenstone belts, reflects the presence of numerous younger round to elliptical batholiths.

The greenstone successions feature mafic to felsic volcanic cycles of mixed tholeiitic and calc-alkalic compositions, commonly with overlying sedimentary rocks. The sequences are commonly synclinally deformed and typically in contact with younger intrusive granitic rocks. They feature low-metamorphic-grade interiors and medium-grade margins. Granitic rocks include a wide variety of older tonalitic, trondhjemitic and granodioritic gneiss and a younger more potassic batholithic suite. The calculated lithic proportion

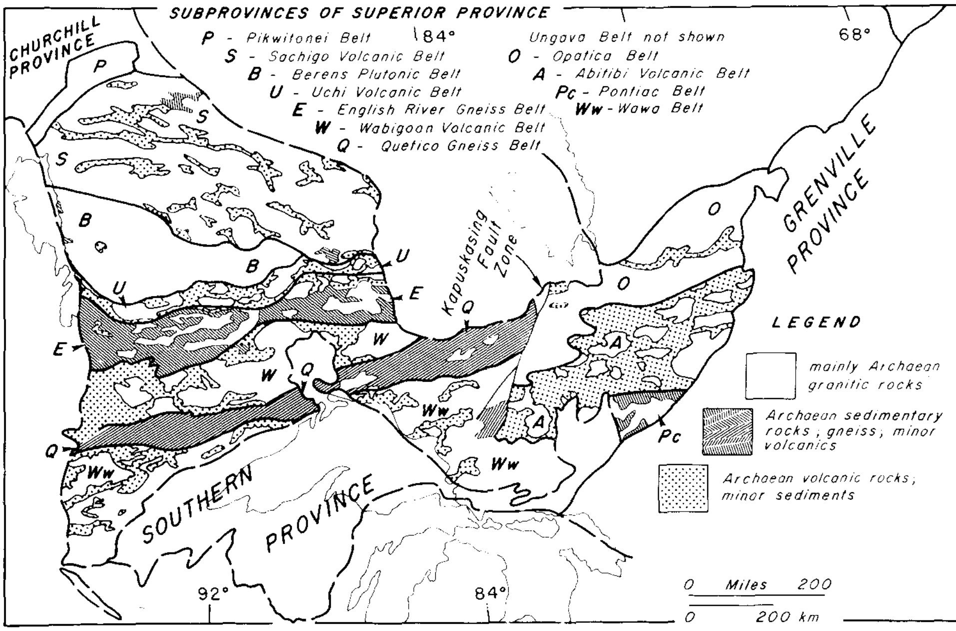


Fig. 5-2. Subdivisions of Superior Province, Canadian Shield showing the relative distribution of volcanic-rich and paragneiss-rich subprovinces (superbelts).

in the large western part of this granite-greenstone terrain is gneiss: massive plutons : metavolcanics : metasediments = 45 : 38 : 12 : 5; or granite: greenstone $\simeq$ 4 : 1 (Goodwin, 1978). The Abitibi greenstone belt itself, in the eastern part of this terrain, contains 49% metavolcanic rocks, 16% metasediments, 3% mafic intrusions and 32% granitic intrusions.

Lithostratigraphy

The greenstone belts of southern Superior Province are characterized by the presence of individual volcanic piles rather than uniform persistent stratigraphic successions (Ayres, 1977; Goodwin, in press). Individual greenstone belts contain a wide variety of stratiform and plutonic rocks. The principal volcanic components are lava flows and pyroclastics of the basalt-andesite-dacite-rhyolite association commonly arranged in mafic to felsic sequences. They are intercalated, especially in upper parts, with greywacke, argillite, tuff, conglomerate and chemical sediments, mainly banded iron-formations and chert. Associated internal granitoid plutons or diapirs are common and predominate in some belts. Mafic and ultramafic sills, dykes and small plutons are widely though sparsely distributed.

Tholeiitic basalt flows are particularly common in the lower stratigraphic parts of the volcanic piles and calc-alkalic andesite in the middle to upper parts. Thick local felsic volcanic accumulations toward the top of the sequence, mainly pyroclastic but locally massive to fragmental lava flows, represent central vent eruptions commonly occupying local subsidence structures. Mg-rich (komatiitic) volcanic rocks may occur at or near the base of each of two or more superimposed mafic to felsic volcanic cycles. Subaqueous volcanic accumulation is most common although subaerial accumulations have been noted in some piles (Ayres, 1977). Lateral facies changes are particularly common.

The volcanic piles vary greatly in stratigraphic thickness, commonly in the range 7—17 km. The accumulative stratigraphic thickness of the Blake River pile, part of the Abitibi Supergroup, for example, has been estimated in excess of 13.5 km (Baragar, 1968).

The larger greenstone belts contain similar proportions of the main volcanic classes. Thus basalt : andesite : dacite : rhyolite = 56 : 30 : 10 : 4 (Goodwin, 1977b). The calculated proportion of volcanic suites is tholeiite : calc-alkalic : alkalic = 57 : 38 : 5. The weighted mean chemical composition of the volcanic assemblages in different belts is remarkably uniform (Baragar and Goodwin, 1969). This suggests that most of the volcanic rocks are products of similar petrogenetic processes with the mantle the predominant if not exclusive source.

Precise stratigraphic correlations are difficult to make in the greenstone belts in all but the most favourable circumstances. The nature and style of supracrustal accumulation was not conducive to correlation. It featured a high degree of local diversity with volcanic piles of varying size and shape together with accompanying clastic wedges accumulating in local shelf and basin sites. Stratigraphic correlations are feasible, however, on a more local scale (Gelinas et al., 1977).

Most Archaean sediments (Pettijohn, in Goodwin et al., 1972) are greywacke with associated slates, some conglomerate, locally very thick, and banded iron-formation of the oxide, carbonate and sulfide facies. The most abundant facies is the greywacke-slate turbidite sequence with characteristic graded bedding. The conglomerates are less widespread but locally very thick, 1000 m or more, and in some cases traceable 50 km or more along strike. Walker and Pettijohn (1971) have shown some to be turbidite deposits, possibly subaqueous fans and perhaps even alluvial deposits.

In addition to these volcanic-associated sediments, large masses of schists and gneiss of sedimentary origin are present in the gneiss belts. Much of this paragneiss represents turbidite facies deposited in the medial portions of troughs. This implies a major sialic crust to supply such large quantities of normal epiclastic sediments (Pettijohn, in Goodwin et al., 1972). Thus there is evidence for major erosion of older sialic plutonic crust as well as of volcanic terrains during Archaean sedimentation.

Geochronology

The oldest dated rocks in Superior Province are tonalite-trondhjemite gneisses in the English River Gneiss belt at > 3040 Ma (Krogh et al., 1976). Development of these rocks may be related to similar events at > 2900 Ma in the Berens Plutonic belt to the north (Krogh et al., 1974; Ermanovics and Davison, 1976) and in the Duxbury massif, Eastmain district, Ungava, at 2915 ± 180 Ma (Verpaelst et al., 1980).

The main period of volcanism and sedimentation giving rise to the greenstone belts has been dated at 2760—2710 Ma (Turek and Peterman, 1971; Krogh and Davis, 1972). In more detail, an age difference of 22 Ma has been established from zircon U-Pb dating in felsic volcanic rocks from the lower (2725 ± 2 Ma) and upper (2703 ± 3 Ma) cycles, respectively, in the Timmins area, Abitibi Volcanic Belt (Nunes, 1980). However, an apparent age difference of 220 Ma was similarly established between felsic volcanic rocks of Cycle I (2958.6 ± 1.7 Ma) and Cycle III (2739 ± 2.5 Ma) in the Uchi—Confederation Lakes area, Uchi Volcanic Belt (Nunes and Thurston, 1980). Regional metamorphism with emplacement of large volumes of granitic rocks occurred in the eastern Lac Seul area, English River Gneiss belt at 2680 Ma and late or post-tectonic granite emplacement at 2660 Ma, followed by post-tectonic pegmatitic granites at 2560 Ma (Krogh et al., 1976). Thus the main phase of this magmatic cycle of events lasted for c. 100 Ma (2760—1660 Ma) and involved early supracrustal accumulation and later orogenic events including deformation, metamorphism and intrusion of large volumes of granitic rocks (Kenoran Orogeny); this was followed by some minor granitic intrusions for another 100 Ma.

Initial $^{87}Sr/^{86}Sr$ values as known are consistently low in rocks of Superior Province greenstone belts (0.7002 ± 0.00019—0.7010 ± 0.0004). This applies not only to the common basalt and andesite of the volcanic piles but also to the associated tonalite-trondhjemitic gneiss and massive to foliated granodiorite-granite batholiths (Goldich, 1972; Hanson, 1972, Peterman et al., 1972; Birk and McNutt, 1977; Hart and Brooks, 1977; Wooden and Goodwin, 1980). Prevailing low initial ratios, together with other isotopic data, suggest that most of the igneous material in the greenstone belts, both extrusive and intrusive, represents juvenile crustal additions rather than recycled substantially older crust.

Still older Archaean crust, 3500—3800 Ma in age has been identified in the Minnesota River valley to the south of Superior Province (Goldich and Hedge, 1974).

Tectonic development

Because of their linearity and volcanic make-up Superior Province greenstone belts have been interpreted by some in terms of the island arc model (Goodwin, 1977b; Ayres, 1978) with one preferred Archaean tectonic model involving interaction of sialic nuclei and oceanic-type crust. Sagging or buckling of thin lithosphere under prevailing high geothermal gradients may have promoted partial melting of mantle peridotite such that tholeiitic and calc-alkalic magmas were developed and extruded sequentially. The percentage of calc-alkalic volcanic components in Superior Province greenstone belts (38%) corresponds to that of Quaternary island arcs of thin to intermediate-type crust approximately 15 to 25 km thick (Miyashiro, 1974). The Archaean volcanic classes and assemblages are generally similar in abundance and composition to modern developed island arcs (Goodwin, 1977b).

Gelinas et al. (1977) invoke the island-arc model to explain Archaean volcanic relations in the Rouyn-Noranda district, Abitibi belt. According to these authors the Archaean volcanics were derived from depths of less than 50 km by high degrees (greater than 30%) of partial melting. The abundance of rhyolite in the higher stratigraphic levels of the Archaean pile suggests that the arc was of continental rather than oceanic-type — possibly resembling the Pleistocene volcanism of the Taupo province, New Zealand. Archaean volcanism, however, was more primitive at all stages of development than more mature modern continental arcs. This could be linked to thinner lithosphere and higher geothermal gradients in Archaean times.

However, others, including Baragar and McGlynn (1976), prefer a model of greenstone belt development involving a continuous comparatively thin sialic crust on which the volcanic piles accumulated (*see, for example, Henderson, this volume, Chapter 9, ed.*). Downbuckling due to the piling of the denser over lighter material results in the base of the crust sinking into the region of melting with consequent upward diapiric rise of granitoids into the growing volcanic-sedimentary piles.

There is general agreement, however, on the prevailing comparative instability of the crust during the volcanic accumulation, there being no substantial evidence in the sedimentary-volcanic record of the presence of stable platform environments.

Summary

Although all the greenstone belts lie in granite-greenstone terrains, contain mafic volcanic rocks and feature low- to medium-grade metamorphic assemblages they otherwise display great diversity in size, form, lithology, stratigraphy, age, basement-cover relations and structural deformation as briefly summarized below (Table 5-I). Such diversity points to equally diverse petrogenetic-tectonic histories.

(1) Individual greenstone belts range in size from small local schist units to the Abitibi belt (Superior Province) 94,720 km^2 in area (Fig. 5-1). Most represent steeply inclined, isoclinally folded, pluton-intruded supracrustal keels. However, Dharwar belts (Peninsular India) include substantial gently folded open structures locally overlying granitic basement.

(2) Greenstone belts are typically linear to curvilinear units of northerly to easterly (Superior Province) trends dispersed in predominant granitic rocks. In southern Superior Province the greenstone units are uniquely concentrated in long, province-wide volcanic-plutonic subprovinces (superbelts) alternating across strike with equally long paragneiss (turbidite)-rich subprovinces. If originally developed elsewhere on earth this superbelt pattern apparently has been destroyed or obscured by later processes including extensive cratonic fragmentation and dispersal.

(3) The four granite-greenstone terrains compare in size and greenstone

Table 5-I

Summary of greenstone belt relations

Type	Size, form and structure	Lithology	Stratigraphy	Greenstone age (Ga) with age of nearby sialic crust in brackets	Basement and tectonic relations
Peninsular India					
1. Dharwar	up to 400 km long, irregular, open structures	gwke, cgl, qte, BIF; bs and ry volcs.	basal shelf facies → turbidites-volcs	2.6—2.3 (> 3.1)	unconformably overlie basement gneiss; stable
2. Kolar	small, narrow, linear belts, 10—50 km long	km, prd, bs volcs; anorth; minor qte, pel, BIF	mainly volcs → metapelites and BIF	> 3.1	→ unstable crust undetermined basement; unstable crust
3. Sargur	small, narrow units 10—20 km long, synformal keels, etc.	km, prd, bs volcs; anorth; qte, carb, pel, BIF	basal shelf facies → volcs and BIF	> 3.1 (?)	sialic basement (?); early stable → less stable crust
Southern Africa					
4. Sebakwian	small, thin discontinuous belts, scattered inclusions	prd, km, volcs; BIF	ultramaf — maf volcs and minor seds	> 3.3 (3.5)	undetermined basement; unstable (?) crust
5. Bulawayan	small-medium size, cuspate to linear; folded units	bs, dc, ry, and + km volcs; qte, cgl, gwke, BIF	cyclic ultramaf-fels. volcs → seds; some early shallow seds	2.7—2.5 (3.5)	at least local gneissic basement; early stable → unstable crust
6. Barberton	small-medium size, elongate folded synformal units	km, bs, dc, ry volcs; sh, qte, gwke, cgl	cyclic ultramaf-fels; volcs → seds	3.4—3.2 (> 3.3)	undetermined basement; mainly unstable crust (final stability)
Yilgarn Subprovince, Australia					
7. Kalgoorlie	small to large (675 km long); synformal	km, bs, dc, ry volcs; gwke, BIF, cgl	cyclic ultramaf-maf fels volcs → sed; some early shallow seds	2.7 (> 2.9)	undetermined basement; early stable → unstable crust
Superior Province, Canada					
8. Keewatin	small to large (650 km long); synformal cuspate; superbelt pattern	bs, and, dc, ry volcs; gwke, BIF, cgl	cyclic maf-inter-fels volcs → seds	2.7 (> 3.1)	undetermined basement; unstable crust

Abbreviations: → = overlain (followed) by; seds = sediments; gwke = greywacks; cgl = conglomerate; qte = quartzite; BIF = banded iron formation; carb = carbonates; pel = pelites; sh = shale; volcs = volcanics; km = komatiite; prd = peridotite; bs = basalt; and = andesite; dc = dacite; ry = rhyolite; maf = mafic; inter = intermediate; fels = felsic; anorth = anorthosite.

content as follows:

Terrain	Total Area (km^2)	Greenstone belts Area (km^2)	%
Peninsular India (excl. NE part)	205,010	42,027	20.5
Rhodesia—Kaapvaal	369,980	66,950	19.0
Yilgarn Subprovince (excl. sediments)	615,250	118,210	19.1
Southern Superior Province	925,825	181,760	19.6

Thus greenstones form about 20% of preserved lithology in widespread Archaean granite-greenstone terrains, suggesting a common limiting degree of crustal uplift and supracrustal erosion during the cratonization process.

(4) Greenstone belts contain varied volcanic-sedimentary proportions. Most are predominantly volcanic with some entirely so; others are predominantly sedimentary (Dharwars); and still other volcanic-rich belts include extensive sediment-rich parts (Yilgarn). Sediments range from chemical (BIF, chert, carbonates) to clastic (greywacke, shale, orthoquartzite, conglomerate). Turbidites are most common. However, some belts (Dharwar, Bulawayan and Kalgoorlie) include shelf facilities at the stratigraphic base and/or top (Barberton).

Volcanic compositions are dominantly tholeiitic but vary from komatiite (high Mg)-rich and andesite-poor (Onverwacht, Kalgoorlie) to calc-alkalic rich basalt-andesite-dacite-rhyolite (multimodal) assemblages with or without minor komatiite (Superior). A mafic (ultramafic)-to-felsic cyclic succession is common. There is no apparent age and space restriction of komatiitic classes which occur in belts ranging in age from 3500 Ma to at least 2500 Ma and in stratigraphic position from preferred basal to upper horizons. However, komatiite-rich assemblages are far more common in terrains of the former Gondwanaland supercontinent. Andesite-rich assemblages, as presently known, are most common in the 2750—2550 Ma age range (Superior, Bulawayan). The calc-alkalic rich andesites typically overlie equally thick tholeiitic basalt sequences (Superior).

Thus komatiitic or high-magnesium volcanic rocks are widely distributed, though not necessarily abundantly, in Archaean greenstone belts but are far less commonly reported in younger rock assemblages. Their extrusion is considered to require high magma temperatures at surface, possibly around 1650°C (Green, 1975).

Tholeiitic volcanic rocks display chemical trends and parameters closely comparable to those of modern oceanic basalt-plagiogranite suites and, by analogy, may be attributed to crystal fractionation from a mantle-derived basaltic parent (Smith et al., 1977; Smith, 1980).

Calc-alkalic (high-Al_2O_3) volcanic rocks show REE patterns and LIL element abundances comparable to those of modern andesitic rocks in island arcs (e.g. Taylor and Hallberg, 1977). These data indicate that volcanic rocks comparable in some respects to those present in modern island arcs are also present among Archaean volcanic assemblages.

(6) Most greenstone assemblages exhibit evidence of widespread instability during supracrustal accumulation. Whereas basal and even uppermost parts of some successions (e.g. Dharwar, Kalgoorlie, Bulawayan) may point to shallow-water stable environments, the bulk is commonly attributed to rapid geosynclinal-type, basinal or fault-bounded trough accumulation featuring thick greywacke (turbidite)-volcanic assemblages.

(7) Greenstone belts considered herein range in age from c. 3500 Ma (Barberton) to 2600—2300 Ma (Dharwar), a span of 900—1200 Ma, with a notable cluster at c. 2700 Ma (Bulawayan, Kalgoorlie, Keewatin). Most of the granite-greenstone terrains contain relicts of still older supracrustal assemblages. When it is considered that still younger (c. 1800 Ma) greenstone belts of similar type occur in North America (e.g. Manitoba, Wisconsin, Arizona) and elsewhere, the accumulative time span of greenstone development must exceed 1.7 Ga.

(8) Most, but not all, greenstone belts have intrusive-migmatized margins; accordingly original basement relations have been destroyed. However, with one possible exception (Kaapvaal), pre-volcanic granitic rocks have been identified near greenstone belts in all terrains either by direct observation (basement-cover relations) or by radiometric dating. Thus greenstone belts typically developed in close proximity to older sialic crust. But the original extent and possible basement role of this older sialic crust during greenstone accumulation is largely unresolved.

(9) Low initial $^{87}Sr/^{86}Sr$ ratios are widely reported in Archaean volcanic rocks as well as in accompanying granitic intrusions (Moorbath, 1975). The data have been interpreted in terms of massive additions of juvenile material to the crust without substantial involvement of older sialic crust by way of melting, assimilation and mobilization.

(10) Finally it is noted that ophiolites, "blue schist" assemblages and paired metamorphic belts, all characteristic of modern convergent plate boundaries, are absent in Archaean terrains. These, together with rarity of kyanite and lack of eclogite, support a hotter, thinner lithosphere model lacking subduction (Benioff) zones during Archaean times. However, this does not preclude the presence of local, comparatively thick, stable middle—late Archaean crust (e.g. Kaapvaal craton).

ARCHAEAN PLATES

Consideration of the above Archaean relations and constraints leads to the following simplified rift-and-sag model to explain the varied Archaean greenstone belts:

(1) The prevailing mantle—crust during Archaean time, following Green (1975), featured a high surface heat flow; a thin lithosphere (c. 60 km) and correspondingly comparatively thin but irregular crust (averaging 20—30 km) over an asthenosphere with c. 5% melting from depths of c. 60 km to beyond

200 km; and a crust which included a thin, discontinuous, protocontinental sialic component (Fig. 5-3.I).

(2) Mantle diapirism from considerable depth (possibly c. 200 km), in response to deeper convective motions, involved rapid diapiric uprise and attenuation of the lithosphere including the sialic crust along lines of crustal extension or "hot lines" (Richter, 1973) with rapid decompression of the underlying mantle leading to a high degree of partial melting. The stage was then set for greenstone belt development.

(3) Plate thinning and incipient separation ensued. Tholeiitic and komatiitic magmas resulting from the high degree of melting in the mantle reached the surface by way of the resulting fractures. At sites limited to crustal attenuation and fracturing only (Fig. 5-3.IIa), the volcanic material accumulated ensialically together with sediments under early stable but later increasingly unstable tectonic conditions (Dharwar-type). Continued fissuring led to clear separation of sialic plates with massive ensimatic volcanic accumulation in the resulting fissures; volcanism led to construction of numerous large downsinking, in part superimposed, volcanic piles each dominantly tholeiitic-komatiitic below and, in some cases, increasingly calc-alkalic above (Keewatin-type)(Fig. 5-3.IIb). On a regional scale involving a number of parallel close-spaced simultaneous or sequential "hot-lines", each the site of crustal separation, a variety of local environments provided equally diverse volcanic-sedimentary assemblages. Three greenstone stages are envisaged: early attenuation promoted dominantly clastic accumulation under stable conditions upon sialic basement; further attenuation with fissuring provided increasingly unstable conditions leading to thick turbidite-volcanic geosynclinal accumulations; and continued fissuring leading to clear separation of the plates promoted ensimatic accumulation of numerous thick volcanic piles. Depending on the strength and longevity of individual "hot line" thermal convections in a region undergoing crustal extension, the three greenstone stages could develop in almost any age, stage and space relationship to provide the observed greenstone diversity (*see also Lambert, this volume, Chapter 18; Kröner, this volume, Chapter 3, ed.*). Accordingly a local basement-greenstone cover relationship may mark either a continuous sialic basement to the greenstone belt or merely a restricted edge effect. Also basement relations in one belt need not be the same in adjoining belts. In practise it is often difficult, if not impossible, to establish the original basement relations in a greenstone belt with intruded or migmatized contacts. Even recourse to indirect geochemical methods may not lead to unambiguous results.

(4) The proposed rift-and-sag process involved sagging of the ocean-type floor beneath the accumulating weight of the very large volcanic piles. This vertical sag in conjunction with ongoing horizontal motion consituted a sag-subduction (sagduction) process (Goodwin and Smith, 1980).

As previously outlined, the major geochemical discontinuity represented

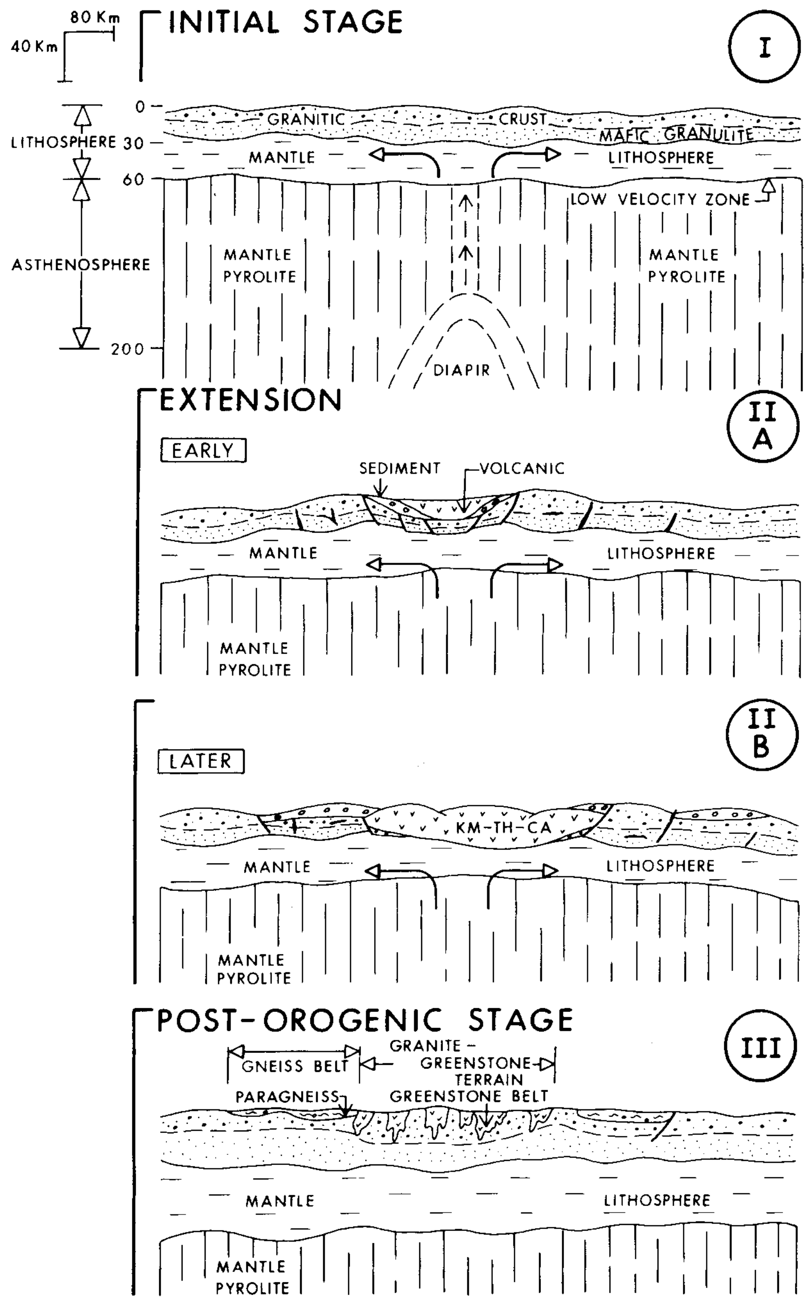


Fig. 5-3. Hypothetical relation of Archaean crust to greenstone belt development: I = initial stage; II = extensional stage, a = early and b = later; and III = final stage. A variety of tectonic, lithologic and age relationships are developed in Archaean greenstone belts.

by the early tholeiitic to later calc-alkalic transition which occurs at mid-thickness of many Superior Province and some Rhodesian volcanic piles may have been influenced by one or all of: (1) a lower degree of partial melting during the calc-alkaline stage in response to lower temperatures in the mantle source because of interaction with deeply sagged, relatively cool lithospheric material (e.g. Smith and Wooden, in prep.): (2) a direct response to the increased supracrustal loading (e.g., Hawkesworth and O'Nions, 1977); or (3) interaction between crustal and mantle materials to produce a chemically evolved source material which, on melting, yielded calc-alkalic magmas. However, the paucity of calc-alkalic components in the komatiite-rich Barberton and Yilgarn belts clearly demonstrates that thick volcanic accumulation by itself did not necessarily affect the mantle source in the direction of calc-alkalic development. Indeed, the differences in volcanic sequence between greenstone belts pose many unresolved petrogenetic problems.

(5) Thus the large volcanic-dominated greenstone belts are considered to represent mainly those uplifted and rapidly rifted lineaments that were ensimatically filled with volcanic materials including clastic and chemogenic sediments (BIF, chert, carbonates). The adjoining gneissic terrains (e.g. gneissic superbelts of Superior Province) represent the pre-existing fissured sialic crust with new epiclastic cover of both crystalline basement and volcanic provenance. This rifted and in part lowered and buried sialic crust with clastic cover underwent higher grade (amphibolite-local granulite) metamorphism and considerable deformation with development of autochthonous migmatites and synkinematic granitic intrusions (Sun and Hanson, 1975).

(6) Plate spreading and consequent greenstone belt development proceeded in response to the regional thermal convection system. Two factors theoretically limited this spreading: (1) waning of the thermal system; and (2) interference by adjoining parallel convection systems, the stronger overpowering the weaker. In either event individual plate separation ceased and, as a result of plate lowering, closing or both, plate compressions ensued. In the existing Archaean high geothermal regime, as proposed by Green (1975), eclogitization of mafic crustal material was not possible hence the subduction process involving return of mafic crust to the mantle along Benioff zones, was inoperative. Instead, the newly formed volcanic-derived crust remained "aloft" to be piled up between and against the sialic nuclei. Extensive isoclinal folding resulted, a deformational pattern accentuated by numerous rising granitic diapirs, products of partial melting of the lowered and substantially thickened crust. Crustal underthrusting, imbricate interleaving and nappe stacking may also have contributed to the thickening (Bridgwater et al., 1974).

(7) The diapiric rise of calc-alkaline granitic magmas from deep crust and/or mantle sources resulted in repeated intrusion and disruption of the over-

lying supracrustal-rich crust. The crust became increasingly stratified by igneous and metamorphic processes to produce upper silicic and lower mafic granulitic parts. Vestiges (greenstone belts) of the parent volcanic-rich assemblages in the upper crust mark the sites of the former "hot line" — induced fissures. Adjoining gneiss terrains feature: (1) pre-volcanic sialic crust; (2) paragneiss coeval with the volcanics; and, in common with the greenstone belts, (3) numerous syn- to post-kinematic granitic intrusions (Fig. 3-3.III).

(8) Greenstone belt development accordingly involved at least some lateral movement of sialic plates. The scale of individual place movement, presently undetermined, was presumably severely restricted by mutual interference of numerous small closely spaced plates, at least within those broad regions of Archaean protocontinental growth. However, the aggregate plate movement was probably very substantial with due allowance for the large number of plates. Archaean plate tectonics operated on a very different scale than at present. Yet the net effect was in the same direction — addition of large masses of juvenile mantle-derived material to the crust in response to successive orogenies. That crustal material was not returned to the mantle and thereby consumed implies large additions to the crust during Archaean greenstone development. The great diversity in Archaean greenstone belts clearly implies a corresponding flexibility in the tectonic process, a process lacking many of the characteristic ingredients of modern plate tectonics, yet resulting in the addition of many similar if not identical volcanic and plutonic components to the growing and consolidating Archaean sialic crust.

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Chapter 6

GEOTECTONIC EVOLUTION OF THE ARCHAEAN SUCCESSIONS IN THE BARBERTON MOUNTAIN LAND, SOUTH AFRICA

CARL R. ANHAEUSSER

ABSTRACT

The debate on the geotectonic evolution of Archaean granite-greenstone terranes is briefly reviewed to highlight some of the principal avenues of contention that surround this formative period in the earth's history. It is concluded that the majority of current opinion concedes that the Archaean record reflects broad-scale similarities with modern arc-trench systems but that the number of reservations expressed preclude direct comparisons with contemporary plate-tectonic mechanisms.

The controversy regarding the nature of the early crust is discussed and evidence from the Barberton Mountain Land relating to the problem is considered. It is concluded that the oldest recognizable rocks in this area are ensimatic in character (peridotitic and basaltic komatiites dated at 3.54 Ga) and are compositionally compatible with the probable source material from which large volumes of intrusive tonalites and trondhjemites were generated following partial melting at depth. These early sialic additions, some of which yield ages approximating that of the Barberton greenstone belt, both arose from and interacted with the supracrustal assemblages at deep and shallow crustal levels, respectively, to form complex migmatites and the protocontinental crust of the developing cratons.

A model of crustal evolution is outlined which draws heavily on the available evidence accumulated in the Barberton Mountain Land and adjoining territory of Swaziland. It is maintained that there is no evidence, either in the Barberton greenstone belt or in the surrounding granite-greenstone terrane, to suggest that plate tectonics, in the modern sense of the term, may have been responsible for the development of the region. Rather, it is claimed, the ancient crust evolved as a response to vertical tectonics involving the sinking of simatic lithospheric slabs and the diapiric upwelling of a succession of granitoids that commenced with early Na-rich phases but which later changed to K-rich magma types during the final stages of Archaean crustal consolidation 3.2—3.0 Ga ago.

INTRODUCTION

What might be termed the "Grand Debate" as to the nature and evolution of the Archaean crust of the earth continues to occupy the thoughts of a wide range of earth scientists. Advances in the knowledge and understanding of the low-grade granite-greenstone areas as well as the high-grade gneissic segments of the continents have progressed significantly since the early accounts of the Barberton Mountain Land in South Africa and the Godthaab District of West Greenland first appeared (compare Anhaeusser et al., 1969; McGregor, 1973; Viljoen and Viljoen, 1969a, b; with Windley, 1977; *see also Windley this volume, Chapter 1, ed.*).

The influence of plate tectonics has been profound. Initially, attention was devoted to plate tectonic theory as it affected post-Mesozoic geosynclines, island arcs, mountain ranges and oceanic domains. In recent years the success of these efforts led to more ambitious applications of the concept to Palaeozoic and late Precambrian orogenic terranes. A sequel to this was the tendency to apply the theory to the entire Precambrian record and to evaluate the possible role of the plate-tectonic mechanism during early continental evolution (Goodwin, 1973; Talbot, 1973; Condie, 1976).

Is there justification for considering Precambrian crustal evolution in terms of uniformitarian principles? This problem can be formulated in terms of three related questions:

(a) Does modern-type plate-tectonic modelling apply to the early Precambrian crustal history?
(b) Does some other form of "primitive" plate tectonics apply instead?
(c) Must we seek an entirely new model to explain early crustal evolution?

The principal support for some variant of primitive plate activity in the Archaean stems from the general similarity of a wide range of igneous and volcanic rock types which were initially regarded as little different in major- and trace-element chemistry from modern island-arc assemblages (Folinsbee et al., 1968; Condie, 1976; Engel et al., 1974; Taylor, 1977). However, to some, the island-arc analogue remains only partly acceptable as an explanation for Archaean greenstone belt development when examined in detail. Conspicuously absent in Western Australia, northeastern Minnesota and South Africa (but present locally in some of the younger greenstone sequences in Canada and Zimbabwe (Rhodesia) are andesites which abound in island arcs and active continental margins (Viljoen and Viljoen, 1969b; Hallberg, 1972; Arth and Hanson, 1975). Absent also are blueschist facies rocks and the "paired metamorphism" so characteristic of the circum-Pacific island arcs (Miyashiro, 1961; Engel, 1970).

The general concept of continental growth through the addition of material derived by a multi-stage process involving convection in the asthenosphere has been appealing to a number of investigators. Although the processes envisaged may be likened in a broad sense with sea-floor spreading, plate tectonics and island-arc development, the proponents of this view require that some modifications of the modern plate-tectonic concept are necessary to account for the evolution of the Archaean lithosphere (Engel and Kelm, 1972; Burke et al., 1976; Glikson, 1976a; Tarney et al., 1976). Caution was also expressed by O'Nions and Pankhurst (1978) who maintained that the compositions of volcanic and plutonic rocks formed in recent times are difficult to relate unambiguously to a specific geological environment — an exercise they considered to be even more hazardous in the Archaean.

A further group of specialists holds the opinion that no plate-tectonic model is applicable to the Archaean (Burke and Dewey, 1973; Dewey and Spall, 1975; Glikson and Lambert, 1976; Van Bemmelen, 1976; Glikson, 1979) and a wide range of alternatives has been suggested. These include catastrophic events that may have triggered widespread or local melting and include: (1) rapid formation of the earth's core after or during accretion; (2) surface impact of extra-terrestrial matter; and (3) lunar capture (Green, 1972, 1975; Condie, 1976; Glikson, 1976b). Further considerations envisage that, in order to account for the production of mafic and ultramafic magmas which require a steep geothermal gradient (Green, 1975) and granitic magmas which appear to require more moderate geothermal gradients, melting processes were initiated by mantle plumes (Hunter, 1974; Condie, 1975; Archibald et al., 1978).

It is clear that Archaean heat flow was high since radiogenic heat production was so much greater at that time (Birch, 1965). In an environment where thermal gradients were probably much steeper than today convective processes must have been different in scale and intensity to the motions which we associate with modern plate-tectonic phenomena. These factors led to models of smaller convective cells (with approximately 100 km wavelength patterns -- Bridgwater and Fyfe, 1974) and a surface of the globe influenced by tempestuous "hot spot" and tectonic activity (Fyfe, 1978; Hargraves, 1976; *see also Hargraves this volume, Chapter 2; Lambert, this volume, Chapter 18, ed.*).

In reviewing the literature on all the alternative views on crustal evolution one point seems to stand out above all others. A majority of opinion appears to favour non-uniformitarian crustal evolution and most would probably ascribe this to a secular reduction in the thermal energy available to drive asthenospheric convection.

With this background in mind it is clear that an analysis of the events in the Barberton Mountain Land alone will not provide a unique solution to the problems of earth history. The latest age of the volcanic sequence in the Barberton greenstone belt has been put at 3540 ± 0.03 Ma (Hamilton et al., 1979), this resulting in over 1 billion years of unaccounted for time from the earth's beginnings. This formative episode in the history of the region must of necessity be speculative in the extreme and will undoubtedly reflect the prejudices of the writer.

THE PRIMORDIAL EARTH'S CRUST

The "Grand Debate" embraces the controversy regarding the nature of early crust of the earth. Two issues are involved. The first relates to the period of earth history older than the oldest preserved crust yet recognized in the siliceous high-grade metamorphic terranes of West Greenland, Minnesota, the Indian Shield and the Limpopo belt in southern Africa, where ages of up to 3.8 Ga have been reported. The second relates to the nature of the crust on which Archaean greenstones were deposited. To some, separation of these questions might appear superfluous. However, as pointed out by Hargraves (1976), there is no assurance that the oldest terranes are displaying the pre-3.8 Ga primordial crust and there is speculation, based on findings that the oldest lunar crust appears to be composed dominantly of anorthosite, that such a beginning cannot be excluded for the original crust of the earth (Windley, 1970). Others have argued that even in the Greenland (and other) Archaean high-grade metamorphic terranes there are large deformed xenoliths of amphibolite preserved in the siliceous gneisses which could reflect remnants of a primary simatic crust or an older greenstone belt. Examples include the Isua enclave intruded by 3.6—3.76 Ga Amîtsoq gneisses in West Greenland and numerous essentially "basaltic" rafts found both in the Amîtsoq gneisses and the Uivak gneisses of Labrador (Bridgwater et al., 1978).

In general the controversy as to the nature of the ancient crust revolves around supporters of a primary simatic crust, possibly analogous to modern oceanic or island-arc-trench domains, and another group that favours the concept of an early sialic crust. Into the first category are included those who consider that an early basaltic (tholeiitic) crust once existed but has probably not been preserved anywhere (Barker and Peterman, 1974; O'Nions and Pankhurst, 1978; *see also Hargraves, this volume, Chapter 2, ed.*). Minor amounts of acid differentiates may have formed at this stage but are not a significant component of the present continental crust. Others favouring the simatic crust concept have been more concerned with the relationship of greenstone belts to an early crust and claim that remnants of such material may be viewed in the basal stratigraphic sequences of the generally isoclinally deformed, Archaean greenstone belts (Engel, 1970; Glikson, 1972, 1976b; Arth and Hanson, 1975; Glikson and Lambert, 1976, Anhaeusser, 1978; Naqvi et al., 1978).

As greenstone belt sequences have different ages — for example, 3.76 Ga for Isua (Moorbath et al., 1975), 3.54 Ga for the Barberton occurrence (Hamilton et al., 1979), and 2.7 Ga for the greenstones of northeastern Minnesota and adjacent Ontario (Hanson et al., 1971) and also different greenstone successions of the same terrane may have distinct ages, for example, the pre-3.5 Ga Sebakwian and 2.7 Ga Bulawayan in Zimbabwe (Rhodesia) (Hickman, 1974; Moorbath et al., 1977) — it is evident that the interpretation placed on their geotectonic setting will require a number of (continuous ?) stages of oceanic crust formation to have taken place throughout the Archaean.

Proponents of an early sialic crust point to the older ages of some siliceous Archaean high-grade metamorphic terranes (Goldich and Hedge, 1974; Moorbath et al., 1975; Barton et al., 1977), occasional preservation of unconformities where greenstone belt sequences rest on older granitic terranes (Bickle et al., 1975; Baragar and McGlynn, 1976; *see also*

Henderson, this volume, Chapter 9, ed.), and the presence of granite detritus in the sedimentary parts of greenstone successions (Condie et al., 1970; Baragar and McGlynn, 1976). Others favouring a primitive sialic crust on or adjacent to which greenstone sequences are considered to have been deposited include Archibald et al. (1978), Hargraves (1976), Hunter (1974), Wilson (1979) and Windley (1977).

THE ARCHAEAN CRUST OF SOUTHERN AFRICA

The southern African Shield is comprised of low-grade granite-greenstone cratons, the latter bordered by high-grade metamorphic mobile belts (Anhaeusser et al., 1969). Recent age determinations indicate that the oldest rocks found so far occur in the Limpopo belt where tholeiitic dykes and granodioritic gneisses have yielded ages of 3.64 Ga and 3.8 Ga respectively (Barton et al., 1977; *see also Barton and Key, this volume, Chapter 8, ed.*). These ages are approximately 200—300 Ma older than the ages recorded for gneisses and greenstones on the neighbouring Rhodesian and Kaapvaal cratons (Hickman, 1974; Moorbath et al., 1977; Hamilton et al., 1979) and clearly demonstrate the existence of sialic material prior to the development of the oldest greenstone belts yet dated in southern Africa.

However, the presence of old siliceous gneisses in the Limpopo belt provides no assurance that similar rocks need have been present in the areas now occupied by the adjacent granite-greenstone cratonic terranes. Claims to this effect have been made by Hunter (1970, 1974) and Stowe (1968), both of whom visualized areas of gneiss and migmatite in Swaziland and Zimbabwe (Rhodesia) being representatives of ancient pre-greenstone belt sialic crust. However, despite detailed geochronological investigations, the temporal relationships of these ancient gneiss terranes do not exceed ages reported for neighbouring greenstone belts (Hawkesworth et al., 1975; Davies and Allsopp, 1976). An exception to this occurs in the region east of the 2.70 Ga old Belingwe greenstone belt in Zimbabwe (Rhodesia) (Bickle et al., 1975) where gneisses exceeding 3.5 Ga in age have been reported by Hickman (1974) and Moorbath et al. (1977). However, these gneisses contain numerous Sebakwian mafic and ultramafic xenoliths of an earlier greenstone generation which might suggest that no initial sialic crust existed in this area.

THE BARBERTON MOUNTAIN LAND

The essential geological features of the Barberton greenstone belt and surrounding granitic terrane have been documented and reviewed in numerous publications (for reference details see Anhaeusser, 1978; Anhaeusser and Robb, 1978; Robb, 1978).

In its simplest form the Barberton greenstone belt comprises a synclinorial succession of komatiitic ultramafic and mafic lavas, tholeiitic basalts, intermediate to felsic lavas and pyroclasts, cherts, banded iron-formations and clastic argillaceous and arenaceous sediments. The entire greenstone belt is enveloped by intrusive granitoid rocks consisting of tonalitic and trondhjemitic gneisses and a variety of younger potassic bodies including granodiorites, adamellites, granites and syenites. Numerous greenstone remnants occur in the surrounding granitic terrane and complex migmatites are developed at granite—greenstone interfaces.

The chemical character of the volcanics displays strong affinities with modern oceanic or island-arc environments and the metamorphism is of low grade. The sediments are either volcanogenic, chemical, or clastic, varying according to their position in the stratigraphic record. Deformation is ubiquitous throughout the region but is most intense near the granite-greenstone margins.

The recognition of primary volcanic, plutonic, sedimentary and geochemical features in the Barberton Mountain Land resulted in the collation of data that placed constraints on models of early crustal evolution. The presence of a basal ultramafic—mafic extrusive assemblage of pillowed and massive komatiitic basalts and peridotites (Viljoen and Viljoen, 1969a) demonstrated that fundamental distinctions exist between Archaean and younger volcanic activity. Furthermore, the volcanic and geochemical changes that occur in ascending stratigraphic order in the greenstone pile (including cyclical calc-alkaline associations of basalt-dacite-rhyodacite-rhyolite) demonstrate further distinctions between Archaean and younger volcanic activity.

Geochemical studies from around the world confirm that Archaean greenstone belts have close similarities with modern island arc and Cordilleran chains (Folinsbee et al., 1968; Goodwin, 1968; White et al., 1971). Closer inspection reveals, however, that important geochemical differences exist between Archaean and younger volcanic-plutonic suites (Glikson, 1971; Hallberg, 1972; Jahn et al., 1974; Arth and Hanson, 1975; O'Nions and Pankhurst, 1978; Sun and Nesbitt, 1978). The absence of andesites in most greenstone belts (excluding some of the younger greenstone occurrences in Canada and Zimbabwe (Rhodesia) is particularly noteworthy. In addition, the Archaean volcanics comprise mainly low-K oceanic-type tholeiites that, in distinction from modern mid-ocean ridge basalts, possess flat REE patterns. Acid volcanics and pyroclastics are common and show highly fractionated REE patterns. By contrast, komatiitic basalts are rare and komatiitic peridotites appear to be absent in modern island arcs and the tholeiitic and calc-alkaline volcanic suites contain abundant andesites and show Al-rich affinities. Furthermore, alkaline (shoshonitic) volcanics are not uncommon in these younger sequences and both the mafic and the acid volcanics display fractionated REE patterns (Condie, 1976).

The volcanic record in the Barberton greenstone belt provides both similarities as well as significant differences between the Archaean and Mesozoic/Cenozoic island-arc volcanic suites. Interpretations of the contrasting characteristics noted above will doubtlessly vary but, in the absence of evidence to the contrary, the writer favours the view that the greenstone assemblage developed in an ensimatic domain. Had a sialic substratum underlain the basal komatiitic volcanic sequence in the Barberton greenstone belt, the intrusive diapiric plutons would be expected to be highly fractionated products of ensialic melting (adamellite, granite) rather than tonalites and trondhjemites diagnostic of ensimatic melting. Whether or not the broad similarities permit the application of a plate-tectonic-type mechanism for the genesis of the magmas remains debatable. It could be argued that the chemical differences recorded between the ancient and the modern examples merely reflect secular changes accompanying a reduction in the earth's heat-flow pattern as well as an evolving asthenosphere, and that primitive sea-floor spreading and subduction may well have occurred in the Barberton area.

For the analogy to be further substantiated a proto-continental mass would have to be sought in the regions flanking the greenstone belts. In the Barberton area proponents of an early ensialic crust would doubtlessly have no hesitation in supporting the suggestions that the Ancient Gneiss Complex of Swaziland (Hunter, 1970, 1974) acted as a protocontinental mass and that the Barberton volcanic pile is the adjacent primitive island arc.

Considerable disagreement has, however, been expressed as to the validity of the concept of the Ancient Gneiss Complex representing a pre-greenstone belt crustal entity (Viljoen and Viljoen, 1969c; Jenner and Gorman, 1977; Anhaeusser, 1978; Glikson, 1979; Robb and Anhaeusser, 1979). The Ancient Gneiss Complex, as originally defined, comprises a suite of extensively granitized metamorphites, the latter resembling the supracrustal rocks in the Barberton greenstone belt (Hunter, 1970). On the basis of metamorphic grade and structural data it was claimed, however, that the Ancient Gneiss Complex pre-dated the Swaziland successions. In a subsequent account Hunter (1974) described the Ancient Gneiss Complex as a suite of gneisses consisting of two main types, namely; (a) a bimodal association of interlayered leuco-tonalitic gneisses and amphibolites; and (b) metamorphites that include quartzites, quartzofeldspathic gneisses, siliceous and biotite-rich garnetiferous gneisses, quartz-diopside and diopside-plagioclase granulites (textural term), quartz-magnetite-grunerite gneisses, iron-formations, and biotite-hornblende gneisses. These rock types led Hunter et al. (1978) to subdivide the Ancient Gneiss Complex into three major units comprising: (1) a bimodal suite of closely interlayered siliceous, low-K gneisses and metabasalt; (2) homogeneous tonalite gneiss; and (3) interlayered siliceous microcline gneiss, metabasalt, and minor metasedimentary rocks — termed the metamorphite suite.

Davies and Allsopp (1976) presented Rb-Sr isochron ages and corresponding initial $^{87}Sr/^{86}Sr$ ratios of gneisses and hornblende-biotite tonalite samples from the Ancient Gneiss Complex. Ages range from 3323 ± 86 Ma (Ro = 0.7006 ± 0.0012) for the tonalite gneisses to 3270 and 3072 Ma (Ro = 0.7022 ± 0.0013 and 0.7060 ± 0.0002, respectively) for the banded gneisses. From their results it was suggested that a genetic relationship exists between the various gneisses of the Ancient Gneiss Complex. Hunter et al. (1978) did not share this interpretation and claimed that, because of several thermal and structural events, the resulting ages of the various gneisses could be regarded as unreliable. In their view the bimodal suite is possibly of 3.5—3.8 Ga age and predates the Swaziland supracrustals by several hundred million years. This, however, has yet to be verified but is considered unlikely as the Ro values place constraints on the crustal pre-history of the gneisses and the tonalitic chemistry of both the bimodal suite and the hornblende-biotite tonalites of the Ancient Gneiss Complex militate against an origin by ensialic anatexis.

Geochemical arguments showing that the siliceous gneisses of the bimodal suite possess low K/Rb and intermediate-to-high Rb/Sr ratios relative to the post-Barberton greenstone belt diapirs were used by Hunter (1974), Condie and Hunter (1976) and Hunter et al. (1978) to negate the views expressed by Viljoen and Viljoen (1969c) and Anhaeusser (1973) that there were grounds for correlating the area to the southwest of the Barberton greenstone belt with the Ancient Gneiss Complex and the Granodiorite Suite in Swaziland. These arguments were considered unconvincing (Glikson, 1976c; Jenner and Gorman, 1977; Robb and Anhaeusser, 1979) as the data made available did not unequivocally substantiate the interpretations placed upon them. Little distinction could, for example, be drawn between the quartz diorites, tonalites and trondhjemites of the Granodiorite Suite and the ancient tonalitic diapirs as both these suites share very low Rb/Sr ratios, high Sr/Ba ratios and positive Eu anomalies. Likewise, as shown by data from trondhjemite-granite batholiths in Western Australia (Glikson, 1979), low K/Rb ratios do not necessarily provide a genetic distinction between K- and Rb-high gneisses and LIL element-poor high-level tonalites. As the diapiric tonalites and the Ancient Gneiss Complex almost certainly represent different crustal levels, it would be unrealistic to expect direct geochemical correlations between shallow and deep batholithic levels, as they are subject to different thermal and differentiation histories. Additional geochemical information recently available (Anhaeusser and Robb, 1980a; Anhaeusser and Robb, unpubl. data) strengthens the view that it would be feasible to promote a correlation between the respective areas mentioned above rather than to polarize them. Support for the correlation of the two areas can also be found following the examination of numerous greenstone xenoliths extending for distances exceeding 60 km in the granitic terrane southwest of the Barberton greenstone belt. The xenoliths which display a wide range of lithological assemblages, including mafic and ultramafic metavolcanic rocks with subordinate felsic interlayers and minor siliceous metasediments, the latter of chemical sedimentary origin (banded chert, banded iron-formation, calc-silicate rocks), can be traced directly (Fig. 6-1) into the lowermost stratigraphic units of the Onverwacht Group of the Swaziland succession (the Sandspruit and Theespruit formations as defined by Viljoen and Viljoen, 1969a).

The rocks, in turn, constitute the oldest recognizable supracrustal sequence of the Kaapvaal craton having yielded a precise Sm-Nd age of 3540 ± 30 Ma (Hamilton et al., 1979). The supracrustal greenstone xenoliths are, furthermore, largely identical to the metamorphites described in the Ancient Gneiss Complex (personal observations), differing only (if at all) in the degree of metamorphism.

Detailed mapping southwest of the Barberton greenstone belt has revealed the existence of at least two generations of tonalite or trondhjemite (Anhaeusser and Robb, 1980a) and the migmatites exposed throughout the region can now be convincingly shown to conform to a predictable pattern, being consistently located in close proximity or adjacent to greenstone remnants (Fig. 6-1). The migmatite textures appear to have originated by progressive stages of in situ granite—greenstone interaction (Figs. 6-2 and 6-3), in a manner seemingly identical to the changes reported by Hunter (1970) in the Swaziland gneiss terrane.

The impasse that exists as to the nature of the earliest crust in the region will probably continue as the emphasis switches from one line of interpretation to another. The writer prefers a view of the variations in chemistry, metamorphic grade and structural style of the low- and high-grade terranes in terms of differences in degrees of partial melting and vertical zonation in the Archaean crust (see also Glikson and Lambert, 1976; Glikson, 1979). These factors probably contribute to the difficulties experienced in resolving the isotopic histories of the greenstones and granitoids as it is not always easy to distinguish discordant age patterns produced by cooling (crystallization stages) and discordant ages produced by overprinting (where igneous and/or metamorphic rocks were reheated but did not undergo complete recrystallization). The complex structural relationships observed within and between the various lithologies in the gneiss terranes enveloping the Barberton greenstone belt (Jackson, 1979; Anhaeusser, 1980; Anhaeusser and Robb, 1980a, b) do not necessarily reflect separate deformational episodes. Instead they may represent a continuum of strain increments (in rocks of widely differing ductility) caused by the emplacement of a succession of buoyant granitoid diapirs. These diapirs now comprise foliated to gneissic tonalite/trondhjemite and may be the product of the mechanical remobilization of early, possibly tabular granitoid sheets which originally intruded and underplated the oldest supracrustal rocks presently exposed — ideas similar to those that have been expressed by Schwerdtner et al. (1979) and Schwerdtner and Lumbers (1980) to account for features observed in the Superior Province in Canada.

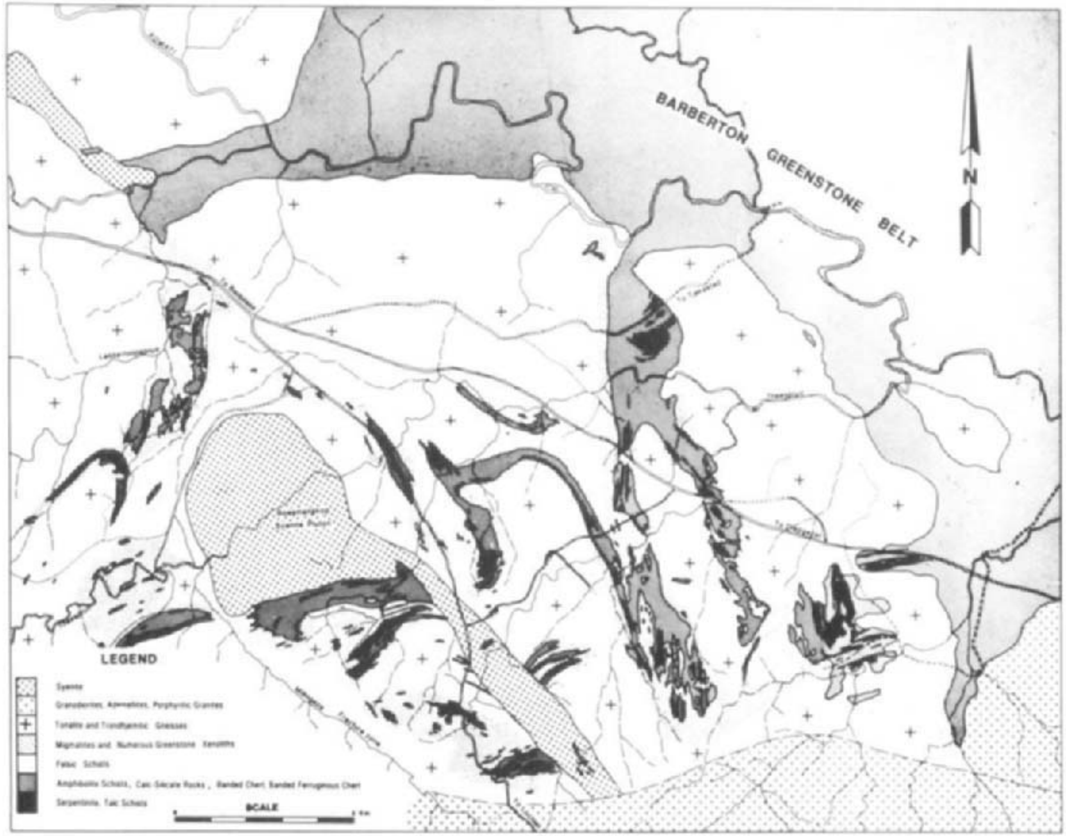


Fig. 6-1. Geological map of the granite-greenstone terrane south of the Barberton greenstone belt, modified after Anhaeusser and Robb (1980a). The greenstone remnants, which can be traced directly into the Sandspruit and Theespruit formations of the Onverwacht Group, are intruded by tonalitic or trondhjemitic gneisses resulting in the local development of complex migmatites like those shown in Figs. 6-2 and 6-3.

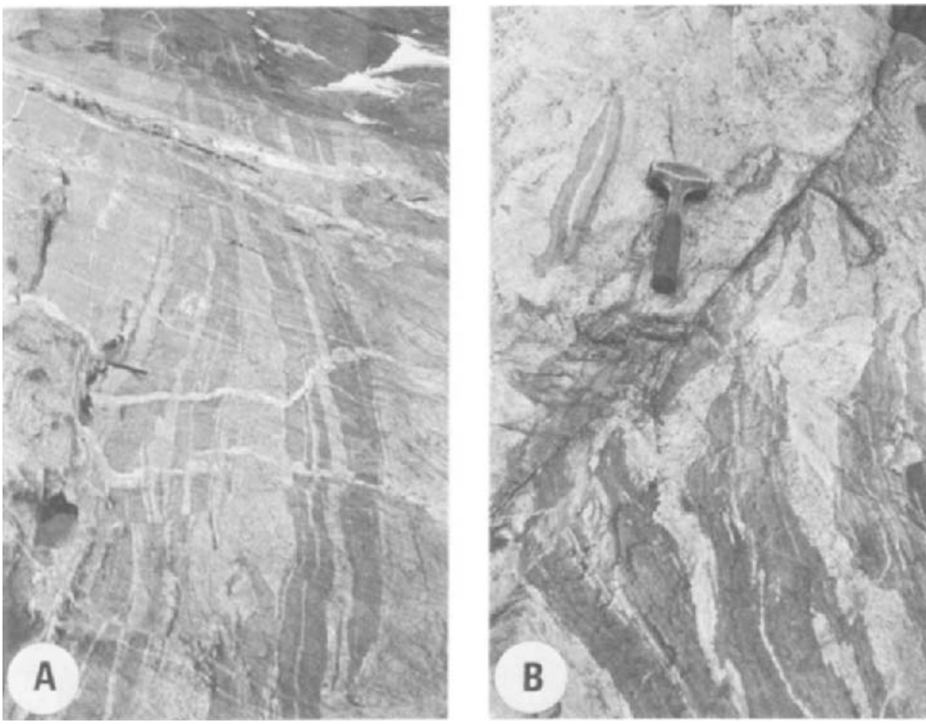


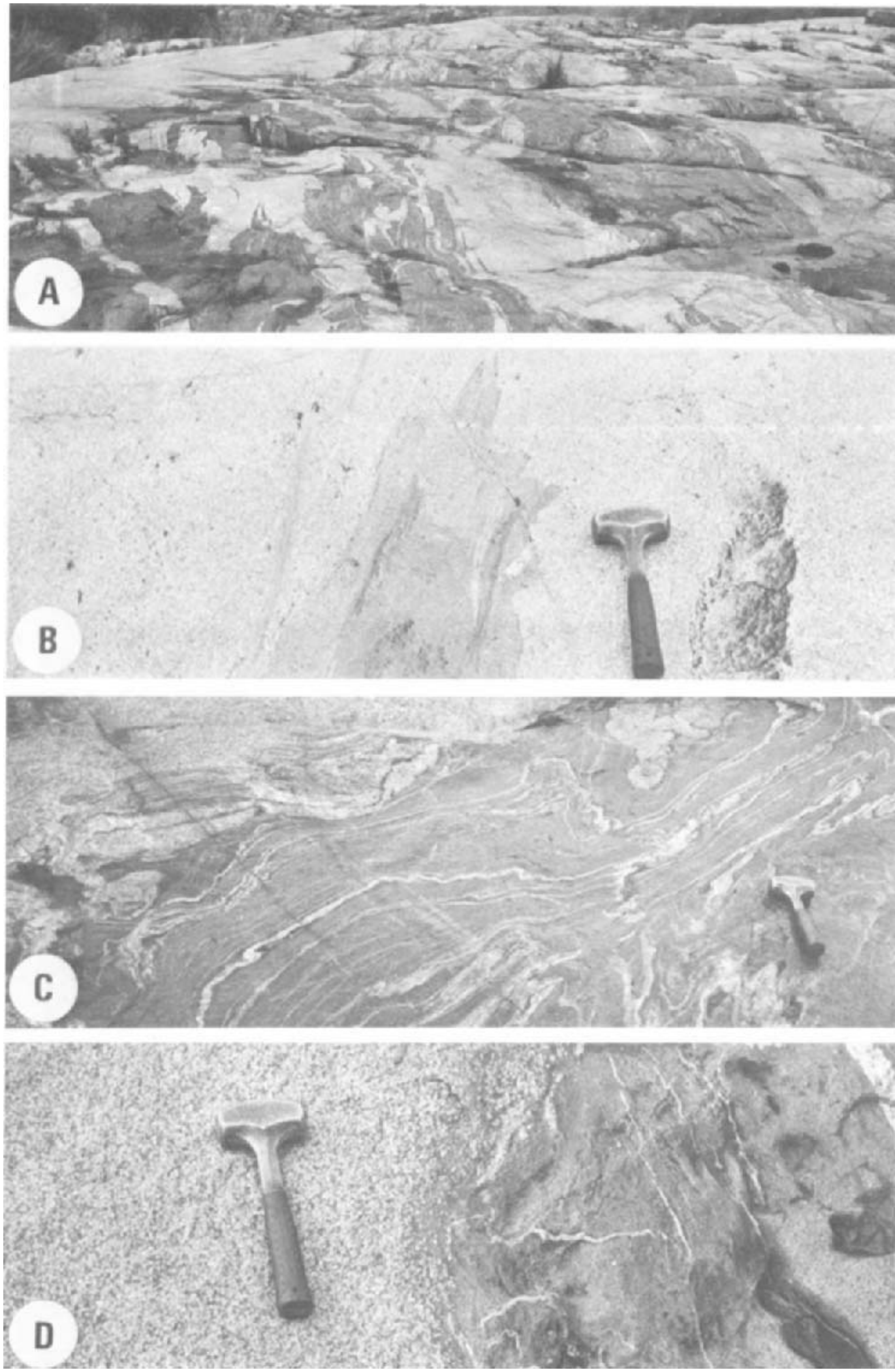
Fig. 6-2. Migmatites formed by granite—greenstone interaction.
A. "Lit-par-lit" amphibolite layering in trondhjemitic gneiss south of the Barberton greenstone belt. In the field the amphibolite bands can be followed along strike until they eventually merge into a large greenstone xenolith.
B. Bands of amphibolite intruded and disrupted by trondhjemite producing partially dissolved paleosome "rafts" in the slightly heterogeneous neosome.

As a preferred evolutionary model for the Archaean crust in the Barberton region it is suggested (as was done for northeastern Minnesota — Arth and Hanson, 1975) that there was probably no older continental crust in the immediate area because:

(1) There is no stratigraphic evidence for it in the Onverwacht volcanic successions (Viljoen and Viljoen, 1969a, b; Anhaeusser, 1978).

(2) The area has as yet yielded no isotopic age that, within experimental error, is greater than 3540 Ma.

(3) The available initial $^{87}Sr/^{86}Sr$ ratios for the 3.1 to 3.3 Ga old igneous rocks lie on a Sr-evolutionary curve that suggests the oldest units have a close genetic relationship with the upper mantle and that younger rocks may be related to the older ones (Davies and Allsopp, 1976). The low initial Sr ratios (Ro) appear to preclude derivation of these rocks from pre-existing continental crust.


A
B
C
D

(4) Trace-element models suggest that the older trondhjemitic and tonalitic gneisses can be generated without considering a derivation from, or assimilation of pre-existing crust (Glikson, 1976c; Hunter et al., 1978). The geochemistry of the tonalites and trondhjemites also probably precludes an original sialic basement, as far more fractionated granitic magmas would have ensued by partial melting of the latter.

TOWARDS AN EVOLUTIONARY MODEL

In attempting to model the evolutionary history of the Barberton Mountain Land and surroundings it must be remembered that over 1000 million years of pre-greenstone geologic time are not represented in the available record. The nature of the crust during this period must remain speculative but was probably characterized by great mobility caused by steep thermal gradients. Burke and Dewey (1973) named this and the subsequent period of Archaean greenstone development the "permobile phase" of earth's history. In their view this mobility was so great as to preclude the development of any form of plate-tectonic system. In the latter part of the permobile regime ($\simeq$ 3.5—2.5 Ga) Burke et al. (1976) envisaged that most of the continental material had differentiated and consisted of numerous small masses but that the remaining two-thirds of the earth's surface was oceanic. Because heat-generation was very much greater, thermal energy was dissipated through greater ocean ridge-type activity. If this appears reasonable it could be assumed that convection cells or centres of "hot-spot" activity would be closely spaced (hundreds, rather than thousands, of kilometers in the past — Fyfe, 1978) and crustal tectonics may have been influenced by a dominantly vertical component in the manner visualized by Ramberg (1967).

Adapting this line of reasoning to southern Africa it could be argued that the ancient successions recorded in the Limpopo belt might qualify as one of the early, small continental blocks that withstood destruction and that the areas to the north and south were dominated largely by a primitive ensimatic domain (*see also Barton and Key, this volume, Chapter 8, ed.*). In this latter environment a small cell convection pattern may have existed (Fig. 6-4a), resulting in the rapid upwelling of high-magnesian basalts and komatiitic peridotites which flowed laterally from the areas of convective upwelling.

Fig. 6-3. Styles of granitic interaction with greenstone xenoliths.

A. River pavement showing extensive fragmentation and migmatization of amphibolites near margin of a large lower Onverwacht xenolith (Sandspruit Formation) south of the Barberton greenstone belt.

B. Nebulitic migmatite. Amphibolites projecting into invading trondhjemitic gneisses display progressive stages of resorption (metasomatic granitization).

C. Tonalitic gneiss (upper left) intruding heterogeneous grey gneiss (metamorphosed greenstone unit, Anhaeusser and Robb, 1980a). Anatectic melting results in the development of contorted felsic veins in the migmatite.

D. Ferromagnesian-rich tonalite (left) formed as a result of assimilation of an amphibolite xenolith (basaltic komatiite of the Sandspruit Formation).

Progressive thickening of the volcanic piles and subtle density contrasts (cooler, higher-density submarine lavas above a hotter, lower-density ensimatic substratum) would be sufficient to initiate gravitative subsidence, the latter partly assisted by convective processes. Once started the process would progressively accelerate and large segments of ensimatic crust ($\simeq$ 100 km long in places) would become gravitationally unstable, with the lower parts of the basaltic crust first being converted to amphibolite and then ultimately, at depths in excess of 100 km, to eclogite ($\rho \simeq 3.5$) overlying mantle peridotite ($\rho \simeq 3.3$) as suggested by Ringwood and Green (1966). The foundering of lithospheric slabs might be viewed as the vertical motion equivalent to the modern subhorizontal subduction process occurring along plate boundaries. Phase changes to amphibolite, basic granulite or eclogite would proceed and the generation of tonalitic liquid by partial fusion of the basic parent would ensue in a manner similar to that now well-documented by the experimental work of Green and Ringwood (1968) and Lambert and Wyllie (1972). Assuming, for instance, a geothermal gradient of 40°C/km (as outlined by Glikson and Lambert, 1976), partial melting of H_2O-saturated olivine tholeiite should commence at depths of approximately 20 km and segregation of dacitic (tonalitic) melts would be widespread at depths between 20 and 30 km. Under anhydrous conditions this process may only commence at depths of about 35 km. Liquids of this composition would, due to their lower densities ($\rho \simeq 2.8$), rise through the sinking ensimatic lithosphere as a series of diapiric plutons, thereby initiating processes of protocontinental growth and accretion. This event would manifest itself in the form of extensive areas of trondhjemite or tonalite, the latter virtually replacing the original basic and ultrabasic crust above the zone of foundering (Fig. 6-4b, c). The convective cells would appear to have been unlike modern spreading centres where the ocean floor generally moves away from the axis of an oceanic ridge as new crust forms. In the modern examples magnetic anomaly patterns are considered to reflect elongate zones of basaltic lava flows or dykes. The spreading, although horizontal, is achieved, essentially by vertical sheeted dyke emplacement and appears to preclude massive subsidence of the oceanic lithosphere. Only at distant sites of subduction are the calc-alkaline volcanic and plutonic rocks generated.

An alternative possibility, suggested by Glikson (1978), is a model involving rifting of simatic sections overlying mantle diapirs, together with concomitant adiabatic melting of the sima and ongoing volcanism in overlying troughs. Tonalitic magma would begin to segregate when the root zones of such simatic rifts intersect the solidus.

The massive influx of trondhjemitic magma, generated by partial melting of simatic lithosphere in a manner similar to that outlined experimentally by Green and Ringwood (1968) and Lambert and Wyllie (1972), either totally or partially replaced the overlying oceanic crust. Where only partial replacement occurred tracts of bimodal migmatite were developed, the

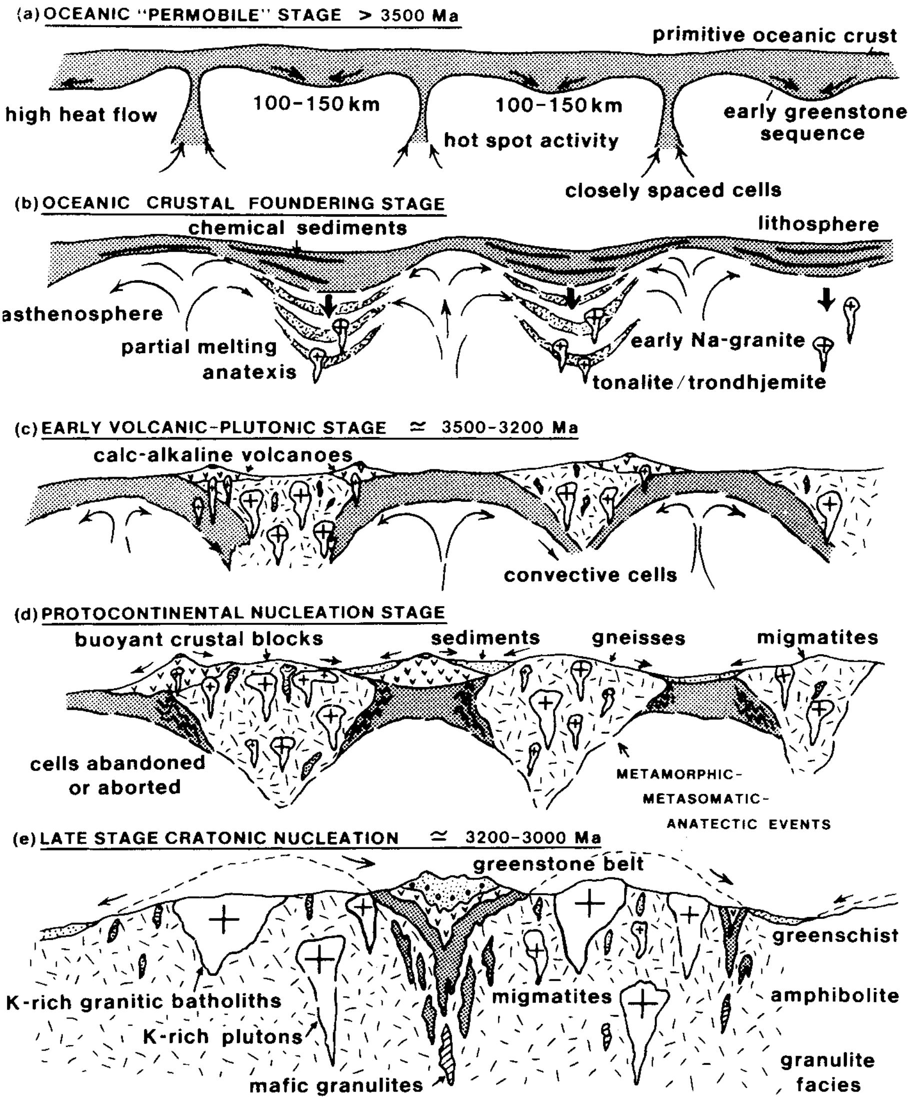


Fig. 6-4. Schematic cross-sectional sketches illustrating the progressive stages of Archaean crustal development in southern Africa. The preferred model, based on investigations in the Barberton granite-greenstone terrane, envisages that processes involving partial melting and granite underplating have been responsible for successive stages of transformation of primitive ensimatic crust and island-arc-type volcanic piles into complex migmatite-gneiss terranes in which major greenstone remnants survive as greenstone belts. The granitic rocks are later additions responsible for protocontinental nucleation and the development of stable cratons or shield areas.

latter terranes frequently shredded with primitive simatic greenstone remnants. Convective cells probably became self-destructing with time but may have contributed sporadic spurts of volcanism of a calc-alkaline or island-arc-type at the interface between the newly formed protocontinents and the remnant simatic domains (Fig. 6-4c). It is conceivable that these interface regions preserved sequences of "older" and "younger" greenstones. In some places conformable relationships between simatic sequences and island-arc-type assemblages were preserved (Barberton and Murchison greenstone belts) and juxtaposed with the ancient gneissic terranes. Major greenstone belts, like the Barberton example, were preserved near or adjacent to the sites of original convective upturn where the successions were spared large scale destruction by anatexis. However, they ultimately found themselves wedged between large, buoyant, sialic masses and eventually underwent gravitational infolding (Fig. 6-4d). This resulted in the development of variably plunging, isoclinal folds formed in preferentially developing synclinoria, and steeply inclined longitudinal faults or slides were generated, the latter frequently eliminating intervening anticlinal folds (Stage 1, Anhaeusser, 1975).

Deeply infolded greenstone belts like Barberton and the Murchison Range could have had their root-zones affected by differential anatectic melting, producing what may have been discrete second-generation trondhjemite/tonalite plutons like those found emplaced around the greenstone belt margins (Stage 2, Anhaeusser, 1975). However, there is no evidence to suggest unequivocally that the Stage 1 (linear) structures were related to early tonalite intrusions rather than reflecting early tectonic lineaments in the simatic crust. The tonalite bodies flanking the greenstone margins have been referred to as second-generation plutons because they deform sedimentary sequences which contain granitic detritus. These sediments may equally have been derived from successive stages of unroofing of piercement granitoid masses that eventually themselves caused deformation as a consequence of continued buoyancy and the development of discordant diapirs. Structures produced during Stage 1 were predominantly linear features (major folds and faults) whereas the deformation produced by Stage 2 diapiric plutons was responsible for the superimposition and intensification of the greenstone belt structural complexity.

Some might wish to see the events described as reflecting a number of mutually exclusive steps in the development of the Archaean crust and, in support of this, age differences in the range from 3.3 to 3.1 Ga are apparent in the trondhjemite/tonalite gneisses surrounding the Barberton greenstone belt (Oosthuyzen, 1970; Davies and Allsopp, 1976). It is, nevertheless, the preferred interpretation of the writer to view these granitoid crustal additions as being partial melt responses to a continuous process of magma generation consequent on the destruction of ensimatic lithosphere (komatiite basalts and peridotites). Remnants of this material occur now as the basal stratigraphy of the Onverwacht volcanic pile (Tjakastad Subgroup, Anhaeusser, 1978) as well as xenoliths in the surrounding granites.

Discussion, up to this stage, has dealt mainly with the volcanic components of the Barberton greenstone belt and some of the early granitic events consequent on the destruction of the primitive lithosphere. There is, however, within the greenstone sequence, no evidence for the existence of any neighbouring continental crust having been present at the time the volcanics were deposited.

Suggestions have been made that a marginal basin model, of the type similar to the "Rocas Verdes" complex in southern Chile, provides a satisfactory actualistic counterpart of Archaean greenstone belt development (Burke et al., 1976; Tarney et al., 1976; Windley, 1977). However, a serious difficulty with this approach, and acknowledged by these authors, includes the absence, in greenstone belts, of mafic igneous rocks in the form of gabbro — sheeted dyke — pillow lava sequences (i.e. typical ophiolites). Likewise, the modern marginal basins have only very rare developments of basaltic komatiite but are devoid of komatiitic peridotites as well as some of the minor, yet characteristic, components found in most greenstone stratigraphy. These include high alumina felsic interlayers in the high magnesian mafic and ultramafic successions, banded iron-formations (oxide-sulphide-carbonate-silicate facies types), and calc-silicate members, the latter possibly representing altered pelagic sediments. The cyclical volcanicity, with rapid alternation of rocks ranging from peridotite, Mg-rich basalt, tholeiitic basalt, dacite, rhyodacite, rhyolite and chert, also does not feature as prominently in the modern marginal basins.

These difficulties, particularly those relating to the nature of the volcanism and mafic igneous series in the Archaean and modern settings, did not appear insurmountable to Tarney et al. (1976) who appealed to the higher geothermal gradient in the Archaean to account for and influence crustal behaviour. Tarney et al. (1976) conceded, however, that the "Rocas Verdes" complex mirrors a single greenstone belt but is not readily accountable for multiple sub-parallel greenstone belts such as those of the Canadian Superior Province, the Yilgarn Block of Western Australia and the eastern Transvaal, South Africa. An added difficulty would emerge in their argument if it were to be established unequivocally that there had been no earlier protocontinental masses at the time of greenstone formation. Their marginal basin model requires back-arc tension and crustal thinning of older continental crust — a factor by no means yet established in the case of the Barberton region. Further, a back-arc spreading-basin model, implying lateral isotopic age zonation, should be observed across granite-greenstone terranes, but such has not been reported to date.

In the Barberton evolutionary model the stage following the infolding of the early volcanic successions (Fig. 6-4d) was marked by the onset of clastic sedimentation which first comprised considerable quantities of greywackes and shales (Fig Tree Group) derived from the erosion of early Ni-and Cr-rich volcanic rocks but which showed, stratigraphically upwards, an increasing proportion of detritus derived from sialic sources (Condie et al., 1970). These and ensuing arenaceous assemblages, comprising conglomerates, quartzites, subgreywackes and shales (Moodies Group), were developed as a result of the sporadic, but progressive, unroofing of the buoyant granitic rocks that began to envelop the Barberton greenstone remnant.

The next development in the chronology of the region (Fig. 6-4e) relates to the emplacement of batholithic massifs of potassic granite, both to the north (Robb, 1978) and to the southwest of the Barberton greenstone belt (Anhaeusser and Robb, 1980a, b). These potassic granites (including coarse porphyritic granites, adamellites and granodiorites) range in age from approximately 3.2—3.0 Ga and are characterized by low initial $^{87}Sr/^{86}Sr$ ratios (E. S. and J. M. Barton, unpubl. data; Oosthuyzen, 1970; Robb, 1978). The absence of published information on these potassic bodies places constraints on interpretations of their genesis, but it appears that they may have resulted from processes associated with the mantle or lower crust, and were probably not involved in any significant crustal pre-history. However, derivation of the K-rich granites by anataxis of trondhjemites and tonalites of the type found in the area and which are known to possess very low Rb/Sr ratios is possible and cannot be ruled out at this stage. Precisely what triggered the invasion of the crust by these K-rich granite bodies remains problematical. They are extensively developed not only on the Kaapvaal craton but also on the Rhodesian craton. Their style of emplacement again suggests vertical plutonism (or diapirism?) resulting in the elimination of vast tracts of earlier tonalitic or trondhjemitic gneiss, greenstones and Na-rich migmatites. The batholithic bodies are relatively homogeneous except near their margins where K-rich migmatites predominate. These migmatites display higher initial $^{87}Sr/^{86}Sr$ ratios (Ro $= 0.7052 \pm 0.0019$, De Gasparis, 1967), supporting the idea that they represent the products of crustal reworking of the earlier siliceous gneisses and their contained greenstone relics. The erratic REE distributions shown by Glikson (1976c) for migmatites north of the Barberton greenstone belt further strengthen the view that concomitant fusion of mafic and/or ultramafic rocks with the ancient tonalites took places when the batholiths intruded.

The earlier emplacement of the tonalites and trondhjemites was responsible for a major crustal thickening episode. At depth these rocks are probably transitional into siliceous granulites. Condie and Hunter (1976) visualized "plume subsidence" at this stage in their evolutionary model, the latter resulting in the crust sagging downwards into an area where the melting point of the siliceous granulites in the lower crust was exceeded. The magmas produced in this way were believed to be manifest in the K-rich bodies described above as well as in the series of later potassic plutons found in the Barberton region and in Swaziland. Existing radiometric ages indicate emplacement times for these later plutons at between 3.2 and 2.5 Ga (De Gasparis, 1967; Oosthuyzen, 1970) and their initial $^{87}Sr/^{86}Sr$ ratios ($Ro = 0.7065 \pm 0.0016$) are in agreement with the proposed derivation from siliceous crustal reworking. However, as mentioned earlier, criteria diagnostic of the derivation of K-rich magma from siliceous granulites, as opposed to tonalitic or trondhjemitic gneisses, are not unequivocal as chemically both these sources are akin. Some fundamental change from the Na-rich granitic regime to the K-rich regime clearly took place in the Barberton region at approximately 3.2—3.0 Ga ago. Perhaps this was a consequence of the secular changes brought about by crustal thickening or, more likely, a pronounced acceleration of thermal activity (a response to the radioactive decay of greater amounts of K, U and Th?) towards the end of the Archaean era — the latter defined at 3.0 Ga on the Kaapvaal craton (Anhaeusser, 1973).

SUMMARY

(1) The preferred model of crustal evolution in the region of the Barberton Mountain Land acknowledges the similarities that exist with modern oceanic island-arc or trench-back arc-marginal basin analogues. These similarities appear to justifiably allow for the interpretations that have been suggested for the evolution of the granite-greenstone terranes of the shield areas in terms of convection, subduction, partial melting and upwelling of a wide range of volcanic and plutonic products.

(2) The preferred model, however, focuses attention on the differences that exist between the ancient and the modern examples and places constraints on direct comparisons between Archaean and later plate-tectonic derived crustal segments.

(3) There is little or no evidence in the Barberton greenstone belt or in the surrounding granite-greenstone terrane that suggests plate tectonics, in the modern sense of the term, was responsible for the development of the region.

(4) The available evidence, both within the Barberton greenstone belt stratigraphic record, and in the surrounding granitic terrane, does not convincingly support the prior existence of any ensialic crust older than the 3540 Ma basaltic and peridotitic komatiites that constitute the bulk of the lower Onverwacht stratigraphy and which also predominate in greenstone xenoliths distributed across the entire Kaapvaal craton.

(5) The basal greenstone belt stratigraphy in the Barberton Mountain Land is interpreted as a relic of the ancient ensimatic crust in the region and material similar to this was the likely source of the poly-domal tonalite/trondhjemite magmas that streamed diapirically upwards from sites where the primitive ensimatic crust had foundered.

(6) Gravitational instability, coupled with convective upwelling, produced an essentially vertical tectonic domain in which volcanic and plutonic rocks of a wide range of compositions were transported to the earth's surface to

form both the "younger" greenstones (the calc-alkaline upper Onverwacht stratigraphy) and the ancient protocontinental sialic nuclei.

(7) The buoyant granites were successively unroofed to provide clastic detritus in the downsagging troughs wedged between the buttressing masses. Deformation intensified and the greenstones inherited their structural and metamorphic characteristics from diapiric gneiss plutons impinging on the margins of the main greenstone remnant.

(8) Later potassic granite batholiths and small discrete plutons were emplaced into the area again emphasizing the dominant vertical tectonics mechanism that characterize the Archaean basement of southern Africa.

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Chapter 7

THE EVOLUTION OF THE RHODESIAN CRATON AND ADJACENT ARCHAEAN TERRAIN: TECTONIC MODELS

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ABSTRACT

The Rhodesian craton has been a coherent tectonic unit with little internal deformation for at least 2.6 Ga. Evidence from within the craton suggests that a continental nucleus has existed since early Archaean time, and the main (Bulawayan) greenstone belt sequences were erupted onto preexisting continental crust. The Limpopo belt to the south of the craton has had a similarly long history. Various tectonic models are discussed to explain the geological development of the craton: since the craton has behaved as an integral unit with little internal deformation it has been a plate (or part of a plate) since the Archaean, but the ancient tectonic regime may have been very different from that of today.

INTRODUCTION

The Rhodesian craton and adjacent Archaean terrain contain some of the better exposed and preserved Archaean rocks. These range in age from c. 3.8 to 2.5 Ga*. The craton is covered by younger rocks in the west, bounded by younger fold belts in the north and east and passes into the Archaean gneisses of the Limpopo mobile belt to the south and south-west. The Limpopo belt abuts the northern margin of the Archaean Kaapvaal craton.

To discuss possible plate-tectonic processes in the Archaean it is necessary to identify Archaean tectonic environments and to compare these environments with those resulting from modern plate-tectonic processes. In particular the recognition of evidence for constructive plate boundaries through relict oceanic crust, or convergent plate boundaries through their distinctive volcanic, metamorphic, structural or sedimentological associations could be taken to indicate plate-tectonic processes. Recognition of tectonic features distinctive of plate tectonics in areas older than the Mesozoic is inevitably hampered by the self-obliterating nature of the plate-tectonic cycle of creation and destruction of oceanic crust. In the Archaean, comparison of tectonic environments with possible modern counterparts is further hindered by uncertainty as to the nature of the tectonic processes

* Rb-Sr ages are recalculated using a decay constant for Rb of $1.42 \times 10^{-11} a^{-1}$. All errors are 1σ.

responsible for Archaean rocks. This uncertainty is reflected by the very different views published on fundamental questions concerning Archaean tectonics, such as the nature of the basement to the Archaean sedimentary-volcanic greenstone sequences. In part this is due to the limited areas of Archaean rocks preserved, poor exposure and tectonic complexity. Undoubtedly it also results from the lack of detailed structural mapping in Archaean terrains compared with younger tectonic areas such as the Alps.

In the Archaean of Zimbabwe (Rhodesia) we consider that detailed mapping by geologists of the Geological Survey, mostly at a scale of 1: 100,000, over the last 70 years, coupled with more recent detailed structural and geochronological studies by geologists from a number of universities have provided answers to at least two of the questions which have bedevilled recent discussions of Archaean tectonics. Firstly, it is apparent that greenstone belt sequences within the craton differ significantly both in place and time and probably evolved in more than one tectonic setting. Secondly, as summarized by Wilson et al. (1978), and discussed in detail by Wilson (1979), there is good evidence for an older continental basement to the main greenstone belts (Bulawayan and Shamvaian Groups). Major stratigraphic units within these belts can be correlated with some confidence across most of the craton.

It is generally accepted that Archaean granite-greenstone terrains can be distinguished from younger orogenic areas. It is difficult to define their distinctive features or even the significance of the difference. However, stratigraphic sequences dominated by basaltic volcanics preserved in synclinal belts between areas of granite and gneiss appear to be restricted to the Archaean, as do the more magnesian peridotitic komatiite lavas which form small but distinctive parts of these sequences.

STRATIGRAPHIC SUMMARY

Before discussing tectonic schemes it is necessary to summarize the stratigraphy of the area. For details the reader is referred to recent publications such as Wilson et al. (1978), Wilson (1979) and Stagman (1978).

The oldest rocks are those of the basement complex gneisses around Beitbridge, in the Limpopo belt (Light and Watkeys, 1978), which have been dated (in South Africa) at around 3.79 Ga (Barton et al., 1977, 1978), with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7012 (*see also Barton and Key, this volume, Chapter 8, ed.*). Younger rocks in the area include supracrustals, consisting of shallow-water sedimentary rocks such as arkose sandstones, pelites, limestones and quartzites, as well as some lavas, now all metamorphosed to high grade. Some rocks contain evidence of metamorphism under conditions of about 11 kb and 860°C (Chinner and Sweatman, 1968). This metamorphism probably took place over 3 Ga ago.

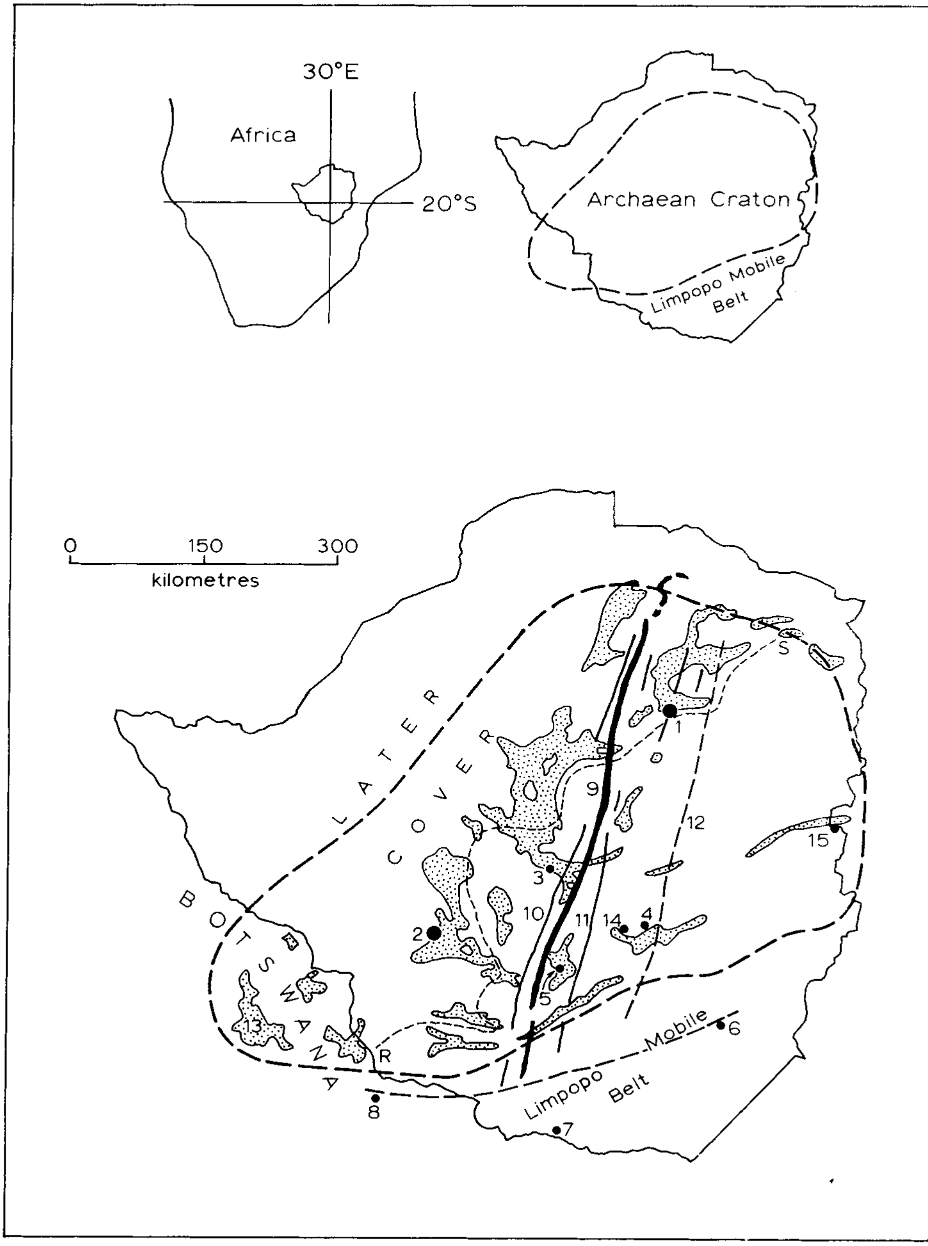

Fig. 7-1. Outline map of the Rhodesian Archaean craton and Limpopo belt. Stippled areas indicate greenstone belts: *1* = Salisbury; *2* = Bulawayo; *3* = Gwelo; *4* = Fort Victoria; *5* = Belingwe; *13* = Matsitama. Other features: *6* = Triangle area; *7* = Beitridge; *8* = Pikwe; *9* = Great Dyke; *10* = Umvimeela Dyke; *11* = East Dyke; *12* = Popoteke Fault; *14* = Mashaba; *15* = Umtali. Dashed line *8* to *6* approximately shows Tuli-Sabi shear zone. Dashed line *R* to *S* indicates the eastern limit of the bimodal and calc-alkaline suites in the west of the craton in the 2.7 Ga old Upper (Bulawayan) Greenstones.

On the Rhodesian craton itself there are only scattered remains of c. 3.5 Ga old rocks, but it is evident that an extensive continental terrain had been established by this time. This extended from the Transvaal to perhaps the Salisbury area, although within Zimbabwe (Rhodesia) conclusive radiometric evidence is restricted to the south of the country.

On this pre-existing terrain a series of "events" took place. The dominantly sedimentary supracrustal rocks around Beitbridge have already been noted. Much better known are the "Sebakwian" greenstone-belt lavas and sediments of the Lower Gwelo, Selukwe and Mashaba areas. Elsewhere in much of the c. 3.5 Ga terrain, and particularly well developed between Selukwe and Mashaba, are abundant, strongly foliated, greenstone belt remnants infolded with gneisses. Clearly there had been an extensive history of supracrustal rocks by c. 3.5 Ga ago.

The Selukwe area is of great significance and is worth considering in some detail. Cotterill (1976; 1979) and Stowe (1968) have shown that the c. 3.5 Ga old (or older) rocks consist of a variety of lavas and metasedimentary rocks. A lower sequence includes basalts and komatiites as well as minor sediments containing quartz clasts and ultramafic detritus. Later, after emplacement of a number of ultramafic bodies, followed by deformation and erosion, the Wanderer formation was deposited. Its basal beds include conglomerate with clasts of talc-carbonate rocks, chromitite, metabasalt, jaspilite, chert, granite and gneiss. The overlying rocks are silts and sandstones as well as oxide, silicate, carbonate and sulphide facies ironstones. This all indicates an extensive and varied source terrain of considerable relief. The topmost part of the preserved succession consists of basaltic lavas. Cotterill (1979) has speculated that a succession of Selukwe type may have covered a large area of southern Zimbabwe (Rhodesia). Re-examination of the Mashaba area (Wilson et al., 1978) and recent work in the Lower Gwelo area (Cheshire et al., 1977), have attributed similar lithologies to the Sebakwian Group, the oldest development of greenstone belts in Zimbabwe (Rhodesia).

Cotterill (1976, 1979) and Stowe (1968) have shown that the extensive Sebakwian terrain around Selukwe is inverted and was emplaced as a nappe or series of nappes. This inverted terrain covers at least 1200 km^2 as now exposed and movement may have been 20 km or more (Cotterill, 1976). Emplacement must have occurred before intrusion of the crust-derived Mont d'Or granite (3.35 ± 0.06 Ga, initial $^{87}Sr/^{86}Sr = 0.711 \pm 0.002$, Moorbath et al., 1976). Key et al. (1976) have suggested that this nappe may have extended over hundreds of kilometres into Botswana but no evidence for this has been found in Zimbabwe (Rhodesia). Coward et al. (1976a, b) have reported major overturned structures in the Matsitama and adjacent greenstone belts in Botswana, but these structures are probably of the same age as major deformations in the Limpopo Belt (c. 2.7—2.5 Ga), and the greenstone belts involved may be significantly younger (2.9 or 2.7 Ga) than the Selukwe rocks.

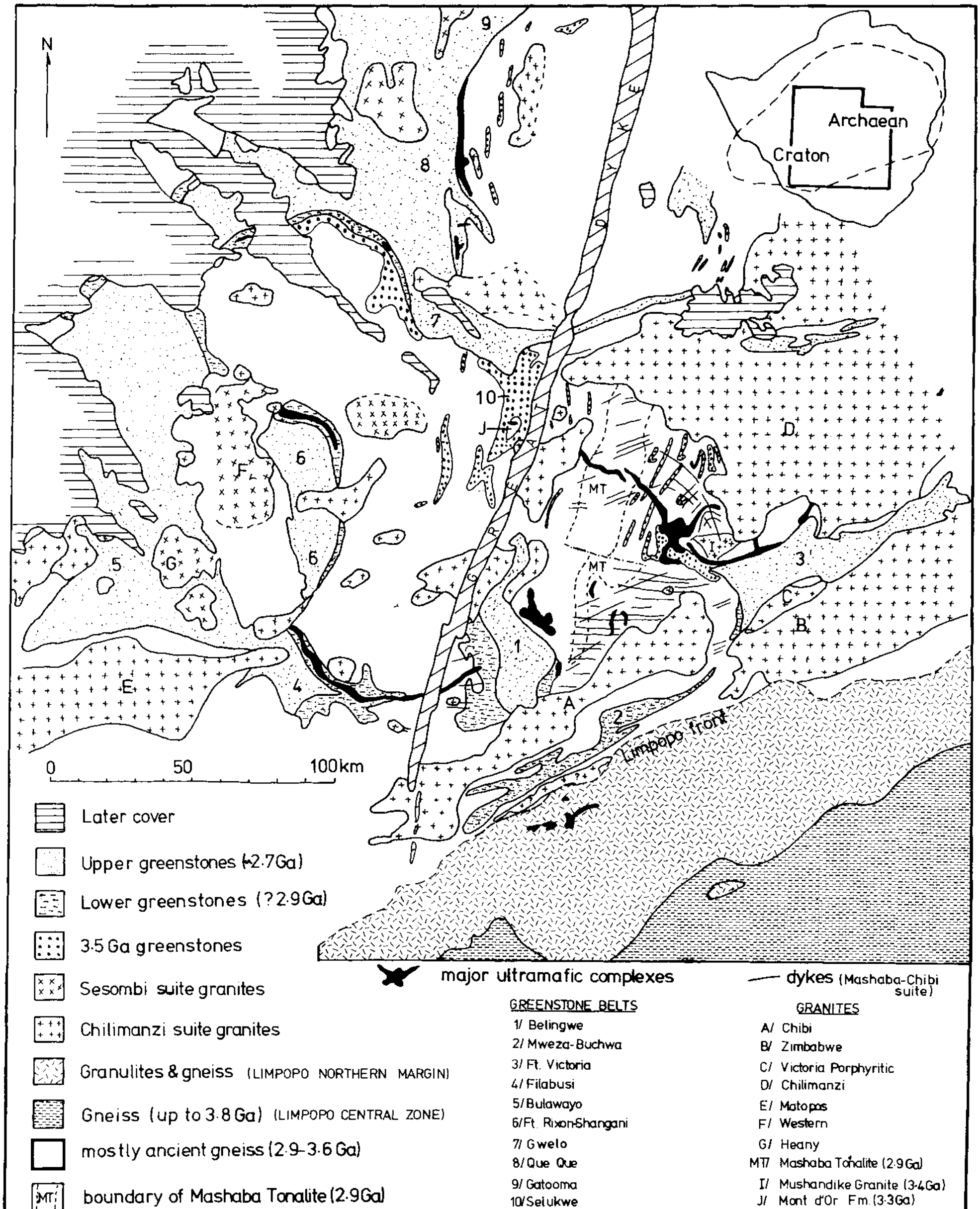

Fig. 7-2. Geological sketch map of part of the Rhodesian Archaean craton and Limpopo belt, modified after Wilson (1979). The northern marginal zone of the Limpopo belt consists of a granite-greenstone terrain of various ages at generally high metamorphic grade; the central zone contains varied gneisses and supracrustal rocks. The various ultramafic complexes shown on the craton constitute the Mashaba Ultramafic Suite and are possibly related to the 2.7 Ga old Upper Greenstones.

An extensive c. 2.9 Ga old granite-gneiss terrain has recently been identified within the area between Belingwe and Mashaba (Fig. 7-1) (Wilson et al., 1978; Hawkesworth et al., 1979) and this may extend into the Limpopo belt. Hickman (1976) has dated granulite facies rocks from the Bangala area at 2.87 ± 0.06 Ga (initial $^{87}Sr/^{86}Sr = 0.7014 \pm 0.0012$). In the Belingwe area (Orpen, 1978; M. J. Bickle and E. G. Nisbet, unpubl. data; Fig. 7-2) an important lower sequence of greenstones is recorded. This has not been dated satisfactorily but the structural and sedimentological relations suggest that it was deposited on 3.5 Ga as well as on 2.9 Ga old gneissic crust and that it is overlain unconformably by the 2.7 Ga old greenstone sequence. Accordingly it is at present provisionally regarded as "Lower Bulawayan" in local stratigraphic terminology. A range of lithologies is involved. The upper parts embrace an extensive suite of peridotitic and basaltic komatiites, coarse clastic sedimentary rocks including clasts derived from a pre-existing granite-gneiss-greenstone terrain, siltstones, quartzites and ironstones. The lower part comprises dominantly dacitic pyroclasts and flows. In detail, regional correlation of these rocks is not certain, but it is probable that the Mweza and Umtali belts and parts of the Salisbury, Lower Gwelo, Filabusi and perhaps Fort Victoria belts are of similar age (Wilson et al., 1978). These greenstone belt rocks of possible 2.9 Ga age were almost certainly wholly deposited on the pre-existing continental craton which included the remains of the Sebakwian (3.5 Ga) greenstone belts. This conclusion is based on the sedimentological evidence which indicates deposition in a basin surrounded at first by high-relief granite-gneiss-greenstone terrain, and also on the inference of a basal unconformity in the eastern development of the sequence at Belingwe (Bickle et al., unpubl. data).

The most studied and most widespread greenstone belts in the Rhodesian craton are the "Upper Greenstones" of the Belingwe area (Bickle et al., 1975; Nisbet et al., 1977) and elsewhere, known as the "Upper Bulawayan". These, and the locally developed overlying sedimentary Shamvaian Group, are about 2.7 Ga old (Wilson et al., 1978). They were deposited unconformably on pre-existing granite-gneiss-greenstone terrain, including the Lower Greenstones, and contain sediments (mainly shallow-water and including stromatolites), peridotitic and basaltic komatiites and thick sequences of tholeiites and andesites. Wilson et al. (1978) have shown that a correlation may be set up across the Rhodesian craton over a lateral distance of more than 200 km and a length of perhaps 700 km. In the east and in the west the lower part of the sequence consists of a thick succession of komatiites and tholeiites. In the west this succession, in its upper part, is interbedded with dacite flows and pyroclasts; this essentially bimodal western succession is overlain by a thick andesite-dominant calc-alkaline suite (Harrison, 1970; cf. also Hawkesworth and O'Nions, 1977). Soon after their formation and before the intrusion of the Great Dyke, the Upper Greenstones were intruded by granites. In the west the largely tonalitic Sesombi Suite may have been mantle-derived (initial $^{87}Sr/^{86}Sr = 0.701$), whereas in the south and east the more potassic Chilimanzi Suite may have had a significant crustal component (initial $^{87}Sr/^{86}Sr = 0.7025$—0.7045, Hawkesworth et al., 1979). Robertson (1973) has suggested that part of the Chilimanzi Suite may have been derived from the Limpopo belt terrain. Isotopic evidence (Hickman, 1978) suggests derivation from 2.9 Ga (but not older) terrain.

A series of dyke swarms of various ages has intruded the craton. These include the dominantly east—west-trending swarms of the pre-Chilimanzi Suite, Mashaba-Chibi dykes (? 2.7 Ga) and the post-Great Dyke NNW-trending Sebanga Poort swarm (Wilson, 1973). The Great Dyke was intruded at 2.46 ± 0.016 Ga ago (Hamilton, 1977), and the Umvimeela and East Dykes and Popoteke Fault are probably of comparable age.

Fig. 7-3. LANDSAT photograph of the SE part of the Rhodesian craton, showing an area broadly similar to Fig. 7-2. The continuity of the ancient basement across the Limpopo front shows clearly. Note that some prominent rectangular features (e.g. NE of the Belingwe belt) are land-use, not geological patterns.

RELATIONSHIP BETWEEN THE RHODESIAN CRATON AND THE LIMPOPO MOBILE BELT

The boundary between the craton and the Limpopo belt has been studied by James (1976) in the area south of Fort Victoria. The first phase of deformation noted produced large scale SW- to NE-trending upright folds with subhorizontal plunge both in the Fort Victoria greenstone belt and in the granulite terrain. During the second deformation heterogeneous simple shear deformed much of the northern margin of the Limpopo belt with northerly overthrusting. The contact between the Limpopo belt and the Rhodesian craton south of Fort Victoria is marked by a low angle ductile shear zone in which Limpopo granulite was thrust over cratonic granite-gneiss from southeast to northwest. The contact here is a zone of extensive mylonitization. James (1976) noted that the thrust dies out westwards which is consistent with Robertson's (1973) observation that farther west the contact is gradational.

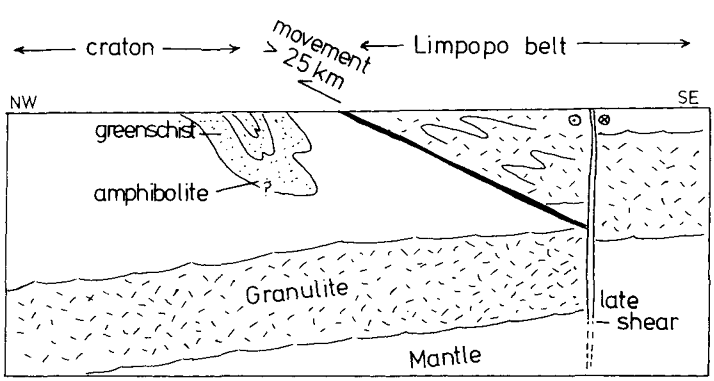

Fig. 7-4. Sketch section from NW to SE across the Rhodesian craton/Limpopo belt contact in the Fort Victoria district. Main thrust probably of late Archaean age. Further SW the contact is more gradational. Arrowhead and tail in circles show lateral component of movement direction on late (2.0 Ga?) vertical shear (for simplicity the late shear is here assumed to be vertical). The granulites are assumed to have extended under the southern margin of the craton. Not to scale.

The third deformation noted by James (1976) was most marked in distinct shear belts, in the Fort Victoria greenstone belt, along the Tuli-Sabi shear zone and in the minor Manganu shear zone. In general this was a period of ductile dextral shear movement, and in the Triangle (= Tuli-Sabi) shear zone, which affects the mobile belt, the southern mass was also displaced upwards, over granulites, to the north.

The timing of these various deformations is as yet not clear, and James (1976) emphasizes their diachronous nature. His first and second deformations, however, in the Limpopo belt and in the craton, took place after the formation of the main "Upper Greenstones" and the Shamvaian Group of the Fort Victoria area (2.7 Ga) and before the intrusion of the adjacent members of the Chilimanzi Suite granites (2.57 ± 0.015 Ga, Hickman, 1978). In the Limpopo belt, however, certain porphyritic granites, which can be considered as an early phase of the cratonic Chilimanzi Suite, show effects of the second deformation (cf. Hickman, 1976). Movement in the Tuli-Sabi shear system has been suggested as predating 2.5 Ga by Key et al. (1976) and synchronous with the intrusion of the Great Dyke (2.46 Ga) by Coward et al. (1976a). James (1976), Hickman (1976) and Hickman and Wakefield (1975) favour dates closer to 2.0 Ga and this is more consistent with available evidence. Recent unpublished Sm-Nd work (C.J. Hawkesworth, pers. commun., 1979) indicates an age close to 2.0 Ga for the Triangle shear zone (qtz-cpx-plag mineral isochron). Renewed movement on earlier lines of weakness is indicated, with the later movements in Zimbabwe (Rhodesia) not represented in Botswana, except near Pikwe.

In summary, the Limpopo belt and Rhodesian craton appear to have been an integral unit for over 3.5 Ga. Movement between them has been restricted to intracratonic shearing, sliding and thrusting. James (1976) has estimated that the early thrusting of granulites over the craton south of Fort Victoria involved an uplift of 24 km and a horizontal northerly movement of some 26 km (Fig. 7-4). Coward et al. (1976a, b), in a regional appraisal of the c. 2.7—2.6 Ga deformations, argue for an overall southwesterly movement of the craton relative to the Limpopo belt of up to 200 km. James (1976) estimates that later (Triangle) shear zones recorded a displacement of 30—50 km. But, although there have been considerable movements jostling the cratonic elements of the area, no major break comparable either to the formation of an oceanic rift, to a subduction zone or to a substantial long-lived transform fault has been recorded from the area. In many ways the Limpopo belt can be regarded as another segment of the craton, uplifted over the Rhodesian craton (sensu stricto) and eroded to greater depth as a result. It is possible that episodes of brittle faulting on the Rhodesian craton (e.g. the late dextral movement in the Jenya fault system cutting the Great Dyke) were high-level equivalents of shear zones in the Limpopo belt.

CONSTRAINTS ON TECTONIC MODELS

Whatever the model adopted for the area there are certain constraining factors which must be considered.

(1) *A continental nucleus has been present since the early Archaean and, from the earliest time, there is sedimentological evidence for shallow-water deposition in some areas and for local high relief on land.* Since erosional processes were almost certainly rapid, this implies active mountain building events. Models such as those of Hargraves (1976) and La Barbera (1978) do not apply to this particular craton. Locally, and probably transiently, the crust (on metamorphic evidence, e.g. Chinner and Sweatman, 1968)

TABLE 7-I

Selected events in the Kaapvaal craton, Limpopo belt and Rhodesian craton. For aspects of this table not discussed in the text see Wilson et al., 1978, Barton and Key (this volume, Chapter 8) and Hawkesworth et al., 1979.

Age (Ga)	Kaapvaal craton	Age (Ga)	Limpopo belt	Age (Ga)	Rhodesian craton
1.79	Waterberg Supergroup (intrusives and extrusives)	1.77	Tuli and Soutpansberg troughs (b)	1.7–1.85	Umkondo and Mashonaland dolerites (c)
1.92	Bushveld Granites (a)	1.95	Termination of uplift (b)	2.0	Various mineral ages (e, l) Probable thermal event in Limpopo belt and southern margin of Rhodesian craton (l)
2.05	Bushveld layered mafic sequence (a)	2.0	Shear zones, uplift (d) Probable thermal event		
2.2–2.3	Transvaal Supergroup (a, f)		? Uplift and deformation (b)	(?)	Deweras, Lomagundi and Piriwiri Groups (lavas and sediments)
2.6	Ventersdorp Supergroup (f)	2.6	Deformation, ? uplift (b)	2.46	Great Dyke (g)
2.65	Witwatersrand Supergroup (a)	2.7	Bulai pluton (b)	2.6	Late granites, Chilimanzi and Sesombi suite (h)
2.8	Dominion reef acid volcanics (a)			2.6–2.7	Upper (Bulawayan) sequence of greenstones and Shamvaian Group
2.75–2.6	Post Pongola granites (a)			2.7	Gwenoro and Umniati gneisses (h)
3.0–2.8	Younger Barberton granites (i)	2.9	Metamorphism (b)	2.9	? Lower (Bulawayan) greenstones Bangala and Chingezi gneisses Mashaba Tonalite (h, j, l)
3.0	Pongola Supergroup (a)	3.06	Dykes (b)		
3.3–3.1	Gneisses in Barberton area (i)	3.15	Deformation (b)		
		3.35	Messina Layered Intrusion (b)	3.35	Mont d'Or Granite, Selukwe (h)

				3.4–3.5	Mushandike Granite
3.5	Onverwacht Group and probably whole sequence (k)	3.35–3.6	Supracrustal Sequence in Limpopo Belt (b)	3.5–3.6	(?) Deposition of Sebakwian Greenstones. Tokwe and Shabani Gneisses, local tonalites (h, e)
		3.6	Dykes in basement (b)		
		3.8	Basement gneisses (b)		? Old greenstone remnants in ancient gneiss
					? Ancient basement
			Possible deposition of sediments on older crust		

References: (a) Hunter and Hamilton, 1978; (b) Barton and Key this volume; (c) Stagman, 1978; (d) C.J. Hawkesworth, pers. commun., 1979; (e) Bickle and Hawkesworth, unpubl. data; (f) Button, 1976; (g) Hamilton, 1977; (h) Wilson et al., 1978; (i) Davies and Allsopp, 1976; (j) Hickman, 1978; (k) Hamilton et al., 1979. (l) Hawkesworth et al., 1979. Rb decay constant: $\lambda = 1.42 \times 10^{-11} a^{-1}$.

must have been up to 45 km thick. Even though this may have been a temporary feature it implies a strong continental crust overall.

(2) *No relict Archaean oceanic crust or ophiolite complexes have been identified.* It is possible that partial melting in Archaean oceanic ridges would have been very much greater than in ridges of the present day, producing a much thicker oceanic crust (Moores, 1973). Thus emplacement of oceanic fragments would not necessarily also mean the emplacement of deep level residual mantle; an Archaean "ophiolite" might consist of lavas, dykes and cumulate peridotites only. Such a body would be very difficult to identify as an "ophiolite" complex. Nevertheless the available evidence in Zimbabwe (Rhodesia) indicates that, at least, the main greenstone belts (? 2.9 and 2.7 Ga) were erupted on to continental crust. With the preserved remnants of the Sebakwian the position is not so clear, but there is no positive evidence of an Archaean ophiolite complex, let alone ancient oceanic crust.

(3) *The major stratigraphic units of the 2.7 Ga greenstone belts have been correlated across the craton.* If these correlations (Wilson et al., 1978; Wilson, 1979) are correct then it is probable that, at that time, a much more continuous cover of greenstone belt rocks with a lateral extent of several hundred kilometres overlay the cratonic area. On this premise the present-day distribution of the greenstone belts does not reflect an original pattern of separate isolated depositories although it may, in part, reflect original depositional basins in the continuous sequence (Fig. 7-5).

(4) *The "Upper Greenstones" contain a thick craton-wide sequence of komatiite and tholeiites, overlain by a major calc-alkaline sequence now preserved in a belt less than 100 km wide and extending over 700 km along the present western exposed margin of the craton.* Predominantly basaltic volcanism occurred repeatedly during the time interval ? 2.9—2.7 Ga; dacitic and andesitic volcanism also played a major part. The "Upper Greenstones" succession, however, is particularly intriguing.

The komatiite-tholeiite sequence bears most resemblance to lava suites from modern intra-continental rift settings thought to be associated with continental breakup prior to formation of oceanic crust. The continental basement, thin basal sediments and a 5—10 km thickness of near sediment-free sequence of magnesium-rich and tholeiitic lavas are very similar to the Tertiary sequence described at Svartenhuk, Greenland (Clarke and Pedersen, 1976). Many intra-continental rift volcanics, however, are highly alkaline,

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Model B. (1), (2) and (3) deposition of a uniform layer of sedimentary and volcanic rocks across the whole craton; (4) compression, formation of relatively open synclinal basins (note: this stage is absent in model A); (5) further compression in a different orientation, formation of tight to isoclinal synclines showing two phases of deformation, with either imposition of a S_2 penetrative cleavage, or rotation of any first cleavage into the new orientation. (6) uplift and erosion.
Both models are consistent with the stratigraphic correlations proposed by Wilson (1979), but are very schematic and do not show the necessary diachroneity of transgressions.

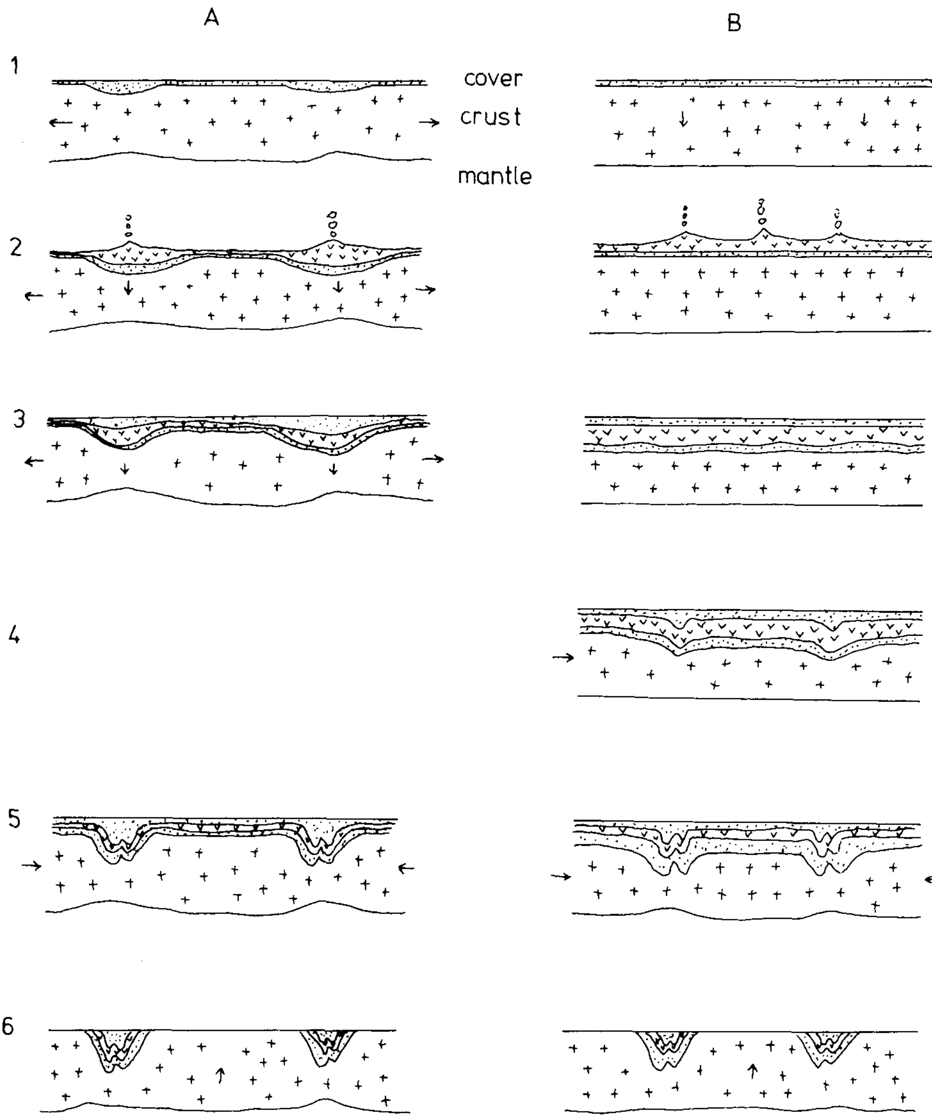

Fig. 7-5. Two schematic models of greenstone belt formation on the craton. Model A: (1) deposition of basal sedimentary rocks in linked basins produced by thinning of the crust under tension; (2) initiation of volcanism in subsiding basins; (3) further subsidence, cessation of volcanism and infill of the basins by later sedimentation. Note that by this stage the deepest rocks, laid down in phase (1), will be fairly steeply dipping; (4) see model B; (5) compression, deformation and tightening of synclinal structures; (6) erosion, leaving isolated synclinal structures, and deformational styles consistent with two phases of deformation, the first non-cleavage producing, the second marked by a penetrative cleavage (a characteristic feature of many Rhodesian greenstone belts). (Text continued on bottom of opposite page).

in contrast to the tholeiites and komatiites of greenstone belts. Even the less alkaline olivine-phyric lavas of Baffin Bay (Clarke, 1970) and Svartenhuk tend to be richer in TiO_2 and related elements. High-magnesian lavas occur in a number of other modern tectonic settings including island arcs, back-arc basins and oceanic spreading ridges, but we reject these environments as analogues of the komatiite-tholeiite volcanic phase of the Upper Greenstones because the structure and stratigraphy of the younger areas are very different. Cameron et al. (1979) have discussed this problem in comparing boninites with komatiitic basalts.

The calc-alkaline belt along the western margin of the Rhodesian Archaean craton has the same order of dimensions as modern calc-alkaline continental margins situated on destructive plate boundaries. The preservation of this belt may merely be a function of differential uplift and erosion, although the association of the tonalite-granodiorite plutonic suite with low $^{87}Sr/^{86}Sr$ initial ratios (Sesombi Suite, Wilson et al., 1978) with the belt is evidence that the calc-alkaline rocks may have been restricted in their original distribution. If so, these volcanics provide perhaps the best evidence for a continental-margin type, convergent plate boundary.

Subtle compositional differences between Archaean and younger andesitic volcanic rocks (Hallberg et al., 1976; Hawkesworth and O'Nions, 1977) and the uncertainty about the actual mechanism for the production of modern andesites combine to reduce the confidence with which inference on the Archaean may be made on the basis of rock compositions. Nevertheless the "Upper Greenstones" volcanic sequence appears to reflect at least two differing volcanic, and presumably tectonic, environments. If the modern plate-tectonic analogy is valid these would most closely correspond to initial rifting of continental crust with the formation of adjacent oceanic crust implied, followed by subduction of the oceanic crust in a convergent plate-tectonic setting (Fig. 7-6; Wilson et al., 1978).

(5) *Initial $^{87}Sr/^{86}Sr$ ratios suggest a variety of source histories for the continental crust.* Some granitoids (especially the Sesombi Suite tonalites) and the komatiites, tholeiites and andesites have low initial ratios, indicating a low Rb/Sr source, i.e. mantle or basalt. Other granites and gneisses (for example Chilimanzi Suite, Mont d'Or granite) have relatively high initial ratios suggesting crustal derivation. It is possible that these granites may have been produced by heating in the base of a thickened continental pile or after major thrusting events. One intriguing possibility is that the Limpopo Belt granulite facies gneisses, metamorphosed at pressures up to c. 11 kb, may be representative of the present crust underlying the main granite-greenstone areas of the craton. 2.9 Ga old gneisses within the Limpopo belt are suitable rocks for derivation of adamellites such as those of the Chilimanzi Suite (2.57 Ga, initial $^{87}Sr/^{86}Sr = 0.704$, Hickman, 1978). However, as yet there is no evidence within the Limpopo belt of rocks depleted by partial melting (Hawkesworth et al., 1979).

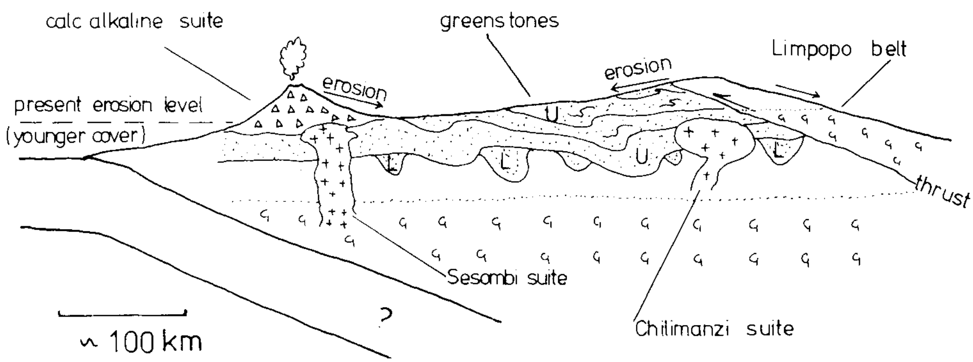

Fig. 7-6. Possible simplistic "plate-tectonic" reconstruction of 2.6 Ga events in the Rhodesian craton. Diagram shows old basement (2.9—3.5 Ga and older) overlain by deformed Lower Greenstone successions, and Upper Greenstone suites undergoing deformation. Uppermost units of the greenstone sequence include the calc-alkaline suite of the western margin of the craton. Intrusion of Sesombi granite suite was nearly contemporaneous and may have been from the same lower crustal-upper mantle parent melt. Intrusion of the Chilimanzi granite suite may have been from a relatively high-level source: in this diagram it is shown as a possible thermal consequence of the Limpopo thrust. Note that this very simple-minded "plate-tectonic" reconstruction, although possible, is unsatisfactory in that it is very speculative. It probably contains several anachronisms (allowed by the present poor state of knowledge) and makes assumptions about the structural history of the craton. Furthermore, the critical plate margin, if any, is now conveniently buried under younger cover (but see Cooper, 1978, for a suggestion that this cover was deposited in a deep trough, possibly over a fundamental weakness in the crust).

(6) *Throughout the craton and the Limpopo belt there is common evidence for regional compression (e.g. folds), tension (dyke swarms), and shear.* Away from the margin of the Limpopo belt the general form of the main greenstone belts is synclinal, although some belts are the remains of larger structures disrupted by invading granites. The relative importance of interference folding and diapiric uprise of basement granite-gneiss domes is still debated (e.g. Snowden and Bickle, 1976) but the present configuration of the belts seems best explained as the combined effects of regional folding, granite intrusion and erosion.

Early folding of the (? 2.9 Ga) Lower Greenstones was about east-northeast-trending axes. This was followed by the first phase of folding of the Upper Greenstones about north-northwest-trending axes, at least in the Belingwe and Lower Gwelo and possibly the Bulawayo areas. Intervening areas of granite and gneiss are often significantly less deformed by these deformations (Coward et al., 1976b) and extensive areas of c. 3.5 Ga old gneiss have survived internally undeformed since that time. Deformation at c. 3.5 Ga ago, however, included the recumbent folding of the Selukwe nappe and must have been in part responsible for the foliated greenstone remnants now preserved within the c. 3.5 Ga old gneisses. Similar thrusting and shearing on the margin of the Limpopo belt appear to postdate the deposition of the Upper Greenstones (2.7 Ga) and, locally, greenstone belts and gneisses are involved in major horizontal structures (Coward et al., 1976a).

It is possible, on present rather scattered evidence, that the Rhodesian craton reflects a sequence of tectonic events (3.5—2.9 Ga and 2.7—2.6 Ga) broadly similar to that of the Kaapvaal craton (3.5 Ga; 3.4—3.1 Ga; 3.0—2.8 Ga; 2.7 Ga). The sequence in the Limpopo belt may also be similar (*see also Barton and Key, this volume, Chapter 8, ed.*). Available evidence is probably still too unrepresentative of the whole area to make detailed correlations, and it is possible that further detailed work will imply very different histories in the two cratons, perhaps suggesting a plate boundary between them. Recognition, however, of narrow suture zones in highly deformed rocks in an incompletely exposed terrain is inevitably uncertain and probably only possible where adjacent rock masses are easily distinguished. Nevertheless, the balance of available evidence as outlined above suggests that there is no major suture between the Rhodesian and Kaapvaal cratons. Table 7-I summarizes available information about events in the two cratons and in the Limpopo belt.

To summarize, evidence of horizontal tectonics possibly indicative of convergent plate boundaries is restricted to the c. 3.5 Ga old greenstone remnants and to the c. 2.7—2.6 Ga tectonic events within the Limpopo belt. In neither case is there sufficient supporting evidence for volcanic or sedimentary rock types or metamorphic belts, or evidence for major tectonic discontinuities, to make a clear-cut case for a convergent plate boundary. As discussed above the 2.7 Ga old calc-alkaline volcanics form a belt apparently not related to the Limpopo belt. Relatively high-pressure metamorphic rocks (Chinner and Sweatman, 1968) preserved within the Limpopo belt must reflect a tectonically thickened crust, and rapid uplift would have been necessary to preserve such a low-temperature assemblage at that depth, even with modern continental thermal gradients (e.g. England, 1979).

TECTONIC MODELS

From the discussion above it is clear that at present there is insufficient evidence from the geological record either to support or refute "plate tectonics" in the modern sense as the dominant tectonic mechanism in the Archaean of Zimbabwe (Rhodesia). Perhaps the only points of significance to emerge are: (a) the predominance of the komatiite-tholeiite suite of volcanic rocks which might have been erupted during intra continental rifting; (b) the presence of very high temperature melts within the volcanic sequence; (c) a calc-alkaline "line" which might imply subduction of oceanic crust; (d) the preservation of areas of crust little deformed since 3.5 Ga; and (e) belts of intense deformation including thrusting. Given the marked lack of success of geologists in their attempts to infer the plate-tectonic mechanism from recent tectonic areas in the absence of critical geophysical observations, one might be justified in remaining sceptical of attempts to infer tectonic mechanisms responsible for the fragmentary and often poorly exposed Archaean remnants.

An alternative approach to the understanding of Archaean tectonics is to try to extrapolate models of modern tectonics to higher heat production within the earth. For example Bickle (1978) has shown that if plate tectonics did not exist in the Archaean, some other mechanism must have existed to manage the earth's thermal budget. It is difficult to conceive models for Archaean tectonics in which active areas of oceanic type were not responsible for the loss of the bulk of the earth's heat. The sort of model proposed by Bickle (1978), however, requires the assumption that the heat input into the base of the continental and oceanic lithosphere was approximately equal and was transported by small-scale convection cells (McKenzie and Weiss, 1975). If the earth's mantle convects through the 650 km discontinuity, or if the continental lithosphere were capable of insulating continents from this heat input by channelling convection towards oceanic areas, then the equality of oceanic and continental heat flow might not hold.

From the geological history of the Rhodesian Archaean craton, and indeed most other cratons, it is clear that these areas had a somewhat complex thermal history for the approximate time span 3.5—2.5 Ga ago. This history, including extrusion of voluminous high temperature lavas, massive granite intrusion and remelting of parts of the continental crust, in places repeated at intervals of only 100 or 200 Ma., would provide sufficient thermal "shocks" to keep a thermal boundary layer type lithosphere relatively thin. If the continental lithosphere were formed of more refractory and less dense depleted material (for example see Oxburgh and Parmentier, 1977) the chances of its own survival as well as that of its associated crustal component might be increased.

McKenzie and Weiss (1975) have made the only serious attempt to model mantle convection under higher heat flow conditions. They considered that modern plate tectonics involves two scales of flow in the mantle: a large-scale flow responsible for the surface plates, and a much smaller-scale convective flow within the domains defined by the plates. The form of circulation of the smaller-scale flow would depend on the magnitude of the large-scale shearing. In regions of strong shearing, the small-scale rolls would be aligned along the direction of shear; in regions of weak shearing a three-dimensional small-scale flow would be set up. McKenzie and Weiss (1975) further suggested that in the early part of the Precambrian temperatures below the thermal boundary layer near the earth's surface were perhaps 200°C hotter than today (an observation compatible with what is known of komatiite melting temperatures), and that the small-scale flow would have been very much more active in the Archaean than today. Possibly stresses generated by this flow would have been as much as ten times their present magnitude (perhaps enough to prevent the formation of major plates), with convective velocities as fast as 1 m/a. McKenzie and Weiss (1975) suggested that this model might very well account for the tectonic style of greenstone belt terrains but did not consider this in detail. A geological application of this model has been suggested by Williams (1977).

The average thickness of the Archaean continental lithosphere remains a controversial problem. The Archaean thermal gradients derived from granulite facies metamorphic belts are not easily reconciled with models involving a very thin continental lithosphere (Bickle, 1978; England, 1979). To manage the earth's thermal budget by a plate-tectonic system similar to that operating today would require a much more rapid production and consumption of plates than today. Bickle (1978) has discussed this problem and suggested that plate thickness would be only slightly less than at present. It is possible that plate thickness was substantially different under the continents and oceans, but no information is available as to the nature of oceanic lithosphere in the Archaean.

If small-scale convection cells dominated movement in the Archaean upper mantle, most heat loss would probably be by prolific volcanism. "Plate" motions would probably be on a rough scale of 700 km (the scale of the cells as discussed by McKenzie and Weiss, 1975). The dimensions of the Rhodesian Archaean craton (700 km × 300 km) make it difficult to test the hypothesis that small-scale plate tectonics operated, although the deformation patterns in the Limpopo belt are not inconsistent with smal-scale convection. Without areas of Archaean crust much larger than 10^3 km across it is difficult to deduce the maximum extent of tectonic zones.

CONCLUSIONS

What, then, is known about Archaean tectonics? It is evident that the Rhodesian craton has formed an extensive area of mostly rigid crust for at least 2.6 Ga and has probably been near rigid for the last 3.5 Ga. Some deformation has occurred; partial melting has taken place in some areas of this crust and extensively in the underlying mantle. But as far back as the early Archaean there is no evidence to show that the Rhodesian craton has suffered major physical disruption analogous to that associated with modern ocean floor evolution, whatever might be the case in Archaean terrains elsewhere. The Limpopo belt has had a fairly similar history to the craton and is probably best regarded as an uplifted equivalent of the craton. It is possible that the two units, perhaps together with the Kaapvaal craton, have formed an essentially rigid blocks for at least 3.5 Ga with only relatively minor jostling motions occurring between the cratons.

To the extent that such relatively large crustal masses as the cratons have been rigid and undisrupted for so long they are, in effect, plates or part of plates. Their deformational history, moreover, is such as to imply some kind of plate motion in the Archaean, possibly of limited extent on the basis of evidence from the late Archaean of Zimbabwe. What form the tectonic system took that produced the motion and shaped these plates is uncertain, but it was a plate system. Many features, e.g. the komatiites, imply that the system could not have been identical to modern plate tectonics but it may have been analogous.

Figure 7-6 is an attempt to explain the various features of the late Archaean of Zimbabwe (Rhodesia) in an extremely simplistic model comparable to those popular with Phanerozoic geologists. The model probably contains several anachronisms since it fully exploits the lack of resolution in age determinations. It is certainly not a model which would be suggested by the rocks themselves in the absence of "modern" preconceptions, but nevertheless it is a valid possible explanation of how the late Archaean geology of the craton could have been created. It must be re-emphasized that it is a very simplistic model while modern plate processes are often very complex.

As an alternative and very different system, the small-scale convective flows provide a very attractive way of driving relatively small fragments of crust. Figure 7-7 explores some of the possible consequences of vigorous small-scale convective motions which are strong enough to break the crust in places, or to extend or compress it. Most of the geological features shown in the craton and in the Limpopo belt can be explained in this way, but it is difficult to think of diagnostic criteria which would differentiate this style of tectonics from the simplistic "plate tectonics" discussed above. It is difficult to guess what plan form convective motions would have (Richter and Parsons, 1975). The size of the cratonic blocks in the early Archaean suggests plates of 300 km or more across: small-scale mantle convection cells would presumably have at least these dimensions and probably involved the whole upper mantle (to about 650 km depth). If the temperature distribution in the upper mantle were near-adiabatic (given high degrees of partial melt forming the komatiitic liquids) these cells would have Rayleigh numbers around 10^6. This would suggest rather turbulent processes; possibly the 3.5 Ga tectonics reflect this with large nappes and high-pressure metamorphic rocks. Later, "spoke-cell"-like patterns might be set up as the Rayleigh number decreased to about 10^5 (Richter and Parsons, 1976; Elder, 1977) leading to a somewhat more ordered and less vigorous crustal tectonic scheme, perhaps in late Archaean time.

To conclude: data currently available on the Archaean of Zimbabwe (Rhodesia) are inadequate to test tectonic models, either "plate-tectonic" or otherwise. A "small-scale convection" plate system is a valid alternative to the modern plate regime and should be seriously considered in any model of Archaean tectonics.

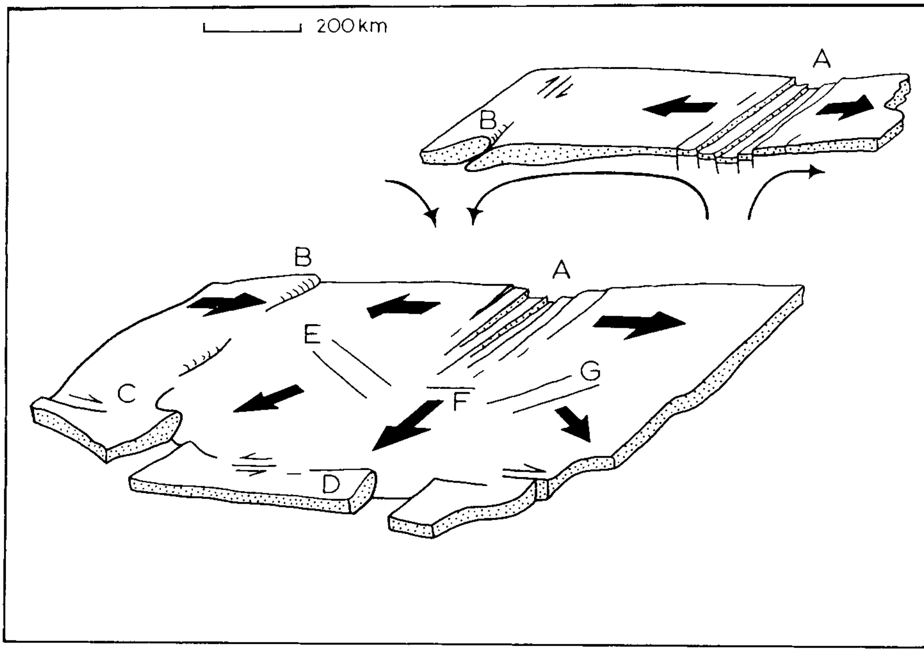

Fig. 7-7. Three-dimensional diagram to illustrate possible continental crustal response to small-scale convective motions in the mantle. Crustal extension takes place above upwellings to produce subsiding, linear, rift-fracture controlled basins (*A*). Fractures at *E*, *F*, and *G* could produce smaller similar basins in different orientations. Crustal thickening takes place above compressional downwellings (e.g. *B*, *C*, *D*). At *A* and perhaps to a lesser extent at *E*, *F* and *G* the deposition of thin basal sediments is likely as a first stage, followed by komatiite-tholeiite volcanism. A widespread continuous volcanic cover could result, thick in the basins and thinner elsewhere (cf. lower part of the Rhodesian craton Upper Greenstones). In the areas above downwellings underthrusting of crustal slabs could occur. From density constraints this would be of limited extent (e.g. *B*, *C*, *D*) and would die out laterally; a component of transcurrent movement is likely where there is opposing or different motion in the underlying convective systems (e.g. near *C* and *D* and cf. the Limpopo belt/ craton contact). Underthrusting could produce various different results. Partial melting could take place in both slabs. Release of volatiles from the lower slab would promote melting in the overlying crust of the upper slab and possibly in mantle material between the slabs. Products could include "crustally" derived granites (cf. Chilimanzi suite and the much older Mont d'Or granite), various felsic magmas (cf. western bimodal suite of the Upper Greenstones, and mantle derived magmas (cf. calc-alkaline volcanics and Sesombi tonalitic suite). If lower crust were returned relatively rapidly to the surface from depth, high-pressure, relatively low-temperature assemblages would occur (cf. Beitbridge area).
Note, though not illustrated here, crustal separation to varying degrees could also occur above upwellings (e.g. at *A*) with the formation of limited regions of "oceanic" crust which might in turn be consumed to produce late stage calc-alkaline igneous activity.

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The order of authors in this paper is more-or-less random, all three authors having ideas not adequately discussed here which will, with luck, surface in later contributions. This is a contribution from the Archaean Crustal Study, University of Zimbabwe (Rhodesia).

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Chapter 8

THE TECTONIC DEVELOPMENT OF THE LIMPOPO MOBILE BELT AND THE EVOLUTION OF THE ARCHAEAN CRATONS OF SOUTHERN AFRICA[1]

J. M. BARTON JR. and R. M. KEY

ABSTRACT

The tectonic development of the Limpopo Mobile Belt is reviewed and a plate-tectonic model is presented to explain this development and the evolution of the southern African cratons. It is postulated that the belt formed $\sim$ 3570 Ma ago with the establishment of a continental rise that collapsed into an aulacogen. The aulacogen and the sedimentary and igneous rocks that filled it were successively deformed as a result of differential movements between crustal plates encompassing what are now commonly termed the Rhodesian and Kaapvaal cratons. The timing of events in the evolution of the Limpopo Mobile Belt correlates well with the chronology of the tectonic development of the granite-greenstone terrains of the Rhodesian and Kaapvaal cratons. The pattern of crustal evolution in southern Africa may be viewed in terms of the accretion of a series of deformed back-arc basins and island arcs onto a nucleus of continental rocks including the Limpopo Mobile Belt.

INTRODUCTION

Various attempts have been made to interpret the evolution of Precambrian polymetamorphic terrains in terms of plate-tectonic models (see e.g. Anhaeusser, 1973; Burke and Dewey, 1973; Dewey and Burke, 1973; Hoffman, 1973, Talbot, 1973; Bridgwater et al., 1974; Hunter, 1974a, b; Tarney et al., 1976; Wynne-Edwards, 1976; Martin and Porada, 1977a, b; Key, 1977; Groves et al., 1978; Barton, 1979a). However, the degree to which any such interpretation can provide a unique solution that is meaningful in detail depends on how well the evolution of the specific terrain is understood. It is rare that both the chronology and the physical conditions during the tectonic evolution of a Precambrian polymetamorphic terrain are understood at all well, especially if the development began during Archaean times. The evolution of the Limpopo Mobile Belt of southern Africa is a notable exception. In this Chapter, the morphology and tectonic development of the Limpopo Mobile Belt as they are presently understood are described. The

[1] A publication of the South African Contribution to the International Geodynamics Project.

constraints that these observations place on plate-tectonic models are discussed and a tentative plate-tectonic model is presented for the evolution of the belt. This model involves the formation of an aulacogen which was periodically deformed over ~ 3500 Ma by differential movements between plates encompassing the Rhodesian and Kaapvaal cratons. While this plate-tectonic model may not provide the only conceivable explanation for the evolution of the belt it is, nevertheless, plausible and consistent with all the observed characteristics. Some of the implications of this model for the evolution of the continental crust of southern Africa are discussed.

THE LIMPOPO MOBILE BELT

Morphology and lithology

The Limpopo Mobile Belt (Mason, 1973) is a polymetamorphic terrain situated between the Rhodesian and Kaapvaal cratons of southern Africa (Figs. 8-1 and 8-2)(see also MacGregor, 1953; Cox et al., 1965). It may be conveniently divided into three zones, a Central Zone bordered more or less symmetrically by Marginal Zones (Cox et al., 1965; Mason, 1973). Although these divisions were originally proposed on the basis of analysis of aerial photographs, subsequent work has proved them to be valid in modified form. The Marginal Zones may be shown by similarity of rock compositions and continuity of certain rock units to be primarily comprised of reworked equivalents of the granite-greenstone terrains of the adjacent cratons (see e.g. Bennett, 1971; Mason, 1973; Robertson, 1973, 1974, 1977; Odell and Phaup, 1975; Key et al., 1976; du Toit and van Reenen, 1977; van Reenen and du Toit, 1977). They are separated from the compositionally distinct supracrustal rocks of the Central Zone by major fault zones (Figs. 8-2 and 8-3)(see e.g. Mason, 1973; Coward et al., 1973; Key and Hutton, 1976; Key, 1977). The contact of the Southern Marginal Zone with the Kaapvaal Craton is gradational with respect to both metamorphic grade and degree of deformation (Graham, 1974; du Toit and van Reenen, 1977; van Reenen and du Toit, 1977, 1978). The contact of the Northern Marginal Zone with the Rhodesian Craton, originally thought to be gradational (MacGregor, 1953; Robertson, 1968), is in part a zone of thrust faults dipping shallowly southward (James, 1975). However, in the west the original contact has been altered by repeated deformation and may have been at one time a zone of thrust faults dipping northward (Key, 1977).

The Marginal Zones are primarily comprised of leucocratic gneisses of variable but often tonalitic composition with small amounts of amphibolite, serpentinite, banded iron formation and other metasedimentary rocks preserved as synformal keels within the leucocratic gneisses. Conformable layers of K-feldspar porphyritic gneisses occur throughout the Northern Marginal

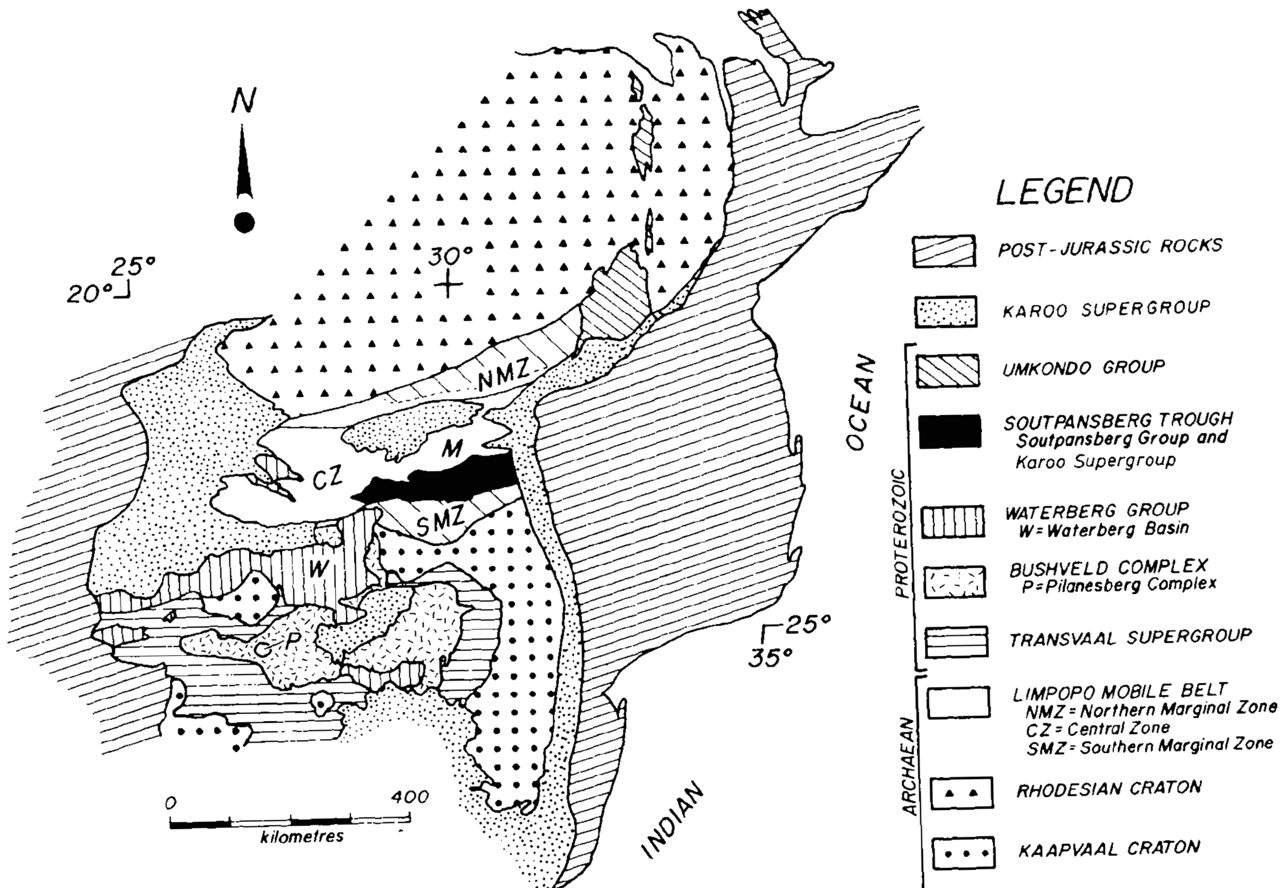
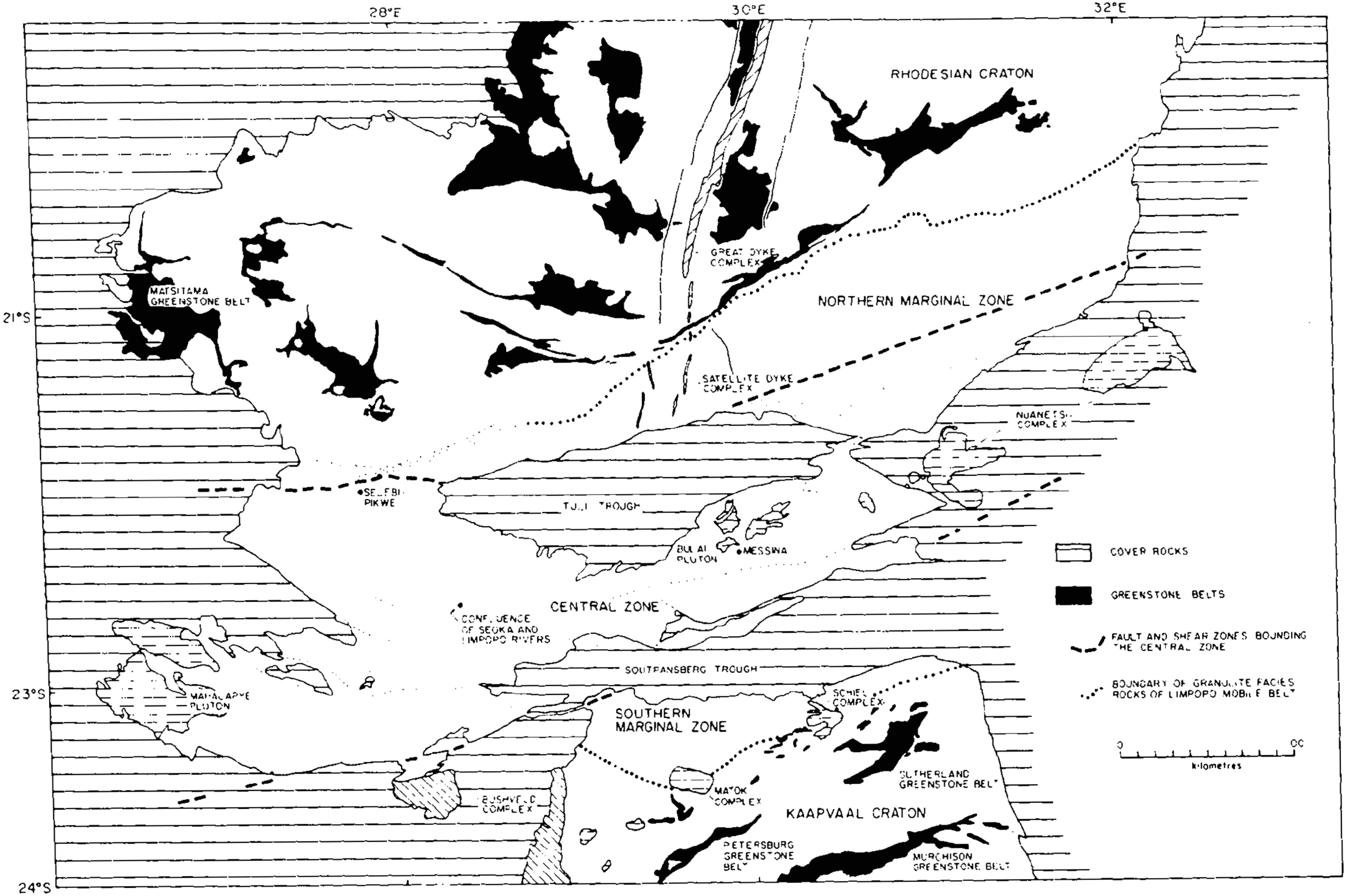


Fig. 8-1. Generalized geologic map of southern Africa showing the exposed portion of the Limpopo Mobile Belt and the surrounding area (modified from Barton (1979a, fig. 1). M = Messina area. Coarse diagonal striping pattern shows the areas in the Marginal Zones containing orthopyroxene-quartz bearing metamorphic mineral assemblages (modified from Coward, 1976, fig. 1).


28°E
30°E
32°E
21°S
23°S
24°S
RHODESIAN CRATON
MATSITAMA GREENSTONE BELT
GREAT DYKE COMPLEX
NORTHERN MARGINAL ZONE
SATELLITE DYKE COMPLEX
NUANETSI COMPLEX
SELEBI PIKWE
TULI TROUGH
BULAI PLUTON
MESSINA
CENTRAL ZONE
CONFLUENCE OF SEOKA AND LIMPOPO RIVERS
SOUTPANSBERG TROUGH
SOUTHERN MARGINAL ZONE
SCHIEL COMPLEX
MATOK COMPLEX
BUSHVELD COMPLEX
KAAPVAAL CRATON
PETERSBURG GREENSTONE BELT
MURCHISON GREENSTONE BELT
COVER ROCKS
GREENSTONE BELTS
FAULT AND SHEAR ZONES BOUNDING THE CENTRAL ZONE
BOUNDARY OF GRANULITE FACIES ROCKS OF LIMPOPO MOBILE BELT
kilometres

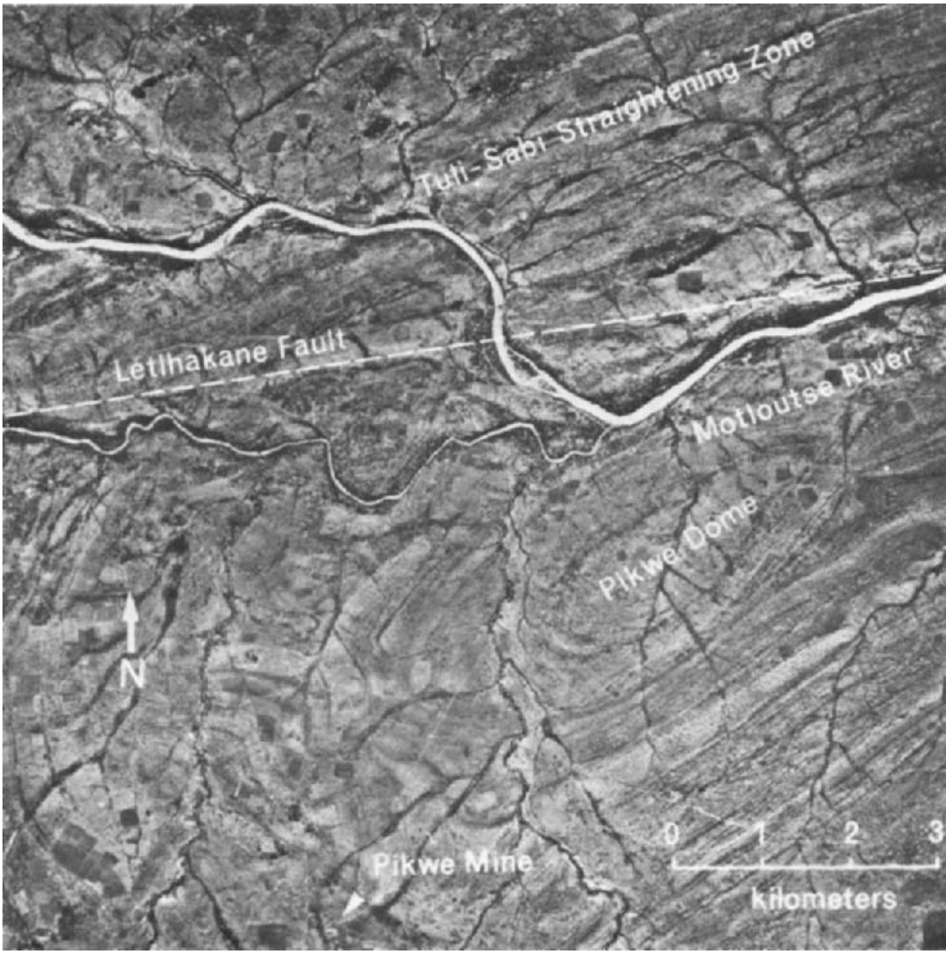


Fig. 8-3. An aerial photograph of a portion of the Limpopo Mobile Belt near Selebi-Pikwe, Botswana, showing part of the Central Zone of multiply deformed rocks (south of the Letlhakane fault) and the northern marginal fault zone of the Central Zone. The marginal fault zone here is called the Tuli-Sabi straightening zone and an important fault in it is called the Letlhakane fault.

Zone and these may be correlated with the units of arkosic sandstone that form the lower parts of the stratigraphic sequences within the greenstone belts in the southern part of the Rhodesian Craton (Key et al., 1976).

Fig. 8-2. Simplified geologic map of the Limpopo Mobile Belt and the adjacent cratons. The narrow horizontal striping pattern designates granitic and syenitic intrusions of diverse age. Except for the greenstone belts, the rocks of the cratons are undifferentiated but consist primarily of leucogranitic orthogneisses and granitoid plutons. The Marginal Zones consist of reworked rocks of the adjacent cratons. The Central Zone consists largely of a unique suite of supracrustal rocks, now present as gneisses.

Batholiths of syntectonic granitic rocks and thick sequences of volcanic rocks are largely lacking in the Marginal Zones. However, a line of plutons including the Matok and possibly the Schiel Complexes (Fig. 8-2) strikes obliquely into the Southern Marginal Zone from the Kaapvaal Craton (Barton et al., 1981a). Those in the Southern Marginal Zone are deformed.

The Central Zone near Messina (Fig. 8-2) is composed of a metamorphosed quartz-dioritic to granodioritic basement complex, including the Sand River gneisses (Barton et al., 1977, 1978, 1981c; Barton and Ryan, 1977; Fripp, 1980), unconformably overlain by a sequence of supracrustal rocks.[1] The basement complex contains a large component of probably metamorphosed greywacke (Barton, 1979b) intruded by gabbroic dykes (Fig. 8-4). The supracrustal rocks consist of metamorphosed sandstone, shale and carbonate rock with small amounts of banded iron formation, metamorphosed basalt and possibly rhyolite (see e.g. Söhnge, 1945; Söhnge et al., 1948; Jacobsen, 1967; Bahnemann, 1972; Mason, 1973; Light et al., 1977; Light and Watkeys, 1977; Fripp et al., 1979). The carbonate rock occurs primarily near the margins of the Central Zone, while the banded iron formation occurs most commonly near the centre, suggestive of sedimentary facies changes. Major differentiated sills of gabbroic rocks, emplaced at high crustal levels and collectively termed the Messina Layered Intrusion, are found within the supracrustal rocks and several generations of gabbroic dykes intrude the supracrustal rocks and those of the Messina Layered Intrusion (Barton et al., 1977, 1979a, 1981b). Deformed granitic bodies such as the Bulai pluton are unusual in this part of the Central Zone and are confined to the area west of Messina (Fig. 8-2)(Söhnge, 1945; Söhnge et al., 1948; Light et al., 1977; Light and Watkeys, 1977). A major body of undeformed granite, the Mahalapye pluton, occurs within the Central Zone in east-central Botswana (Fig. 8-2).

How far this succession of basement and cover rocks can be traced east-northeastward of Messina is unknown. To the west-southwest, however, compositionally similar rocks and rock successions occur as far as Baines

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[1] The basement complex, originally recognized by Bahnemann (1971), has been variously termed and includes the Artonvilla Formation (Jacobsen, 1967) and the Macuville Group in South Africa and Zimbabwe (see e.g. Light et al., 1977). The supracrustal rocks were originally termed the Messina Formation (Söhnge et al., 1948) and later in South Africa and Zimbabwe the Beit Bridge Group made up of several sub-groups and formations (see e.g. Light et al., 1977). Application of these divisions has proven to be nearly impossible owing to the difficulty of recognizing basement and cover rocks in outcrop and the large amount of deformation experienced by the rocks. In addition, the "basement rocks" exposed on Farm Macuville are actually in part supracrustal rocks and in part a phase of the Bulai pluton. Some of the units grouped by Söhnge et al. (1948) into the Messina Formation have been recognized as belonging to the basement complex. For these reasons, we have decided not to use formal stratigraphic names in this paper.

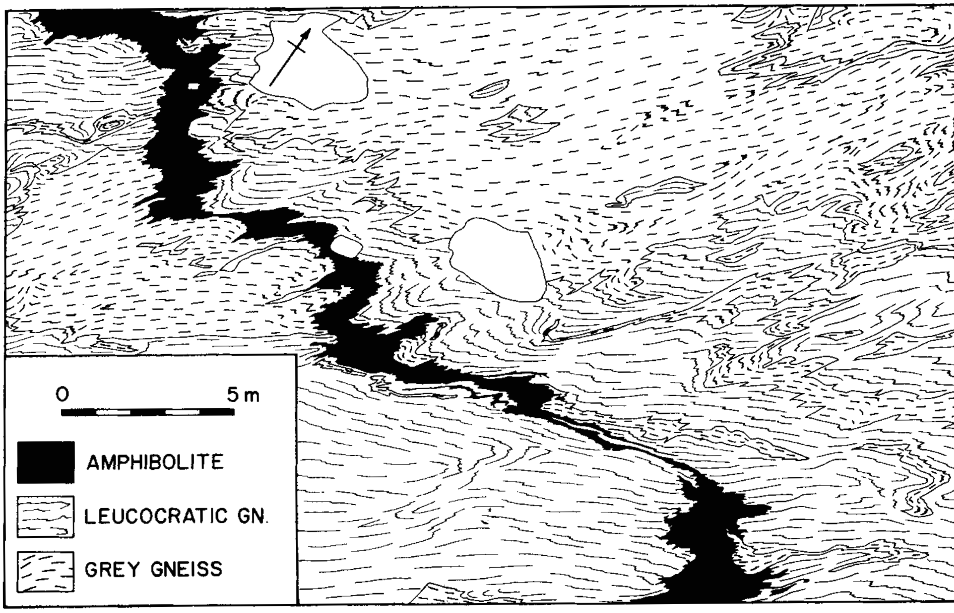

Fig. 8-4. A sketch map showing the relationship in the basement complex of the Limpopo Mobile Belt between the leucocratic and grey gneiss facies of the more than 3790 Ma old Sand River Gneisses and the approximately 3570 Ma old mafic dykes (modified from Barton et al., 1977, fig. 2).

Drift and the confluence of the Seoka and Limpopo rivers (Fig. 8-2; Key, 1977, 1979). Rocks compositionally similar to those of the supracrustal sequence also occur in the area around Selebi-Pikwe (Fig. 8-2; Wakefield, 1974, 1976), but large concentrations of metabasalt also occur here and the rocks that might be correlated with those of the Messina Layered Intrusion are anomalously rich in quartz (Key, 1977), possibly as a result of hydrothermal alteration. Near Mahalapye (Fig. 8-2) the stratigraphic succession appears to be quite different (Ermanovics, 1977), suggestive of a new and probably younger succession of rocks being exposed in the Central Zone west of approximately longitude 27°30′ E.

The fault zones bounding the Central Zone have had long and complex histories involving both strike-slip and dip-slip movements (see e.g. Mason, 1973; Coward et al., 1976a; Key and Hutton, 1976). The distance separating the fault zones and hence the width of the Central Zone decreases from west-southwest to east-northeast (Fig. 8-2), and the apparent amounts of displacement along these fault zones increase going in the same direction. In addition, both fault zones bend slightly to the north at a point in eastern Botswana (Figs. 8-2 and 8-5) and dip steeply to the south (Cox et al., 1965). The fault zones have been intermittently active over the history of the belt and they have controlled the pattern of more recent sedimentation in both the Tuli and Soutpansberg Troughs (Jansen, 1975; Key and Hutton, 1976; Ermanovics et al., 1978; Barton, 1979a).

Fault zones conjugate to those bordering the Central Zone and striking approximately from northwest to southeast exist throughout the Central Zone (Fig. 8-2). These have also been intermittently active during the history of the belt.

Gravity and magnetic character

The gravity pattern over the Limpopo Mobile Belt (Fig. 8-5) reveals two

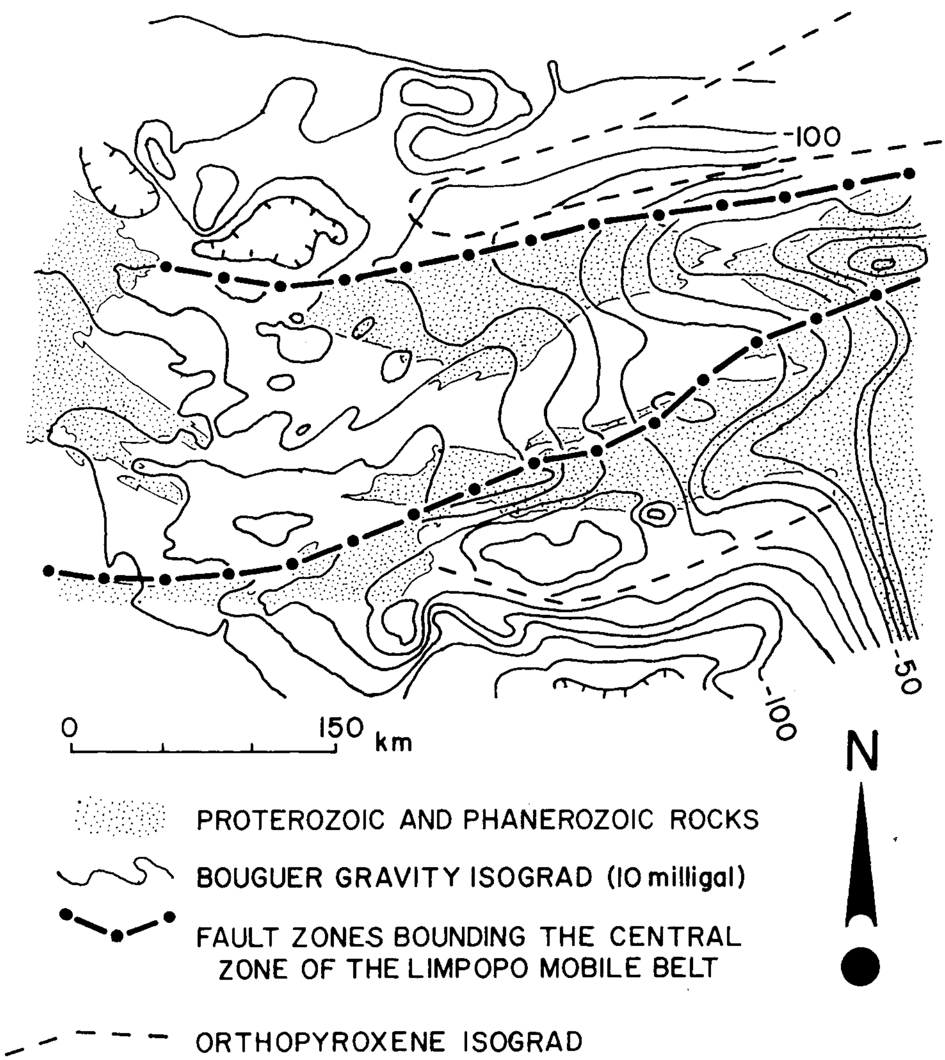

Fig. 8-5. Gravity pattern over the exposed portion of the Limpopo Mobile Belt (modified from Fairhead and Scovell, 1977, fig. 2).

broad lineaments of relatively high value, one along the northern margin of the Central Zone and the other along the northern margin of the Southern Marginal Zone (see e.g. Fairhead and Scovell, 1977; Reeves, 1977). In each instance the gravity high may be interpreted to indicate the existence of a thinner than normal, low-density upper crust in these areas and the presence nearer to the surface of higher-density rocks such as characterize the lower crust and upper mantle. The gravity highs become more subdued towards the west-southwest through the belt and disappear into a rather uniform but low pattern in eastern Botswana (Reeves and Hutchins, 1975). The gravity highs are continuous northeastwards as far as the gravity high associated with the Lebombo monocline which marks the eastern edge of the Kaapvaal Craton.

Although some of the gravity pattern in the eastern end of the belt may be attributable to the Lebombo monocline, it nonetheless suggests that a portion of the Central Zone has been thrust over the Northern Marginal

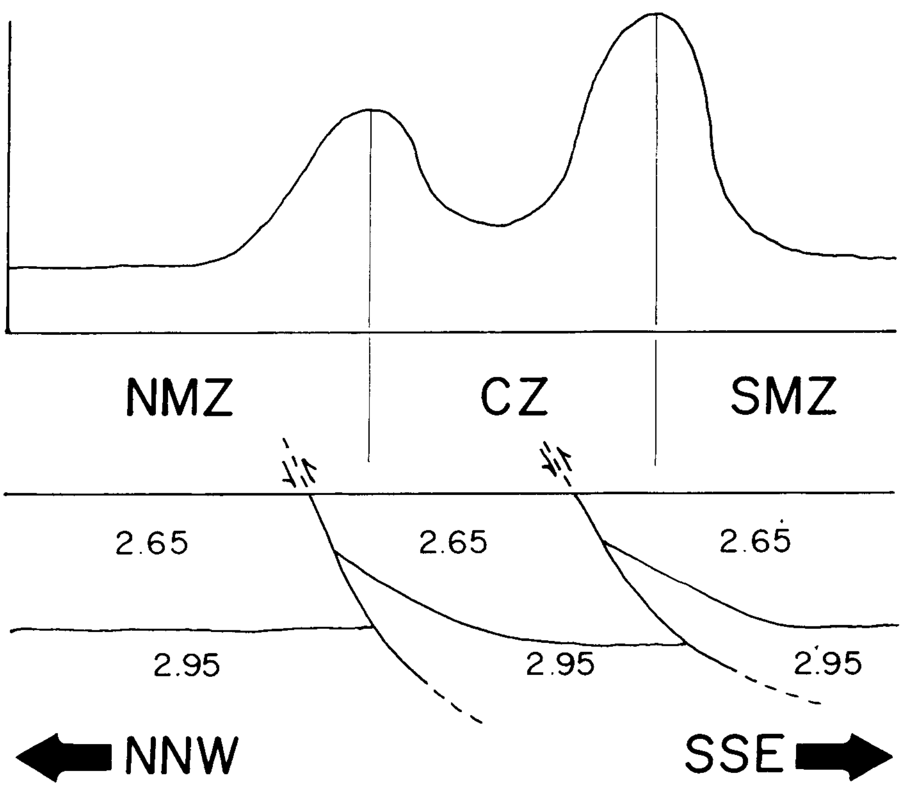

Fig. 8-6. A generalized crustal cross section of the eastern portion of the Limpopo Mobile Belt showing an average gravity profile for comparison. Assumed average rock densities are shown. *NMZ* = North Marginal Zone; *CZ* = Central Zone; *SMZ* = Southern Marginal Zone.

Zone and that a portion of the Southern Marginal Zone has been thrust over the Central Zone (Fig. 8-6). Field evidence for this thrusting may be found in Botswana (Key, 1977). This thrust faulting has resulted in the turning of the continental crust partially on edge in each case, exposing lower levels of crust along the northern margins respectively of the Central and Southern Marginal Zones. The amount of thrust faulting decreases to nothing towards the west-southwest and the convergence and bending of the fault zones bordering the Central Zone reflect the increasing amount of thrusting towards the east-northeast.

No gravity high marks the northern margin of the Northern Marginal Zone, suggesting that any thrust faulting along this margin must have either been along shallowly dipping zones or of a minor amount.

The magnetic and gravity patterns in Botswana (Reeves and Hutchins, 1975; Reeves, 1977, 1978a) show that the fault zone separating the Southern Marginal Zone from the Central Zone continues westward beneath the rocks of the Waterberg Group, Karoo Supergroup and Kalahari Group and is truncated by the Kalahari line (Reeves, 1977)(Fig. 8-7). The Kalahari line is a major crustal discontinuity that marks the western edge of the Kaapvaal and Rhodesian cratons. The fault zone separating the Central Zone from the Northern Marginal Zone becomes progressively less distinct westward and is unrecognizable west of about longitude 27°30′ E (Key and Hutton, 1976; Reeves, 1978a). This may reflect a termination of this fault zone, in which case the rocks of the Central Zone and those of the Rhodesian Craton are continuous between about longitude 27°30′ E and the edge of the Rhodesian Craton marked by the Kalahari line. Alternatively, and perhaps more reasonably, the gradual disappearance of the fault zone may reflect progressively deeper burial beneath younger rocks and this zone may actually continue at depth westward to the Kalahari line. The Kalahari line itself becomes less distinct to the north (Fig. 8-7) and is obscured by a feature termed the Makgadikgadi line (Reeves, 1977) which probably represents a fault zone along which younger rocks from the northwest are thrust southeastward over the rocks of the Rhodesian Craton. This thrust faulting did not distort the rocks containing the Kalahari line (Reeves, 1977). If this interpretation is correct, then the Central Zone of the Limpopo Mobile Belt completely separates the Rhodesian Craton from the Kaapvaal Craton and some of the rocks in the western portion of the central Zone, as suggested in the previous section, are younger than and overlie those in the east. Perhaps these younger supracrustal rocks are correlative with or constitute part of those in the basin postulated by Coward et al. (1976b) to have occupied the region immediately to the north of the Central Zone. The destruction of this basin gave rise to the upper greenstone belts in the southwestern portion of the Rhodesian Craton.

Metamorphism

Large areas of the Marginal Zones are composed of rocks bearing the min-

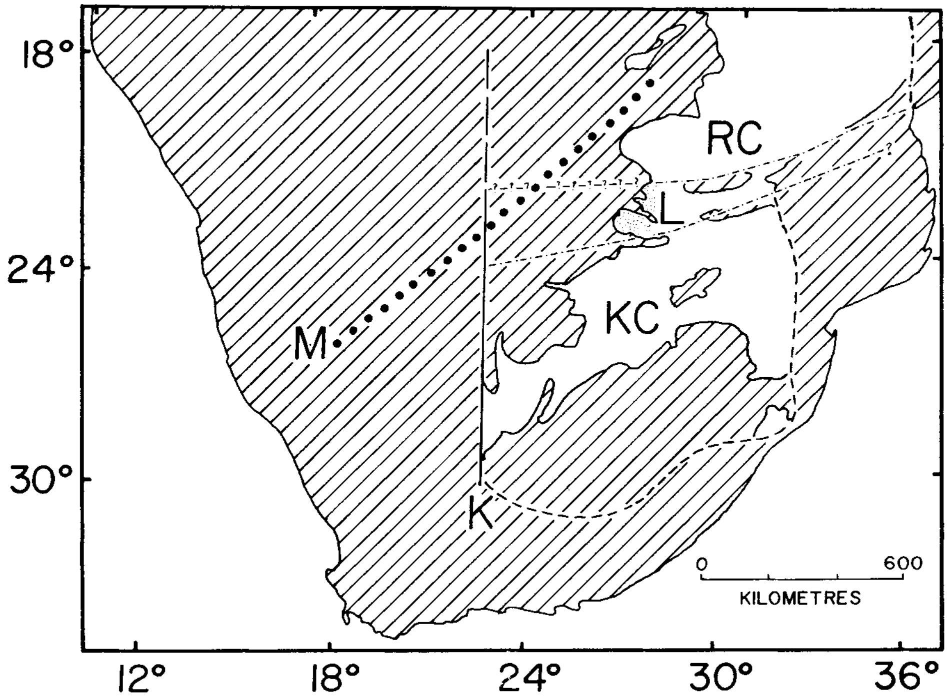

Fig. 8-7. A generalized tectonic map of southern Africa showing the inferred extent of the Rhodesian Craton (*RC*), the Central Zone of the Limpopo Mobile Belt (*L*) and the Kaapvaal Craton (*KC*). *K* = Kalahari line. *M* = Makgadikgadi line. The stippled area shows that portion of the western Central Zone in which presumed younger supracrustal rocks are exposed.

eral assemblage orthopyroxene-quartz (Fig. 8-1) and have been metamorphosed under pressure and temperature conditions characteristic of the granulite facies (Robertson, 1968; van Reenen and du Toit, 1977, 1978). The grade of metamorphism declines, in some cases over a fairly short distance (van Reenen and du Toit, 1977), passing into the adjacent cratons where mineral assemblages characteristic of lower amphibolite facies or greenschist facies prevail. The grade of metamorphism also decreases towards the west-southwest as does the width of the zone of granulite facies rocks (Figs. 8-1 and 8-2). Granulite facies rocks also occur in the zone of deformation in Botswana between the Matsitama greenstone belt and Pikwe (Fig. 8-2). In the Southern Marginal Zone the area containing the orthopyroxene-quartz assemblage may indicate the deepest level of the crust exposed.

In the Central Zone it is unusual to find orthopyroxene in quartz-bearing rocks, the rocks characteristically containing amphibole and biotite. However, petrologic studies suggest that these rocks also have been subjected to pressures and temperatures characteristic of granulite metamorphism

(Bahnemann, 1972; Clifford, 1974; Schreyer and Abraham, 1976; Horrocks, 1980). Rocks in large areas of the Central Zone may have been metamorphosed under conditions of higher water proportion in the fluid phase than have been those of the Marginal Zones. In the Central Zone, as with the Marginal Zones, the grade of metamorphism appears to become lower in the west, perhaps reflecting in that direction rocks from higher crustal levels at the surface.

Deformational history

Locally the deformational histories of the rocks of the Marginal Zones of the Limpopo Mobile Belt are reasonably well understood (see e.g. Coward et al., 1973, 1976a, b; Graham, 1974; Hickman and Wakefield, 1975; James, 1975; Coward, 1976; Key et al., 1976; du Toit and van Reenen, 1977). These histories may be correlated with those of the adjacent cratons, although the regional intensities of the deformational events are variable, accounting for the sometimes different tectonic styles of the Marginal Zones.

In the Northern Marginal Zone an early phase of recumbent folding preceded the formation of tight, upright north—south trending folds that are locally co-axially overfolded. This early folding formed the major nappes defined by the greenstone belts in the southern part of the Rhodesian Craton (Litherland, 1973) and was followed by a major period of shearing that formed the pronounced east-northeast-trending grain of the Northern Marginal Zone. A simple shear model with P_{max} parallel to the regional grain has been proposed for this second event (Key and Hutton, 1976). In addition, a rifting event preceded the intrusion of the Satellite Dykes of the Great Dyke Complex (Robertson and van Breemen, 1969). In the west, a young zone of deformation, spatially confined to a belt trending northwest to southeast between the Matsitama greenstone belt and Pikwe, intersects both the Rhodesian Craton and the Northern Marginal Zone and has affected both areas equally (Fig. 8-2).

In the Southern Marginal Zone a relatively simple tectonic history has been preserved. Here tight, upright east—west-trending folds have been refolded and cut by shears of thrust faults which also trend from east to west or from northeast to southwest (du Toit and van Reenen, 1977). These deformational events gave a pronounced east—west grain to the Southern Marginal Zone.

In the Central Zone the basement complex was penetratively deformed at least once prior to the deposition of the supracrustal rocks and was intruded subsequently by gabbroic dykes (Barton et al., 1977, 1979a). A basin then formed into which the sequence of supracrustal rocks was deposited, and this basin was subjected to tensional stresses during the emplacement of the Messina Layered Intrusion (Barton et al., 1979a). Following this, the rocks of the Central Zone around Messina were deformed under compressional stresses during four major events (Barton et al., 1979a). The first deformational event after deposition of the supracrustal rocks involved isoclinal folding and thrusting about approximately north—south-trending axial planes that dipped gently to the west. The second period of deformation which was the major fabric-forming event affecting both the supracrustal rocks and those of the Messina Layered Intrusion involved upright folding around steeply dipping northwesterly trending axial planes. Gabbroic dykes were intruded into these rocks after the second period of deformation. The Bulai pluton was emplaced after the second deformational event and was affected by and possibly intruded syntectonically during the third deformational event. This third event involved upright folding around axial planes nearly coplanar with those of the second deformational event. The fourth deformational event involved upright folding around east—west-trending axes and thrusting along east—west-striking axial planes. It was during this last major deformational event that the thrust faulting took place by which the rocks of the Central Zone over-rode those of the Northern Marginal Zone and the rocks of the Southern Marginal Zone over-rode those of the Central Zone.

For the rocks northwest of Messina in Zimbabwe a similar tectonic history has been proposed by Light and Watkeys (1977) except that they postulate the emplacement of the Messina Layered Intrusion synchronous with the late stages of the first deformational event affecting the supracrustal rocks.

Farther to the west again a roughly similar tectonic history has been proposed for the Central Zone but different workers have disagreed on points of detail. A possible basement complex, which has been recognized near the confluence of the Seoka and Limpopo Rivers (Fig. 8-2), may have been penetratively deformed and then intruded by gabbroic dykes prior to the deposition of the supracrustal rocks (Key, 1977). A basin then formed into which the supracrustal rocks were deposited, and these rocks in turn were intruded by the rocks of the Messina Layered Intrusion. The supracrustal sequence was tightly folded into recumbent structures with axial planes trending from north-northeast to south-southwest (Wakefield, 1977). A suite of gabbroic dykes was emplaced prior to a second deformational event that involved upright isoclinal folding about north—south-trending axes (Key, 1974; Wakefield, 1977). A third deformational event subsequently involved refolding of the upright folds about the same axial planes. All of the resulting interference structures plunge to the west and the basement complex and the supracrustal rocks appear to pass beneath a younger sequence of supracrustal rocks at about longitude 27°30′ E. Thrust and strike-slip faulting accompanied and followed the generation of the fold structures. Structures associated with the last deformational event near Messina have not been recognized in the rocks near Pikwe.

Chronology[1]

The rocks in the Marginal Zones of the Limpopo Mobile Belt, being metamorphosed equivalents of the granite-greenstone terrains of the adjacent cratons, are older than ~ 3200 Ma (see e.g. Allsopp, 1961; Oosthuyzen, 1970; Hickman, 1974, 1976, 1978; Hawkesworth et al., 1975; Davies and Allsopp, 1976; Key et al., 1976; Moorbath et al., 1976; Barton, 1980). However, ages only as old as ~ 2870 Ma have been reported for rocks from the Northern Marginal Zone (Hickman, 1976, 1978) and ages only as old as ~ 2650 Ma have been measured for rocks from the Southern Marginal Zone (Barton and Ryan, 1977; Barton et al., 1981a). In the Southern Marginal Zone nearly all of the deformed rock units have yielded ages of ~ 2600 Ma and the post-tectonic units yield slightly younger ages. It appears probable, therefore, that the metamorphism that affected the Marginal Zones was sufficiently severe to erase at least the Rb-Sr whole-rock isotopic evidence for these older ages. The rocks of the Marginal Zones were eroded to approximately their present level by ~ 1950 Ma ago (Barton and Ryan, 1977; Jansen, 1977), and post-kinematic gabbroic dykes were emplaced in the Southern Marginal Zone ~ 1900 Ma ago (Barton, 1979a).

In marked contrast to the Marginal Zones where metamorphism has obliterated the radiometric record of early tectonic events, a long and complex isotopic history covering c. 3800 Ma is preserved in the rocks of the Central Zone near Messina and probably extending as far west as Selebi-Pikwe (Barton and Ryan, 1977). The basement complex is at least ~ 3790 Ma old and was intruded by gabbroic dykes ~ 3570 Ma ago after suffering at least one deformational event (Barton and Ryan, 1977; Barton et al., 1977, 1978, 1981c). The supracrustal rocks were deposited sometime during the interval between ~ 3570 Ma ago and ~ 3270 Ma ago (Barton and Ryan, 1977; Barton et al., 1977, 1979a). The Messina Layered Intrusion and other related rocks were intruded ~ 3270 Ma ago and, near Messina, the first deformation of the supracrustal rocks occurred sometime between that time and ~ 3150 Ma ago (Barton et al., 1979a, 1981b). The second deformational event occurred ~ 3150 Ma ago and the Bulai pluton was probably emplaced ~ 2700 Ma ago (van Breemen and Dodson, 1972; Barton et al., 1979b). The third deformational event occurred between ~ 2700 Ma ago and ~ 2600 Ma ago and the fourth event occurred ~ 2600 Ma ago (Barton and Ryan, 1977; Barton et al., 1979a, 1981a). A metamorphic event for which no period of penetrative deformation has been assigned occurred ~ 3000 ± 50 Ma ago (Barton and Ryan, 1977; Barton et al., 1981b). This event was, however, accompanied by the emplacement of gabbroic dykes. High level granitic intrusions such as the Mahalapye pluton and some gabbroic dykes were

[1] All radiometric ages mentioned in this section have been calculated with the decay constants recommended by Steiger and Jäger (1977).

emplaced ~ 2200 Ma ago (van Breemen and Dodson, 1972; Barton, 1979a) and uplift and erosion to nearly the present surface level was achieved throughout most of the Central Zone by ~ 1950 Ma ago (Barton and Ryan, 1977). Local thermal metamorphism and uplift preceded the formation of the Soutpansberg Trough ~ 1770 Ma ago (Barton, 1979a) and some movement along the fault zones bounding the Central Zone accompanied the deposition of the rocks of the Karoo Supergroup in the Soutpansberg and Tuli Troughs ~ 180 Ma ago (Manton, 1968; Barton and Ryan, 1977). In general, every tectonic event that has occurred on the Rhodesian and Kaapvaal cratons has been accompanied by some contemporaneous tectonic activity in the Central Zone of the Limpopo Mobile Belt (see the review in Barton and Ryan, 1977).

A PLATE-TECTONIC MODEL

In order to properly apply plate-tectonic models to a polymetamorphic terrain such as the Limpopo Mobile Belt, it is necessary to dissect the evolution of the terrain into distinct tectonic events. These events must be evaluated separately and the final model must be the sum of the separate solutions. It must be remembered that besides the crustal plates themselves, plate-tectonic models involve spreading zones of crustal creation, shortening zones of crustal destruction and zones of predominantly strike-slip motion with minor crustal distortion. These last two zones form major crustal discontinuities that tend to act as permanent zones of weakness that may be preferentially reactivated in response to later crustal stresses.

There is no conclusive evidence that oceanic crust ever existed in what is now the Central Zone of the Limpopo Mobile Belt. Ophiolites have not been recognized and large volumes of volcanic and intrusive rocks of the types that result from destruction of oceanic crust are conspicuously lacking except possibly for the local concentration of metavolcanic rocks west of Pikwe. Therefore, classic (i.e. post-Mesozoic type) subduction of oceanic crust in the evolution of the Central Zone of the Limpopo Mobile Belt seems unlikely unless it was achieved without subsequent magmatism. Furthermore, the lack of large volumes of volcanic and intrusive rocks precludes the possibility that a mantle plume is incorporated into the Central Zone.

The earliest tectonic event that may be considered unequivocally to be a unique part of the evolution of the Limpopo Mobile Belt is the development of a basin, *floored by continental crust*, into which the supracrustal rocks of the Central Zone were deposited sometime between ~ 3570 Ma and ~ 3270 Ma ago. The basement complex, albeit older, is compositionally similar to the Archaean gneiss terrains on both the Rhodesian and Kaapvaal Cratons, and it may represent a lower level within the oldest parts of these cratons (Key et al., 1976). The intrusion of gabbroic dykes ~ 3570 Ma ago into the basement complex is apparently coeval with mafic magmatism of the Barberton Mountain Land and the emplacement of the Sebakwian volcanic rocks of the Rhodesian Craton (see e.g. Jahn and Shih, 1974; Barton et al., 1977; Wilson et al., 1978; Hamilton et al., 1979; *also Anhaeusser, this volume, Chapter 6; Nisbet et al., this volume, Chapter 7, ed.*). The sequence of supracrustal rocks, being composed primarily of clastic rocks of a continental origin and containing few rocks of volcanogenic or chemical provenance, is unique in composition and time of deposition to the belt. The only other sedimentary rocks in southern Africa recognized to possibly be coeval with it are those of the Fig Tree and Moodies Groups of the Barberton Mountain Land (Barton et al., 1977) and the Tutume Group of the western margin of the Rhodesian Craton (Litherland, 1973). However, these sedimentary rocks are derived primarily from volcanic source areas. The linear occurrence of carbonate rocks near the present margins of the Central Zone and the apparent lack of supracrustal rocks related to those of the Central Zone in the Marginal Zones (Key, 1977) suggest that the margins of the basin were not much different than those of the Central Zone today. The obvious plate tectonic analogue for such a basin is an aulacogen or a continental rift valley formed by the collapse of a continental rise emanating from a triple junction (Burke and Dewey, 1973). The fact that the fault zones bounding the Central Zone continue completely across the Rhodesian and Kaapvaal Cratons suggests that the triple junction existed outside of the belt at one end or the other.

The second tectonic event of widespread significance in the development of the Limpopo Mobile Belt was the emplacement of the Messina Layered Intrusion into the supracrustal rocks ~ 3270 Ma ago. This event may or may not have closely followed the actual deposition of the supracrustal rocks. However, the tensional regime during which it was intruded and the high initial $^{87}Sr/^{86}Sr$ ratio (0.7029, corrected for 3270 Ma; see Barton et al., 1979a) of the magma giving rise to the Messina Layered Intrusion when compared to its age, are consistent with it having been emplaced during continental rifting or the formation of an aulacogen (Barton, 1979a; Barton et al., 1979a).

The first three deformational events in the Central Zone affecting the rocks of the supracrustal sequence and the Messina Layered Intrusion are considered to be the result of Riedel shear across the belt resulting from differential movements of the adjacent cratons or plates (see e.g. Coward, 1976; Coward et al., 1976a, b; Key and Hutton, 1976). This means that there have been large components of roughly parallel but opposite motions between the plates encompassing the Rhodesian and Kaapvaal Cratons and that the rocks of the Central Zone have acted as a sort of shock absorber between the plates, deforming in response to externally applied stresses. The first deformational event, occurring sometime between ~ 3270 Ma and ~ 3150 Ma ago, involved refolding about and thrust faulting along approximately north—south-trending axial planes and resulted primarily from right-lateral motion between the plates. The second deformational event, occurring ~ 3150 Ma ago, and the third deformational event, occurring sometime between $\sim$ 2700 Ma and $\sim$ 2600 Ma ago, involved refolding of the north—south structures about, and thrust faulting along, approximately northwest–southeast-trending axial planes and resulted from primarily left-lateral motion between the plates. It is improbable that in these three instances the motions of the plates were exactly parallel. Consequently, there was almost certainly a component of crustal shortening or extension occurring across the Central Zone during each of these deformational events (see e.g. Coward, 1976; Coward et al., 1976a, b). Components of crustal shortening or extension are responsible for the parallelism in the deformational histories of the Central and Marginal Zones (see e.g. Key and Hutton, 1976; Key, 1977).

The fourth deformational event occurred $\sim$ 2600 Ma ago and was caused by a convergence of the plates across the eastern end of the belt, essentially perpendicular to the Central Zone. This resulted in the thrusting of rocks of the Central Zone over those of the Northern Marginal Zone and of the rocks of the Southern Marginal Zone over those of the Central Zone. This thrust faulting occurred along the pre-existing fault zones bordering the Central Zone. Because, however, the thrust faulting dies out to the west-southwest, a clockwise rotation of the plate including the Rhodesian Craton with respect to that including the Kaapvaal Craton of $\sim 15^\circ$ is deduced, apparently occurring around the portion of the belt now exposed in eastern Botswana. This rotation and resulting thrust faulting caused the slight divergence of earlier fold axes and thrust faults in the Messina area with respect to those in the Selebi-Pikwe area. It also explains why the last deformation in the Messina area is not manifested in the rocks of the Selebi-Pikwe area. Part of this rotation may have caused or resulted from the opening of the rift into which the Great Dyke Complex was intruded and the crustal thinning during the formation of the Witwatersrand Basin. It may also have caused or resulted from the deformation along northwest—southeast-trending axes between the Matsitama greenstone belt and Pikwe. It may even have resulted in the fracturing of the southwestern part of the Rhodesian Craton along west-northwest axes. These fractures were reactivated during late Karoo times (Reeves, 1978b). The rotation produced, however, both the convergence and bending of the fault zones bordering the Central Zone and a general tilting of the rocks of the Central and Southern Marginal Zones to the south and to the west. In both cases the crust was turned slightly on edge and deeper levels were exposed by subsequent erosion going northwards. In the Central Zone the basement complex was exposed. In the Southern Marginal Zone the orthopyroxene-quartz bearing granulites were exposed. The blanket $\sim$ 2600 Ma whole-rock ages from these rocks reflect their passing during thrust faulting into conditions of pressure and temperature where the radiometric clocks could start. Minor thrust faulting probably occurred in the Northern Marginal Zone to accommodate some crustal shortening and this faulting may have brought rocks of granulite grade to the surface.

It is important to note that even though the Limpopo Mobile Belt may appear surficially symmetrical with regard to the Central and Marginal Zones and even though it may have been symmetrical before the last deformational event, this event imposed a major structural asymmetry on the belt.

The Rhodesian and Kaapvaal cratons as well as the Limpopo Mobile Belt have acted as a reasonably stable plate since ~ 2600 Ma ago and have formed a major structural domain termed by Clifford (1966) the "Kalahari Shield". A cartoon depicting the plate-tectonic model presented here for the evolution of the Limpopo Mobile Belt is shown in Fig. 8-8 and a summary of the tectonic development is presented in Table 8-I.

IMPLICATIONS FOR CRUSTAL EVOLUTION OF SOUTHERN AFRICA

Plate tectonics as a mechanism implies a global interaction in tectonic activity. The pattern of activity in any one location should not be viewed

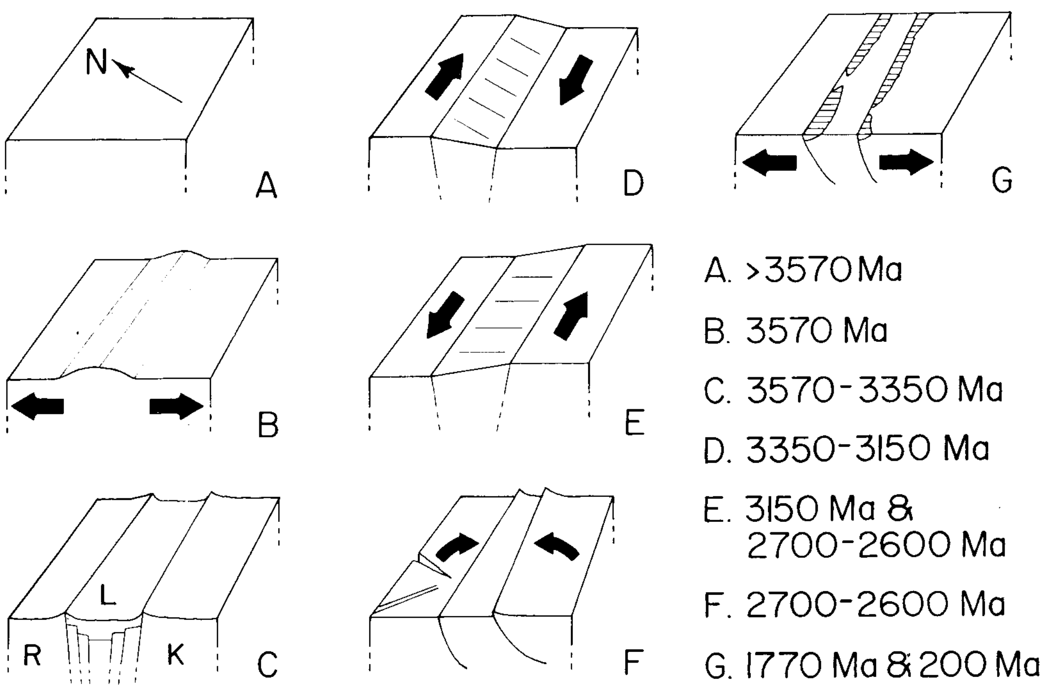

Fig. 8-8. A cartoon showing schematic block diagrams depicting the major tectonic events in the evolution of the Limpopo Mobile Belt (see also Table 8-I). N = north; R = Rhodesian Craton; L = Central Zone of the Limpopo Mobile Belt; K = Kaapvaal Craton. Horizontal pattern in Fig. 8-8G shows the rocks of the Karoo Supergroup and the Soutpansberg Group. The parallel lines in Figs. 8-8D and 8-8E show the primary axes of folding and thrust faulting during the first three deformational events affecting the belt. Solid arrows denote directions of relative motion.

TABLE 8-I

The tectonic development of the Limpopo Mobile Belt

Age*	Description
± 180 Ma	Reactivation of the Tuli and Soutpansberg troughs and the deposition of the rocks of the Karoo Supergroup (Fig. 8-8G). Intrusion of mafic dykes (Reeves, 1978b)
± 730 Ma	Emplacement of the Beit Bridge kimberlite pipes (Allsopp and Kramers, 1977)
± 1770 Ma	Formation of the Tuli and Soutpansberg troughs and deposition of the rocks of the Soutpansberg Group (Fig. 8-8G)
± 1900 Ma	Deposition of the rocks of the Waterberg Group in the western portion of the belt and the emplacement of postkinematic gabbroic dykes in the Central and Marginal Zones
± 1950 Ma	Establishment of ubiquitous mineral radiometric ages possibly related to the termination of uplift and erosion of the entire region
± 2220 Ma	Emplacement of post-kinematic gabbroic dykes in the Central Zone and intrusion of the Mahalapye batholith
± 2550 Ma	Emplacement of post-kinematic granitic plutons in the Southern Marginal Zone and of the Satellite Dykes of the Great Dyke Complex into the Northern Marginal Zone
2600—1950 Ma	Uplift and erosion to nearly the present surface level, the resulting detritus filling the contemporaneously existing basins on the Rhodesian and Kaapvaal cratons. Possibly some deformation occurring in the Botswana segment of the belt
2700—2600 Ma	The third period of deformation affecting the belt resulting from a left lateral motion between the Rhodesian and Kaapvaal cratons (Fig. 8-8E). The fourth period of deformation affecting the belt resulting from a clockwise rotation of the Rhodesian Craton with respect to the Kaapvaal Craton (Fig. 8-8F)
± 2700 Ma	Emplacement of the Bulai pluton in the Central Zone, perhaps syntectonically with the third period of deformation affecting the belt
3050—2850 Ma	Metamorphic event of unknown significance and emplacement of mafic dykes
± 3150 Ma	The second period of deformation and major fabric forming event in the supracrustal rocks. This deformation resulted from left lateral motion between the Rhodesian and Kaapvaal cratons (Fig. 8-8E)
3270—3150 Ma	The first period of deformation affecting the belt resulting from a right lateral motion between the Rhodesian and Kaapvaal cratons (Fig. 8-8D)
± 3270 Ma	Emplacement of the Messina Layered Intrusion in the Central Zone
3570—3350 Ma	Formation of a fault bounded basin and emplacement of the sequence of supracrustal rocks characteristic of the Central Zone (Fig. 8-8C)
± 3570 Ma	Emplacement of gabbroic dykes in the basement complex, possibly as

Table 8-I continued

Age*	Description
	a result of formation of a continental rise in the Rhodesian—Kaapvaal Craton originating from a mantle plume (Figs. 8-8A and 8-8B). *The establishment of the Limpopo Mobile Belt*
± 3790 Ma	A major period of deformation of the Rhodesian—Kaapvaal Craton
> 3790 Ma	Deposition of a sequence of greywacke as a result of erosion of volcanic terrain. This volcanic terrain contained a large component of rocks formed by partial melting of the upper mantle and oceanic crust

* Based on the decay constants recommended in Steiger and Jäger (1977).

singly but instead as part of a total earth pattern. Unfortunately, with the present state of knowledge, this is difficult to recognize in detail for the past 200 Ma of earth history and is nearly impossible to visualize for Precambrian time except in the very broadest terms. Nevertheless, if the model presented here is correct, it may provide some clues to the nature of the development of the Rhodesian and Kaapvaal cratons and some evidence about the style of tectonic activity during Archaean time.

In order to form an aulacogen strong, reasonably thick continental crust must exist. Therefore, it may be inferred that the Limpopo Mobile Belt formed in a continental nucleus containing rocks at least ~ 3800 Ma old. The formation of the aulacogen resulted in the creation of two plates each containing some of the old continental crust. These remnants are now probably incorporated in the Rhodesian and Kaapvaal cratons although they have yet to be identified.

The stresses that caused the deformation of the aulacogen were generated elsewhere within or adjacent to the plates encompassing the Rhodesian and Kaapvaal cratons. It is therefore useful to look at the tectonic activity on these cratons that accompanied each deformational event in the Central Zone of the Limpopo Mobile Belt. The postulated formation of the triple junction and the consequent generation of a continental rise followed by an aulacogen or graben occurred at approximately the same time as the formation of the early granite-greenstone terrains of both cratons (see e.g. Hawkesworth et al., 1977; Wilson et al., 1978; Hamilton et al., 1979; Barton, 1980). Most of the greenstones contain mafic and ultramafic rocks that are believed to represent oceanic crust although in order to explain the formation of high-magnesium basaltic and peridotitic magmas a close genetic relationship between these rocks and sialic rocks of a continental crust must be inferred (see e.g. Anhaeusser, 1973; Hunter, 1974a; McIver, 1975; Key et al., 1976; Hawkesworth and O'Nions, 1977; Groves et al., 1978). This suggests the probability that greenstone belts are actually remnants of oceanic crust formed during advanced rifting of continental crust (Hunter, 1974a; Groves et al., 1978; *see also Goodwin, this volume, Chapter 5, Kröner, this volume, Chapter 3, and Lambert, this volume, Chapter 18, ed.*) or in back-arc basins between a continent and an island arc situated on either oceanic or continental crust (Tarney et al., 1976). As such the Central Zone of the Limpopo Mobile Belt may have originally formed in much the same manner as did the greenstone belts, with the only difference that rifting ceased in the belt before oceanic crust was emplaced.

This does not mean, however, that the evolution of the Limpopo Mobile Belt and of any given greenstone belt on the Rhodesian and Kaapvaal cratons were exactly coeval in every aspect. The only greenstone belt from which sufficient age data are available for comparison is the Barberton greenstone belt of the Kaapvaal Craton (see the synthesis of the tectonic evolution in Barton, 1980 and, for contrast, the discussion by Anhaeusser, this volume, Chapter 6). Here an oceanic crust composed of rocks of the Lower Onverwacht Group was formed ~ 3550 Ma ago (Hamilton et al., 1979) in what we suggest to have been a back-arc environment between an island arc and the continental nucleus ultimately to contain the Limpopo Mobile Belt. The basin containing the oceanic crust was first filled with predominantly volcanic rocks (the Upper Onverwacht Group) and then by sediments of the $> \sim 3450$ Ma old Fig Tree and Moodies Groups (Barton, 1980). These sedimentary units were deposited in a narrow continental shelf-type environment as might exist along the margins of a fault bounded (rift) basin and they were derived from a rapidly eroding source area composed of both volcanic and continental rocks (Condie et al., 1970; Eriksson, 1979, 1980). The amount of continental detritus increases upwards through the Fig Tree and Moodies Groups, suggesting that the source area was composed of continental crust overlain by volcanic rocks. This source area was located to the south and, in a plate-tectonic context, is a good candidate for a volcanic terrain (island arc) situated on continental crust and derived from a subducting piece of oceanic crust. The basin was then compressed and intruded by plutons derived from the partial melting of oceanic crust similar to that flooring the basin. During this deformational event the rocks of the basin were welded onto the continental nucleus. The elapsed time from the formation of the basin until welding of the basin rocks onto the continental nucleus was probably less than ~ 100 (Barton, 1980). Subsequently, the deformed basin was intruded by plutons during three periods:, the first $\sim 3250 \pm 150$ Ma ago, the second $\sim 2900 \pm 150$ Ma ago and the third $\sim 2550 \pm 50$ Ma ago. In each instance the plutons were either derived directly from partial melting of oceanic crust or they represent remobilized material derived originally from partial melting of oceanic crust. The rocks exposed in the Barberton greenstone belt were never deeply buried and have been at nearly their present crustal level for the past ~ 3000 Ma.

The formation of the Sebakwian greenstone belts of the Rhodesian Craton also occurred before ~ 3500 Ma ago (Moorbath et al., 1976; Wilson et al., 1978; Wilson, 1979) but the detailed tectonic evolution of these rocks is poorly understood. However, the younger greenstone belts of the Rhodesian Craton appear to have formed in a similar fashion to the Barberton greenstone belt, and the span of time from the formation of a basin floored by oceanic crust until the destruction of that basin was only ~ 100 Ma from ~ 2700 Ma until ~ 2600 Ma ago (see e.g. Bickle et al., 1975; Hawkesworth et al., 1975; Coward et al., 1976a, b). In addition, igneous and/or metamorphic activity, possibly associated with the formation of a granite-greenstone terrain, occurred in the Rhodesian Craton ~ 2850 Ma ago (Hawkesworth et al., 1979).

The periods of destruction of these Archaean basins and the resulting formation of the greenstone belts coincide remarkably well with the periods of tectonic activity in the Central Zone of the Limpopo Mobile Belt, although the basin that ultimately became the Barberton greenstone belt may possibly have formed at a slightly earlier time than that of the Central Zone, and the basin forming the Central Zone may have been created during the first deformation of the Barberton greenstone belt. This similar chronology suggests that the sialic crust of southern Africa was being stretched intermittently to form basins from ~ 3550 Ma until ~ 2550 Ma ago. If the basins into which the rocks of the Witwatersrand triad, the Transvaal Supergroup, the Bushveld Complex and the Waterberg Group were deposited formed in a similar fashion, then this stretching continued until ~ 1850 Ma ago. However, the early basins tended to spread until the lithosphere fractured and oceanic crust was emplaced in them while the younger ones remained floored by continental crust, albeit thinner. In addition, the early basins were unstable and subject to destruction while the later ones were more permanent. The destruction of any given basin may have been a result of spreading to form other basins.

CONCLUSION

The Archaean tectonic evolution of southern Africa may be viewed adequately in terms of intermittent plate-tectonic activity. The Limpopo Mobile Belt formed as an aulacogen in a nucleus of continental crust onto which deformed back arc basins were accreted in the form of granite-greenstone terrains. However, as most tectonic features such as the Limpopo Mobile Belt and the greenstone belts are only partially preserved and imperfectly exposed, it is difficult to evaluate the time scale involved in each tectonic event and the arial extent and thickness of the plates involved. Therefore, in reality one can only speculate as to whether application of uniformitarian processes is appropriate or whether some other paradigm involving gradual evolution of processes might fit the observed features better.

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Chapter 9

ARCHAEAN BASIN EVOLUTION IN THE SLAVE PROVINCE, CANADA

JOHN B. HENDERSON

ABSTRACT

The Slave Province in the northwestern part of the Canadian Shield is a sediment-dominated Archaean "granite-greenstone" terrane underlying an area of about 190,000 km^2. Archaean supracrustal rocks were deposited in a 10 to 15 million year period 2670 Ma ago in a series of small fault-bounded basins that formed due to regional extension of the c. 3 Ga old granitic to tonalitic basement. Several examples of basin margin complexes have been preserved in which the relationship between various supracrustal facies and the sialic basement is evident. The sedimentary fill, which forms by far the greatest proportion of the supracrustal rocks, consists almost entirely of greywacke-mudstone turbidites derived from a mixed felsic volcanic and granitic source. Minor fluvial sandstones and conglomerates occur at the fault-basin margins. Mafic volcanics consisting mainly of massive and pillowed flows and intrusions occur sporadically in narrow linear belts also at the margins of basins. Felsic volcanic complexes occur within, at the margins of and between basins and consist mainly of volcaniclastic deposits that were in part terrestrial. Some form of interaction between hypothetical Archaean plates remote to that part of the ensialic crustal segment now preserved as the Slave Province may be an explanation for the short period of province-wide extension that resulted in the formation of graben-like basins in which the supracrustal rocks were deposited.

INTRODUCTION

For many years geological thought regarding the Archaean was dominated by examples from relatively well preserved granite-greenstone terranes (e.g. Anhaeusser et al., 1969; Pettijohn, 1972). The Archaean supracrustal "package" consisting of thick sequences of dominantly mafic volcanics and lesser amounts of characteristically immature greywackes and mudstone sediments, the apparent lack of extensive mature sandstones and carbonates so common in the later record, the relatively low pressure metamorphism and the abundant diapiric intrusives, led to visions of a rather unstable Archaean world developed on thin, or at least highly mobile, crust in which the existence of a sialic component was, in the minds of some, in considerable doubt. More recently, attention has been drawn to other Archaean terranes where granulite facies mineral assemblages imply a thick Archaean crust, perhaps not unlike that of the present (O'Hara, 1977). Significant thicknesses of Archaean quartzite in some localities imply periods of considerable stability (Schau, 1975). The recognition of very ancient granitic terranes up to one billion years older than many of the familiar granite-greenstone terranes (Moorbath et al., 1972) implies the existence of granitic crust during much of the early Archaean, even if its extent remains unknown. This has required the application of certain constraints in the development of models explaining tectonic processes during the Archaean. In addition to these extremes there are commonly significant, if somewhat less dramatic, variations within any given class of Archaean terrane. It is not useful and probably counterproductive to think in terms of a "type example" of Archaean terrane such as granite-greenstone belts as suggested by Burke et al. (1976). Indeed, as much can be gained from understanding and appreciating the differences as the similarities and before a worldwide synthesis of Archaean tectonics is attempted, both must be considered. It is with this in mind that the following discussion of the evolution of Archaean basins of the Slave Structural Province, a sediment-dominated granite-greenstone terrane, is presented. This, of course, is only the first part of the story. A synthesis of the subsequent magmatic, thermal and deformational history of the province is not included.

THE SLAVE STRUCTURAL PROVINCE

The Slave Structural Province is located in the northwestern part of the Canadian Shield between Great Slave Lake and Coronation Gulf, where it occupies an area of about 190,000 km^2 (Fig. 9-1). The province is an Archaean "granite-greenstone" terrane that has undergone no major deformation other than faulting in the past 2.5 Ga. The province is unconformably overlain at its southern and northwestern margins and part of its north-central region by flat-lying to gently dipping early Proterozoic sediments. The eastern boundary is a tectonic front marked by the transition from relatively low-grade metamorphic rocks with irregular structural trends on the Slave side to granulite facies rocks with pronounced linear structural trends of the adjacent western Churchill Structural Province. The extent of Archaean rocks is not confined by the boundaries of the Slave Province as the rocks on the Churchill side are also Archaean (Fraser, 1978) as are the rocks of much of that province, although the nearest major well-preserved granite-greenstone terrane occurs 500 km to the east (Davidson, 1972a). In contrast to this terrane and the supracrustal rocks of the Slave Province a major sequence of Archaean quartzites occurs in the northern Churchill Province (Schau, 1977). The Slave Province has been mapped at various scales ranging from 1:500,000 for northern parts of the province to 1:10,000 or better in selected areas of economic interest. The geology has been summarized some time ago by McGlynn and Henderson (1972) and more recently the metamorphism and deformation has been discussed by Thompson (1978).

The geology of the Slave Province is similar to other granite-greenstone

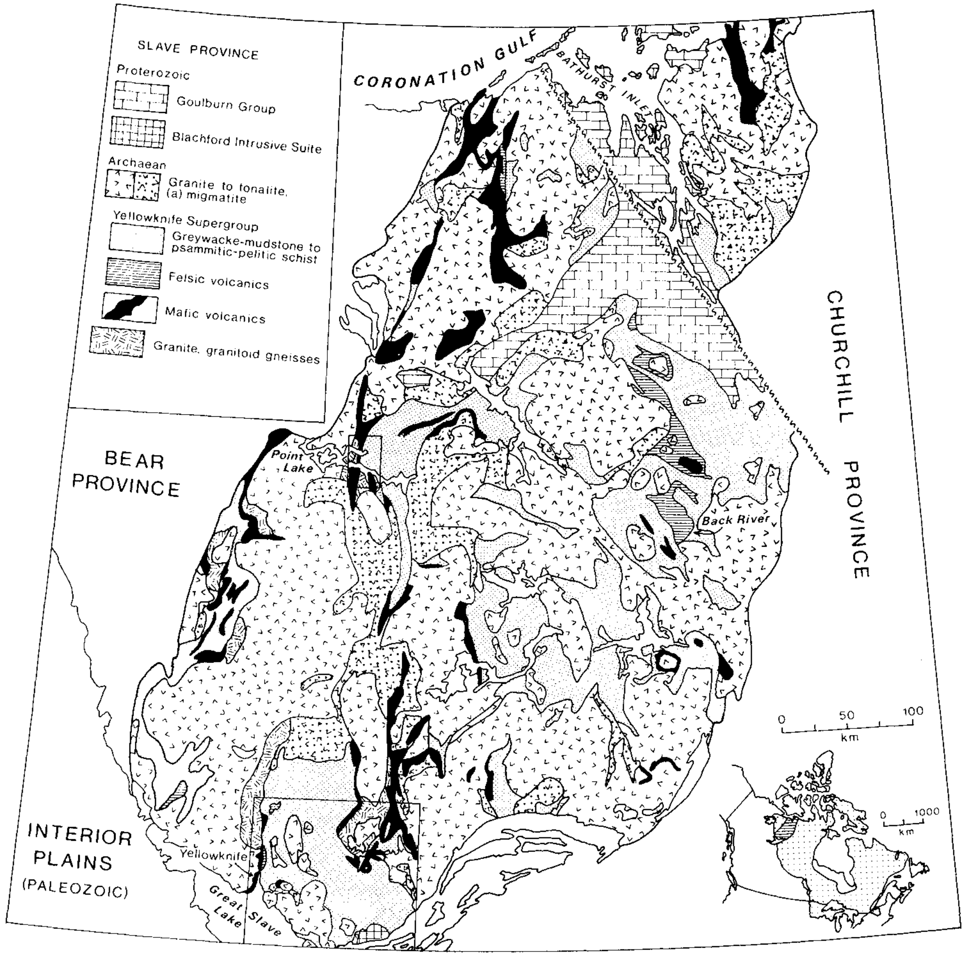


Fig. 9-1. Geological map of the Slave Province showing distribution of Archaean supracrustal rocks, areas of known and probable basement to the supracrustal rocks and later intrusive rocks. Note the high proportion of sediments to volcanics. Small outlined area in the west-central part of the province is the Point Lake area (Fig. 9-3), while the larger area outlined in the southern part of the province is the Yellowknife area (Fig. 9-2). After McGlynn (1977).

terranes of the world in that the supracrustal rocks consist mainly of mafic volcanic sequences and the sediments are greywacke-mudstones, all of which are intruded by tonalitic to granitic bodies. The rocks are metamorphosed to varied degrees in the lower pressure facies series. However, about 80% of the supracrustal terrane of the Slave Province is underlain by sediments while only about 20% by volcanics; the proportion is reversed in most other granite-greenstone terranes. The large volume and the composition of these sediments have important implications as to the nature and extent of the source terrane. Basement to the supracrustal rocks is recognized in several places in the province and is well enough preserved in a few cases that the relationship between volcanics, sediments and basement is evident. This, together with the preservation of a fairly complete section across an Archaean basin segment in the southern part of the province (Fig. 9-2), allows reasonable speculations as to their origin to be considered.

SUPRACRUSTAL ROCKS — THE YELLOWKNIFE SUPERGROUP

Approximately half the province is underlain by Archaean supracrustal rocks of the Yellowknife Supergroup and their more highly metamorphosed and migmatized equivalents (Fig. 9-1). Of this, about 20% is underlain by dominantly mafic volcanic sequences with local felsic centres. The remainder consists almost entirely of greywacke-mudstone turbidites. The supracrustal rocks are contained in a sea of granitic rocks, almost all of which are intrusive into them.

The Yellowknife rocks occur in patchy, more or less connected areas of varied size and shape (Fig. 9-1). In general, the volcanic units occur along the contact between the intrusive granites and the sedimentary terrane. Structural trends as defined by the steeply dipping volcanic units are northerly, although in detail there are many deviations. The sedimentary rocks are typically complexly folded into steep irregularly trending isoclines. Metamorphism is in the lower pressure facies series and ranges from greenschist to upper amphibolite (Thompson, 1978). In many areas the change from greenschist to amphibolite facies is easily mapped with the occurrence of coarse cordierite porphyroblasts (Fig. 9-2). The lowest-grade metasediments in general occur within the central part of the large supracrustal areas.

As will be discussed in this paper, the Archaean supracrustal rocks are thought to be deposited in a series of small fault-bounded basins characterized by marginal mafic volcanism and central greywacke turbidite deposition. The fault-bounded basins formed on older sialic basement due to regional extension of the crust over an area at least that of the Slave Province.

Volcanic rocks

Volcanic rocks of the Slave Province range in composition from basalt to rhyolite. No major ultramafic units have yet been recognized. Many volcanic sequences are dominated by either mafic or felsic compositions although in any succession both are usually present to some extent.

Mafic volcanic sequences are by far the most abundant. They occur as

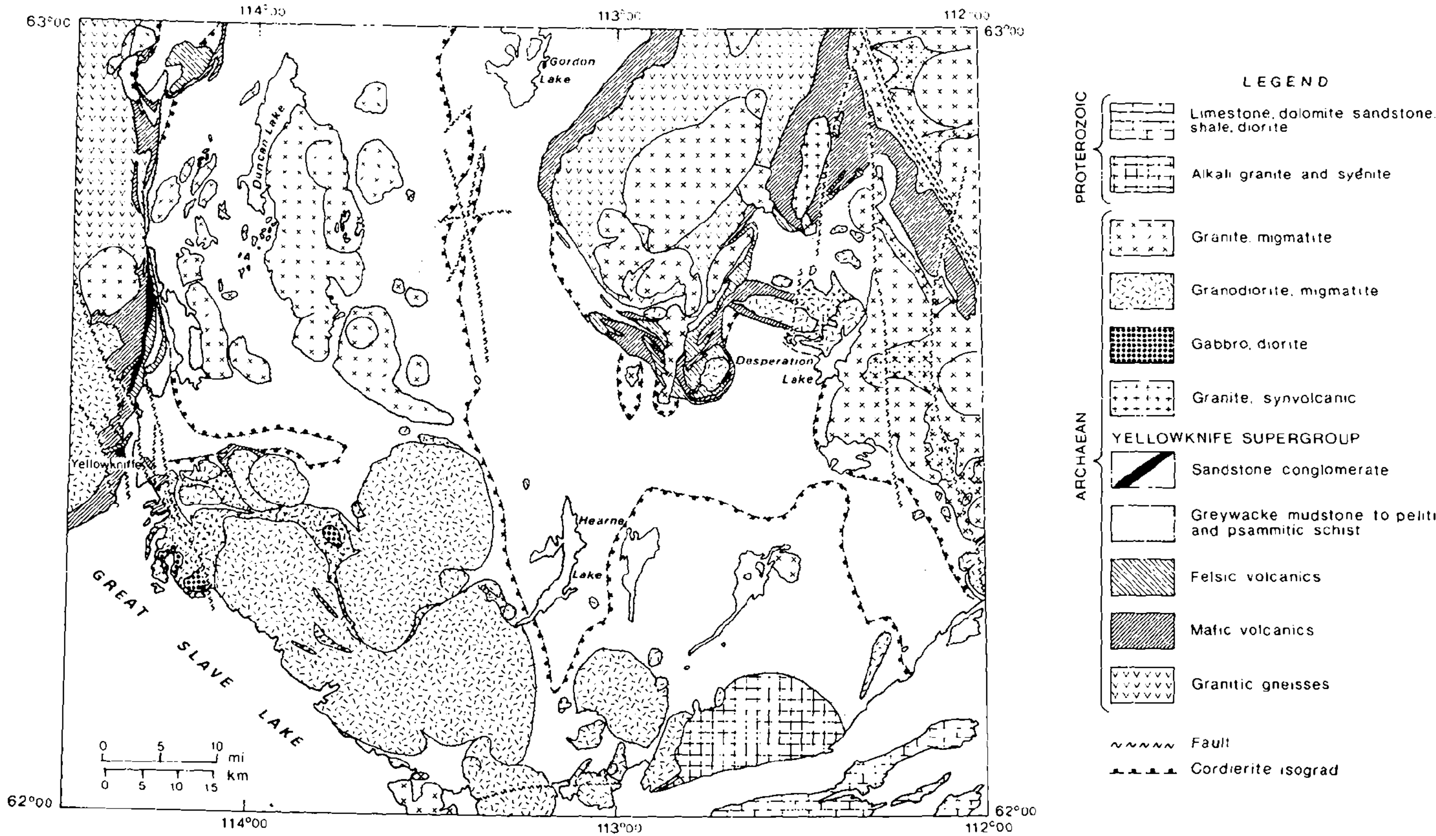


Fig. 9-2. Geological map of the Yellowknife area in the southern Slave Province. A segment across part of an Archaean basin is preserved within the map area. After Henderson (1976).

elongate belts normally at the edge of supracrustal areas. The best known belt is at Yellowknife (Baragar, 1966; Henderson and Brown, 1966; Fig. 9-2) and is thought to be representative of mafic volcanic dominated sequences. This sequence has an apparent thickness of about 7000 m although, as is normal, the base is not preserved due to later granodioritic intrusions. The top is locally unconformably overlain by terrestrial sediments and felsic volcanics. The sequence consists dominantly of basaltic and, to a lesser extent, andesitic massive and pillowed flows, indicating subaqueous accumulation. Individual flows are, in some cases, in excess of one hundred metres thick but are not particularly extensive laterally. Towards the top of the sequence volcanic breccia and other volcaniclastic units are more abundant. Significant argillaceous sedimentary units within the succession are unknown. The volcanic flows are intruded by several series of dykes, sills and irregular intrusions that are more or less contemporaneous with volcanism and are thought to represent part of the conduit system for volcanism at higher levels in the succession. A series of variolitic flows and tuffaceous horizons occur within the sequence and have proved to be useful marker horizons. A 400 m thick felsic volcanic unit occurs in the central part of the sequence and is similar to the up to 1300 m thick felsic volcanic unit that occurs above the local unconformity.

The mafic succession at Yellowknife comprises two calc-alkaline cycles separated by the central felsic unit (Baragar, 1966). Superimposed on these calc-alkaline trends are several tholeiitic trends of short duration although there is no overall enrichment of iron through the sequence. Baragar suggested the calc-alkaline trends are due to contamination of tholeiitic magma as it passed through sialic crust while the periodic tholeiitic differentiation pattern is due to magma having been tapped from successive small magma chambers.

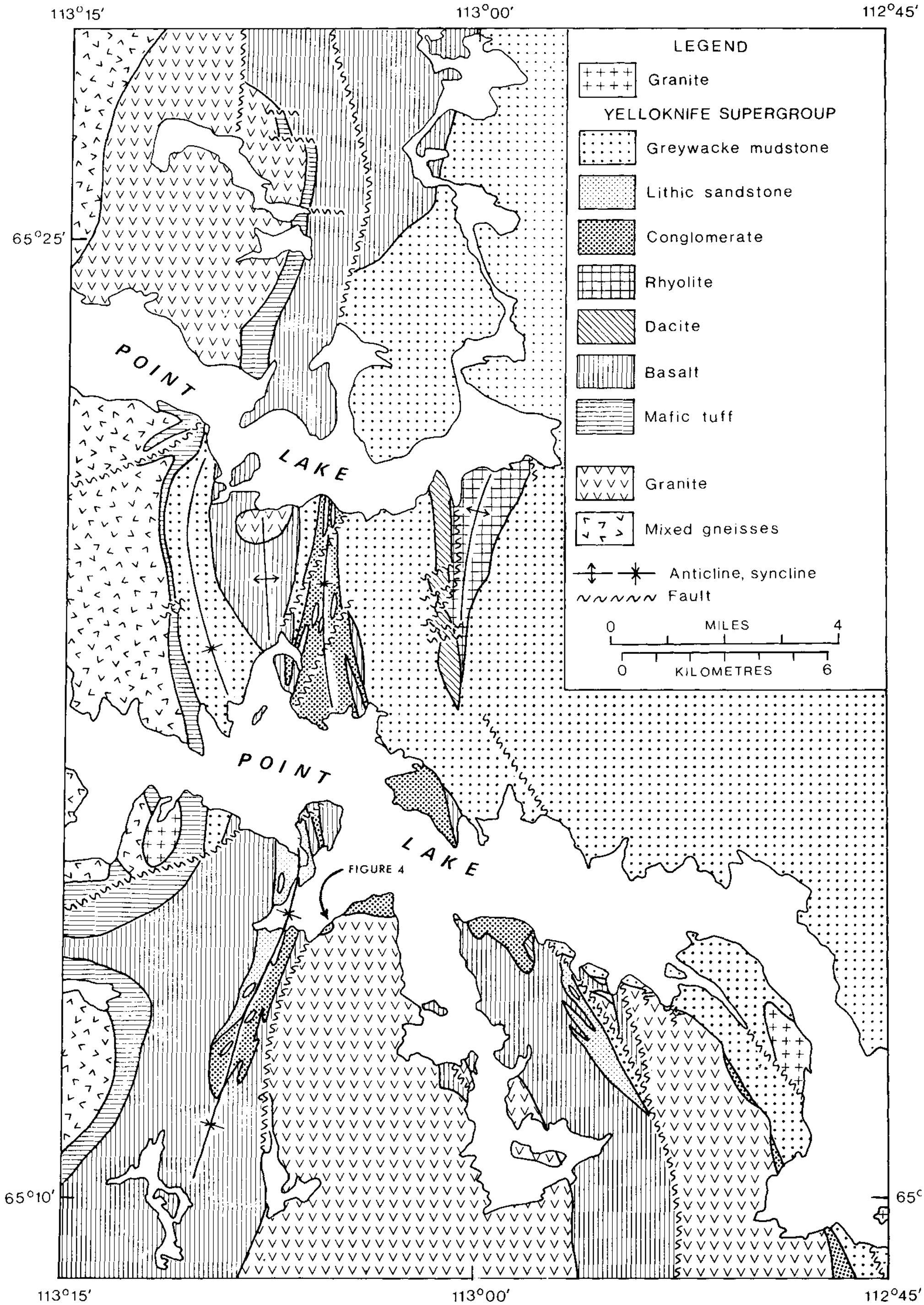
In most of the mafic volcanic belts the volcanic sequence is conformably overlain by deep water greywacke-mudstone turbidites. An important exception, however, occurs at Yellowknife on the margin of a depositional basin where a local angular unconformity separates the main mafic volcanic sequence from overlying fluvial sediments and a felsic volcanic unit. The distal part of the mafic sequence, a few kilometres to the east out in the basin, is more typical in that it is conformably overlain by deep water sediments. The local unconformity can be explained by uplift along the basin margin while mainly felsic volcanism was still active. Further evidence of this uplift is that prior to the formation of the unconformity part of the volcanic edifice slid to the east into the basin (Henderson, 1978).

The volcanic sequences are almost always overlain by sediments but there is no evidence that volcanics extended any distance into the basin beneath the sediments. Gravity profiles across the volcanic sequence and adjacent sediments indicate that no significant thickness of mafic volcanics occurs below the sediments. Direct evidence of this relationship is rarely evident but at Point Lake, in the west-central part of the Slave Province, basement to the supracrustal succession is preserved (Fig. 9-3). Here, resting on the basement, a thick volcanic sequence marginal to the basin and overlain by sediments, thins into the basin. Some 15 km into the basin the volcanics are no longer present and only the normal basinal turbidites overlie the basement. Thus the mafic volcanic sequences appear to have erupted from centres along long narrow zones of weakness at the margins of basins.

Volcanic complexes dominated by felsic material contrast markedly to the mafic sequences. Felsic complexes are much less abundant and those that are preserved tend to occur as more or less equidimensional complexes within the sedimentary terrane. Commonly there is evidence of subaerial accumulation in part (Lambert, 1977, 1978). The most striking example is the Back River complex in the west-central Slave Province (Lambert, 1978). It underlies a roughly rectangular area, 40 km long by 20 km wide. Most of the complex consists of volcaniclastic deposits that range in composition from andesite through dacite to locally rhyolite. These were deposited as both air-fall and water-lain tuffs, ash flows and pyroclastic breccias. Some of the dacites and rhyolites also occur as lava flows and domes. Both subaerial and subaqueous depositional environments are represented, together with intrusive bodies contemporaneous with volcanism. In its later stages of development the complex collapsed to form a cauldron subsidence feature. Only a few other dominantly felsic complexes are known. These include the volcanics 10 km north of the Back River complex which is the southern part of a large mixed felsic and mafic belt (Frith and Percival, 1978), a unit 6 km in diameter 50 km north of Yellowknife (Fig. 9-2) and a felsic complex 100 km northwest of Yellowknife (Fig. 9-1). None of these is known in any detail.

On the basis of the few examples at hand, these felsic complexes have a rather random pattern of occurrence. Some occur at the edges of supracrustal areas while others are completely surrounded by sediments. They also occur together with dominantly mafic volcanic sequences if some of the thicker felsic units associated with the mafic sequences are considered (i.e. up to 1300 m at Yellowknife). There is no tendency for them to line up or form linear features as is the case with mafic volcanic sequences. It is suggested later in this paper that many more felsic volcanic centres must have existed at one time than is indicated by the preserved record, to account for the high proportion of volcanic clasts in greywackes that form by far the greatest part of the Yellowknife Supergroup. Although not preserved, these complexes may have been located in areas now occupied by intrusive plutons. It is suggested that the preserved felsic volcanic complexes within the basins may represent the extrusive equivalents of plutons similar to those found elsewhere in the basins that are not exposed at the present erosion level.

Between the extremes of felsic and mafic-dominated volcanic complexes are volcanic sequences in which neither end member is dominant. In such



113°15'
113°00'
112°45'
65°25'
65°10'
65°
LEGEND
Granite
YELLOKNIFE SUPERGROUP
Greywacke mudstone
Lithic sandstone
Conglomerate
Rhyolite
Dacite
Basalt
Mafic tuff
Granite
Mixed gneisses
Anticline, syncline
Fault
0
MILES
4
0
KILOMETRES
6
POINT
LAKE
POINT
LAKE
FIGURE 4
113°15'
113°00'
112°45'

sequences all ranges of composition are represented and are complexly interrelated. For example, a small complex 80 km east of Yellowknife (Fig. 9-2) consists of pillowed mafic volcanics interfingered with rhyolite and dacite tuffs, while at the top of the sequence are felsic tuffs and ignimbrites, indicative of subaerial deposition (Lambert, 1978). Just north of the Back River complex is a major belt of basaltic to rhyolitic volcanics in which the occurrence of two cauldron collapse structures has been suggested (Frith and Percival, 1978; Frith and Roscoe, 1980). Associated with these cauldrons are major Cu-Zn mineralized zones. In fact, all the significant base metal deposits in the Slave Province occur in mixed felsic to mafic volcanic sequences.

In summary, volcanic complexes in the Slave Province range from dominantly mafic successions that tend to occur at edges of basins and were extruded along narrow elongate zones to equidimensional, dominantly felsic complexes that have no regular pattern of distribution.

Sedimentary rocks

On the basis of outcrop area sedimentary rocks are by far the most important part of the preserved Yellowknife Supergroup, comprising about 80% of the supracrustal terrane. They can be discussed in terms of three associations: deep water greywacke-mudstone turbidites, terrestrial conglomerates and sandstones, and chemical sediments.

Greywacke-mudstone turbidites and their more highly metamorphosed equivalents are by far the most abundant, underlying almost all the sedimentary terrane. In general, they conformably overlie the mafic volcanic sequences and in some cases are interlayered to some degree. The original thickness of the sediments is difficult to estimate as units that occur above the sediments are rare and, where present, are not extensive. In addition, the sediments are complexly deformed and lack distinctive marker horizons. Where thickness estimates have been attempted they are on the order of one or several thousand metres. At Yellowknife, for example, a thickness of 5000 m is estimated — probably a maximum figure as the possibility of as yet unrecognized tectonic stacking has not been taken into account. Throughout the Slave Province the sediments are similar, consisting of greywacke to siltstone turbidites and interbedded mudstone layers ranging in thickness from less than a centimetre to several metres. Sedimentary

Fig. 9-3. Geological map of the Point Lake area. This area is at the margin of a fault-bounded greywacke-mudstone turbidite-filled Archaean basin with mafic volcanics extruded at the margin on granitic basement in a north-trending linear belt that thins to the east. Conglomerate derived from uplifted marginal basement blocks occur above, within and below the supracrustal section. The relationship between the granitic basement and the supracrustal rocks is revealed in a series of fault blocks that essentially provide a cross-section normal to the basin margin. After Henderson and Easton (1977).

structures characteristic of turbidites are commonly well preserved in the lower grade metamorphic rocks. At Yellowknife it has been suggested that these sediments accumulated on a subaqueous fan complex with the sediment being derived from the west, a direction more or less normal to the nearby basin margin from an area now underlain by intrusive granodioritic rocks (Henderson, 1975a). The only other palaeocurrent data on these sediments is from an area 300 km to the northwest but on the west side of the same granitic terrane. There, Ross (1962) has shown that the sediments were brought into the basin from the east. This opposition of palaeocurrent directions favours the hypothesis of two distinct basins separated by an intervening source terrane.

The coarsest components of the greywackes are quartz, rock fragments and plagioclase that grade down to an argillaceous quartzofeldspathic matrix. Felsic volcanic rock fragments are the most abundant, with only very few mafic clasts in most samples. Quartz-plagioclase rock fragments are rare but ubiquitous. However, the difficulty of preserving quartz-plagioclase aggregates as fragments in a sediment consisting of grains commonly less than 2 mm together with the abundance of individual quartz and feldspar grains suggest the granitic (or possibly gneissic) component in the source terrane is probably significant. Chemically the greywackes are similar to both an average granodiorite and an estimate of the composition of the

TABLE 9-I

Comparison of chemical composition of Yellowknife greywacke, average granodiorite and Canadian Shield

	Yellowknife greywacke (Henderson, 1978)	Average granodiorite (LeMaitre, 1976)	Granodiorite standard deviation	Canadian Shield (Eade and Fahrig, 1971)
SiO_2	68.39	66.80	4.36	66.1
TiO_2	0.66	0.54	0.30	0.6
Al_2O_3	15.77	15.99	1.49	16.0
Fe_2O_3	0.72	1.52	1.06	1.2
FeO	4.68	2.87	1.49	3.5
MgO	2.83	1.80	1.12	2.2
CaO	1.76	3.92	1.49	3.4
Na_2O	3.21	3.77	0.86	3.8
K_2O	1.98	2.79	0.99	3.3

The chemical composition of both an average greywacke at Yellowknife and an estimate of the composition of the Canadian Shield are very similar to LeMaitre's (1976) average granodiorite. The only significant deviations are the somewhat higher proportion of ferrous iron in both the greywacke and Shield estimates and the high proportion of calcium in the greywacke. All other oxides fall within the standard deviation of the granodiorite average.

present Canadian Shield (Table 9-I). It therefore appears that the source terrane for the sediments was chemically similar to the present shield, but consisted mainly of felsic volcanic and granitic and/or gneissic rocks. When the volume of sedimentary material is considered, the area of the source terrane must have been extensive, and in order for sufficient erosion to have taken place to generate the large volume of sediment, the terrane must have been largely subaerial.

Fluvial sandstones and conglomerates are known at several localities in the Slave Province (Henderson, 1975b; Henderson and Easton, 1977; Tirrul and Bell, 1980). At almost all localities they are associated with mafic volcanic sequences although in at least one place they also occur above granitic basement. At Yellowknife a basal conglomerate, composed mainly of subangular volcanic clasts and locally well-rounded granitic cobbles, fills depressions on the unconformity surface that truncates the mafic volcanic succession. Conformably overlying the basal conglomerates are up to 300 m of cross-bedded, volcanic lithic sandstones with thin to lensoid beds of pebbly conglomerate that are more abundant in the lower part of the section. Sandstones occur as heterogeneous, well bedded parallel-sided to lensoid beds in which shallow scours and channels are common. These have been interpreted as braided river deposits (Henderson, 1975b). The sandstones are composed almost entirely of felsic volcanic rock fragments and quartz. A felsic volcanic source is indicated; one that was probably contemporaneous with sedimentation as the fluvial sandstones, in turn, are conformably overlain by a major felsic volcanic unit. They are thought to represent the terrestrial basin margin equivalent to the adjacent basinal greywacke mudstone turbidites.

A similar situation occurs to the north at Point Lake. There, conglomerates and sandstones unconformably overlie granitic basement as well as mafic volcanics with which they are also locally interbedded. The sandstones are more feldspathic, reflecting the granitic component in the source terrane.

Seventy-five km northeast of Yellowknife (Fig. 9-2) up to 60 m of subrounded quartz sandstone, in part carbonate-cemented, overlie and laterally interfinger with the distal volcaniclastic deposits of a major mafic sequence to the north. Scattered granitic clasts in the dominantly quartz-rich sandstone reflect a granitic provenance, presumably the granitoid basement that underlies both the volcanics and sediments in the area.

Conglomerates without associated fluvial sandstones are known from several localities. As elsewhere they are composed mainly of volcanic clasts, but in many cases also contain granitic clasts. They are closely associated with mafic volcanic sequences, either overlying them (McGlynn and Ross, 1963; Fraser, 1969) or interlayered with them (Moore, 1956).

In attempting to understand the nature of the depositional basins, it is of interest that these terrestrial sediments are always associated with the mafic volcanic sequences and, in several cases, with areas of known or suspected granitic basement. With one exception (Heywood and Davidson, 1969) none are found with basinal turbidites in the central parts of supracrustal areas. The fact that the sandstones are fluvial deposits and both the sandstone and conglomerate are composed largely of material similar to units they overlie or are overlain by, suggests that they are of rather local derivation and represent a terrestrial facies of the basinal greywacke turbidites.

The third group of sediments of the Yellowknife Supergroup comprises the chemical sediments — iron formations and carbonates. Iron formations are particularly abundant in the central part of the Slave Province northeast of Point Lake although they have been found in most other regions as well. Oxide, carbonate, silicate and sulphide facies are present, and in some localities zones can be defined on the occurrence of the various types. The iron formations occur mainly with the basinal sediments as thinly layered to laminated units, usually less than a metre thick and a few hundred metres across, interbedded with the greywacke-mudstones. Oxide iron formation is also present locally with some mafic volcanic sequences, although it is much less abundant.

Limestone but more commonly dolomite units are also a minor facies in the supracrustal succession. When present they are most commonly associated with felsic volcanics and black carbonaceous locally pyrite-bearing mudstone and, typically, occur near the transition between volcanics and greywackes. For example, 80 km east of Yellowknife carbonates can be discontinuously traced over a distance of 50 km at this contact. At the Back River volcanic complex carbonate occurs sporadically at the margin both as bedded carbonate and commonly as cement to extensive volcanic breccias (Lambert, 1978). Near High Lake in the northern Slave Province stromatolitic forms occur in a discontinuous carbonate unit at the contact between felsic volcanics and sediments (Henderson, 1975a). If these forms are analogous to more recent stromatolites and are formed by photosynthetic organisms this would suggest that these stromatolitic carbonates at least formed in a shallow-water environment.

In summary, greywacke-mudstone turbidites derived from a mixed plutonic and felsic volcanic terrane comprise the main basinal fill. At the margins of the basins fluvial sandstones and conglomerates occur locally, commonly lying unconformably on mafic volcanics or, in one case, on granitoid basement, and are thought to be the terrestrial facies equivalent of the basinal turbidites. Minor chemical sediments include the various facies of iron formation that occur locally with the basinal turbidites and carbonates, in at least one case of biogenic origin, typically associated with felsic volcanics.

BASEMENT

Within the Slave Province several examples of granitic basement to the

Archaean Yellowknife supracrustal rocks are known and several others have been suggested. Some of these have been discussed by Baragar and McGlynn (1976).

The best exposed and preserved example is at Point Lake in the west-central Slave Province (Stockwell, 1933; Henderson and Easton, 1977; Figs. 9-3 and 9-4). There, the relationship between granitic basement and

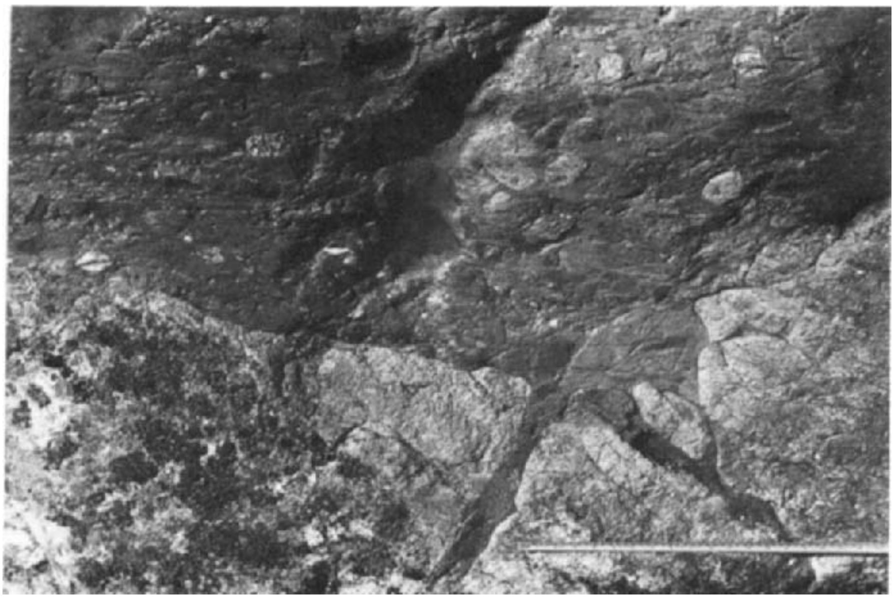


Fig. 9-4. Basal unconformity at Point Lake between granitic basement and overlying granitic and volcanic pebble conglomerate. Fractures in the unconformity surface are filled with pebbles. Scale is 43 cm long. Location noted in Fig. 9-3.

both volcanic and sedimentary rocks is exposed. The basement rock is a massive, equigranular, leucocratic, perthitic granite in which the primary mafic minerals are altered to chlorite. Subsequent to deposition of Yellowknife volcanics and sediments parts of the pluton have been metamorphosed. Chlorite both in fractures and in the granite itself is now fine-grained and interwoven with masses of biotite, indicating both the fracturing and original alteration is pre-metamorphism. In several places an Archaean weathered zone is developed in the granite in which feldspars are extensively altered although the original granitic texture of the rock is still apparent, mainly in the quartz textures. At some localities a thin lag deposit of very angular quartz, presumably derived from the weathered zone, occurs immediately above the unconformity. More commonly, however, a conglomerate occurs in which generally rounded cobbles derived from both the fresh basement granite and the weathered equivalent are a major component. In the western exposures, where mafic volcanics occur in the section, angular mafic volcanic clasts are commonly dominant in the conglomerate. Conglomerates similar to this basal conglomerate, containing both volcanic and the granitic clasts, also occur both within and above the mafic volcanic sequence. In the east, however, beyond the extent of the volcanics, the cobbles in the conglomerate are almost entirely granitic with no mafic volcanic component.

To the west, but in probable fault contact with the mafic volcanics, are tonalitic to granodioritic gneisses, locally migmatitic, with amphibolitic and pelitic gneiss zones (Fig. 9-3). The striking contrast in structural style and metamorphic grade with the relatively low-grade Yellowknife rocks to the east may be evidence that these gneisses are also older than the Yellowknife. A preliminary study of zircons from this unit (Krogh and Gibbins, 1978) indicates the zircons are of two generations with the brown first generation zircons from a single sample having a minimum age of 2730 Ma, and the second generation white zircons a probable age of about 2600 Ma. If both terranes are basement to the Yellowknife they represent a considerable variation in the basement from massive potassic plutons to migmatitic intermediate gneisses.

Another example of basement to the Yellowknife supracrustal rocks occurs 80 km northeast of Yellowknife (Fig. 9-2) and was first recognized by Baragar (1966). The contact between the basement and the overlying mafic volcanics is an unconformity along which some movement has later taken place although locally the unconformity is preserved. In the vicinity of the unconformity the basement rock is a cataclastically deformed granodiorite to tonalite that locally contains highly deformed layers of amphibolite. The cataclastic foliation is locally discordant to the overlying volcanics (Davidson, 1972b). Swarms of mafic sills and dykes, commonly with chilled margins and in some cases with coarse plagioclase phenocrysts preserved despite amphibolite grade metamorphism, are present in the basement. Most occur as elongate lensoid bodies a few metres to several tens of metres wide in the granodiorite parallel to the flows, although some dykes strike directly towards the mafic volcanics. They are similar to the abundant sills and dykes within the volcanic succession but do not occur in the immediately overlying sediments. They are thought to be part of the subvolcanic feeder system. To the south the volcanics thin and eventually give way to the previously mentioned thin conglomerate and sandstone unit. At its base the conglomerate is derived mainly from mafic volcanics but contains angular deformed granitic clasts similar to the underlying basement in the upper part. The sandstones are composed mainly of subangular quartz but occasionally quartz-plagioclase clasts also suggest derivation from the tonalitic basement.

Heywood and Davidson (1969) have suggested that a small metatonalite body, mantled by mafic volcanics, at Brislane Lake 200 km east-northeast of Yellowknife (Fig. 9-1) may be basement to the adjacent supracrustal rocks. As in previous examples, mafic dykes occur in the tonalite and in lower parts of the volcanic sequence but not in the upper parts of the succession. No contact metamorphic aureole in the supracrustal rocks about the tonalite has been recognized.

Evidence for other candidates for basement is less direct. Frith et al. (1977) report a Rb-Sr whole rock isochron age of 2939 ± 51 Ma on the most granitic phases of the tonalitic gneiss 300 km north-northwest of Yellowknife. At Yellowknife zircons in boulders of tonalitic gneiss in a diatreme in the mafic volcanic sequence yield zircon $^{207}Pb/^{206}Pb$ ages up to 3040 Ma (Nikic et al., 1975). Both suggest that 3 Ga old tonalites or tonalitic gneisses existed in the region before deposition of the Yellowknife supracrustal rocks.

GEOCHRONOLOGY

Geochronological data on the Slave Province is sparse, particularly for the Yellowknife supracrustal rocks. Archaean geochronology has been studied by Green and Baadsgaard (1971) in the vicinity of Yellowknife, and Krogh and Gibbins (1978) in the Point Lake area. Frith et al. (1977) have reported on the geochronology of the granitic rocks 300 km north-northwest of Yellowknife, on the western boundary of the Slave Province. Other data is mainly of a preliminary nature but certain patterns are becoming evident. In the following discussion Rb-Sr dates have been recalculated using the $1.42 \times 10^{-11} a^{-1}$ rubidium decay constant.

Geochronological data on three examples of basement to the Yellowknife Supergroup indicate ages close to 3 Ga. At Point Lake, Krogh and Gibbins (1978) report a U-Pb concordia intercept age of 3155 Ma, based on 3 zircon fractions from one sample. As previously mentioned, Nikic et al. (1975) report zircon $^{207}Pb/^{206}Pb$ ages up to 3040 Ma on clasts presumably derived from a tonalitic basement in a diatreme in the volcanic sequence, and Frith et al. (1977) report a 2939 Ma whole-rock age (Rb-Sr isochron, Sr_i = 0.700 ± 0.001) for a tonalitic gneiss unit 300 km north-northwest of Yellowknife. On the other hand, zircons from the basement locality 80 km northwest of Yellowknife have a $^{207}Pb/^{206}Pb$ age of 2640 Ma although a whole-rock Rb-Sr isochron from the same unit indicates an age of 2457 ± 120 Ma (Sr_i = 0.707 ± 0.005) (Green and Baadsgaard, 1971). The volcanics that overlie the basement have a Rb-Sr isochron age of 2574 ± 200 Ma (Sr_i = 0.706 ± 0.005) (Green and Baadsgaard, 1971) which falls between the two basement age estimates. Indeed the basement zircon age is similar to the zircon $^{207}Pb/^{206}Pb$ age of felsic volcanics at Yellowknife (Green and Baadsgaard, 1971) which are believed to be contemporaneous with the volcanics above the basement. None of the ages in this area are regarded as primary ages by the writer, since the isotopic systems may have been disturbed by large post-Yellowknife granitic intrusions or, in the case of the basement zircons, by the Yellowknife volcanism. Similarly, the possible basement tonalite at Brislane Lake has an age of 2656 ± 88 Ma (Rb-Sr whole-rock isochron, $Sr_i = 0.7011 \pm 0.0026$; Davidson, 1972c), but Davidson concluded that the age neither confirmed nor denied that the tonalite was basement to the Yellowknife volcanics that mantle it. Thus geochronological data on documented or proposed basement terranes are either, at about 3 Ga, comfortably older than the supracrustal rocks or so close to them that the indicated age of the basement is not significantly different from the apparent age of deposition of the overlying Yellowknife Supergroup.

For the Yellowknife supracrustal rocks Green and Baadsgaard (1971), as previously mentioned, report a zircon $^{207}Pb/^{206}Pb$ age of 2650 Ma from dacites in the volcanic sequence at Yellowknife and a Rb-Sr isochron age for the same sequence of volcanics at 2570 ± 160 Ma ($Sr_i = 0.7022 \pm 0.0023$). The isochron is very close to the 2574 ± 200 Ma isochron they determined for the volcanics 80 km to the east and both are regarded by the writer as minimum ages that may reflect disturbances by later intrusions. A single population of zircons from greywackes at Yellowknife has a $^{207}Pb/^{206}Pb$ age of 2680 Ma (R.K. Wanless, pers. commun., 1969) and probably reflects the age of the felsic volcanic/granitic source. The Back River felsic volcanic complex in the east-central part of the province has a zircon concordia intercept age of 2667 ± 7 Ma, based on two zircon fractions of a sample from a rhyolite dome in the complex and two fractions from a sample of the adjacent greywackes (R. K. Wanless, pers. commun., 1978). Green and Baadsgaard (1971) have dated zircons from boulders in the conglomerate at Yellowknife at 2595 Ma and 2575 Ma respectively ($^{207}Pb/^{206}Pb$). These dates are somewhat younger than the age of the sequence in which the conglomerate is found, but fall within the range of zircon ages of the granodiorite that intrudes the sequence. Taking the rather meagre zircon data available for the supracrustal rocks at face value it would appear that these rocks were deposited between 2650 and 2680 Ma ago and that deposition of these rocks took place in at least two widely separated parts of the province at this time.

Archaean granites intrusive into the Yellowknife in general range in age from about 2.65 to 2.5 Ga (Green and Baadsgaard, 1971; Frith et al., 1977).

ORIGIN AND EXTENT OF THE YELLOWKNIFE SUPRACRUSTAL BASINS

Yellowknife supracrustal rocks occur throughout the Slave Province. Is it reasonable to make an attempt to outline depositional basins?

Several remnants of basin margins are recognized although it will be some time, if it is possible at all, before the original outline of any single basin can be defined with any degree of confidence. Basin margins are taken to be the lateral transition from terrestrial to deep-water deposition. This may be considered by some to be a somewhat unrealistic definition (i.e. Walker, 1978) as major accumulations of terrestrial deposits would have accumulated in a basin as well. For the Yellowknife, however, if large terrestrial deposits ever did exist, they have not been preserved. What is left are relatively thin terrestrial deposits at the edge of large remnants of much thicker deeper-water deposits, and this association is interpreted as a reflection of a primary tectonic feature. The transition is typically coincident with thick accumulations of mafic volcanics and evidence for basement commonly occurs nearby.

The best example of a preserved basin margin is at Point Lake (Fig. 9-3). There, as previously described, a long, linear, thick sequence of mafic volcanics unconformably overlies basement. The volcanics thin to the east over a few kilometres to where the basement is overlain by greywacke-mudstone turbidites. Locally overlying, interbedded with, and underlying the volcanics are conglomerates composed of both volcanic and granitic clasts. Similar conglomerates but without the mafic clasts occur to the east at the base and lower part of the sedimentary section above the unconformity. A crossbedded sandstone occurs with the conglomerates above the mafic lavas. In this region there is good evidence to support the proposal that the basin was fault-bounded. Mafic volcanism initiated along major north—south-trending fractures resulted in the present linear distribution of the mafic belt and its thinning to the east, away from various centres along the fracture zone. This relationship is well displayed at Point Lake (Fig. 9-3) where, at the margin of the basin, a series of block uplifts occur that generally dip easterly or northeasterly so that on any given fault-bounded block the basement granite occurs on the west side and the supracrustal rocks overlying basement occur on the easterly or northeasterly side of the block. Uplifted basement blocks in the basin-margin area shed granitic detritus as well as detritus from the earliest stage of volcanism caught on the upward rising blocks. Detritus continued to be shed from rising blocks as volcanism continued, resulting in deposition of conglomerate at the base, within and above the volcanic sequence. This intimate association of conglomerate and volcanics implies a source that was consistently, or at least periodically, higher than the accumulating volcanic sequence — a requirement most easily satisfied by a rising fault scarp. On the other hand, if the volcanics had accumulated in the central part of a basin, presumably they would form a positive topographic feature. While it is possible to transport coarse detritus many tens of kilometres into a basin to form deep-water conglomerates, it is unlikely that, having reached the low part of the basin, the detritus could be moved up slope to accumulate on the flanks and on top of the volcanic ridge.

Immediately following the initial formation of the basins felsic volcanism was initiated, possibly in response to the same forces that caused the original faulting. Felsic buildups formed both inside the basin, as shown by the small volcanic body within the sedimentary terrane at Point Lake, but presumably to a greater extent outside the basin. Erosion products of these volcanic edifices were a major contributor to the greywackes which are the main basinal fill. The basin continued to deepen as the supracrustals accumulated in it.

Another example of a basin margin occurs at Yellowknife (Fig. 9-2). Although not nearly as well displayed, the situation appears to be essentially identical with that at Point Lake. Additional evidence supporting the proposed model is that palaeocurrent data from the basinal turbidites indicate sediment transport normal to the fault-basin margin which is outlined by the orientation of the linear mafic volcanic belt. In addition, the apparent tilting of the mafic volcanic pile, followed by subaerial erosion and deposition of fluvial sediments in the basin-margin area, can be explained by differential movement of fault blocks in the marginal zone. Across the basin to the northeast, Lambert (1977) has suggested a similar mechanism for the evolution of the volcanic sequences in that area (Fig. 9-2). There, volcanic sequences encompass what appears to be a basement uplift with volcanic piles developed discontinuously along faults that are interpreted to bound the block uplift. On the west side of the block it is evident that volcanics were not extruded equally along the marginal fault zones. There, the mafic volcanic sequence, up to 4000 m thick and composed mainly of pillowed and massive flows, thins abruptly to the south to dominantly volcaniclastic deposits. These, in turn, are replaced mainly by conglomerates and quartzitic sandstones derived from the volcanic sequence and the granodioritic to tonalitic basement, all of which are overlain by greywackes derived from a mixed felsic volcanic and granitic provenance.

On the basis of these examples the model proposed for explaining the evolution of the Yellowknife supracrustal rocks requires the formation of an extensive block-faulted terrane (Fig. 9-5). Mafic volcanism took place locally along the basin-forming boundary faults but probably never extended very far into the main part of the basin. The basins themselves are underlain by large down-dropped blocks and were filled mainly by turbidites derived from uplifted marginal basement blocks and felsic volcanic centres both inside and out of the developing basin. The felsic volcanism was probably initiated by the same mechanism that resulted in the original faulting. It seems likely that these graben-like basins formed as a result of regional extension across the proto-Slave Province. It seems unlikely that rifting ever extended beyond the graben stage; the basins are floored by granitic basement and there is no evidence that the pre-Yellowknife crust was ever completely rifted such that oceanic crust formed within the basins.

It has proved difficult to extend the presently recognized basin margins

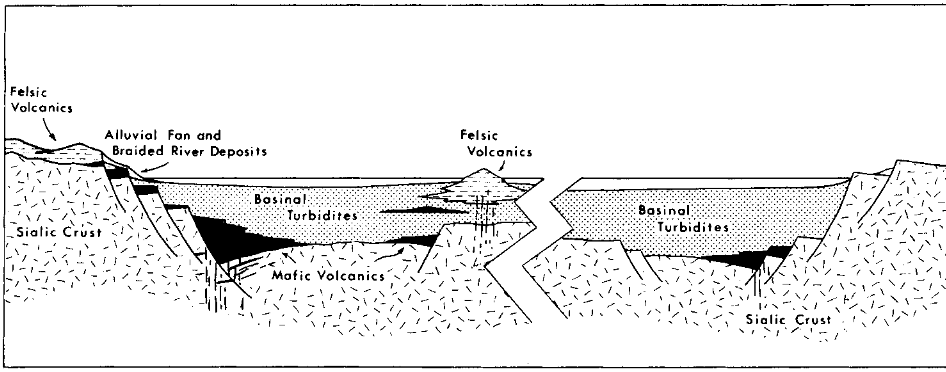


Fig. 9-5. Cross-section through fault bounded Archaean basin in which greywacke-mudstone turbidites derived from the basement and felsic volcanics on the uplifted basin margin are the main basinal fill. Mafic volcanics are restricted to basin-margin fault zones or fault blocks within the main basin. Felsic volcanics occur within and outside the basin and are a major contributor to the sediments.

to outline complete basins with any degree of confidence. The occurrence of other long linear belts of mafic volcanics between dominantly granitic terrane and large areas underlain by sediments, particularly where conglomerates occur with the volcanics, should be investigated as possible basin margin situations. At this time, however, it is difficult to choose between the possibility that the various supracrustal remnants represent several separate basins of the same or different ages or, possibly, one large interconnected basin in which the identified basin margins are only parts of uplifted blocks of unknown size in a much larger basin. The available evidence tends to favour a common age. Nowhere in any given area has more than one age of supracrustals been recognized; that is, no major regional unconformities have been identified that would indicate more than one period of supracrustal deposition. The sparse geochronological data suggests that deposition took place within a period of about 30 million years (2650—2680 Ma), based on zircon data from two widely spaced points in the province. Volcanic piles and associated sedimentary sequences on the scale of the Yellowknife supracrustal succession can accumulate over a relatively short period of time. Folinsbee et al. (1968) were the first to suggest this for the succession at Yellowknife when they compared it to a comparable thickness of Miocene mafic volcanics and turbidites of the fossa magna of Japan. The Miocene sequence accumulated in about 12 Ma and, on the basis of similar thickness, a similar rate may be reasonable for the Yellowknife although Folinsbee et al. (1968) also attempted to demonstrate the rate on geochronological grounds. Thus the entire Archaean supracrustal rock record in the Slave Province may only represent a 10 to 15 million year time span about 2665 Ma ago.

EVIDENCE OF PLATE TECTONICS IN THE SLAVE PROVINCE DURING THE ARCHAEAN

What can our present understanding of the supracrustal geology of the Slave Province contribute to the question of plate tectonics during the Archaean?

Attempts have been made in the past to relate aspects of Archaean geology to recent environments thought to be an expression of plate tectonic processes. For example, Folinsbee et al. (1968) have suggested that the mafic volcanic sequence at Yellowknife is similar to an island arc using, as an example, the Miocene rocks of the fossa magna of Japan. An unstated implication is that the major volcanic piles such as that at Yellowknife are directly related to crustal subduction. Others (McGlynn and Henderson, 1970; Baragar and McGlynn, 1976) have proposed completely ensialic models in which subduction mechanisms were not invoked.

Given that plate-tectonic processes occur at present, it is reasonable to consider the possibility that these processes or variations of them worked in the past, even during the Archaean. Burke et al. (1976) have suggested on theoretical grounds that such processes were likely, given the greater heat production during the Archaean due to radioactive decay and the fact that fast-moving spreading ridges can dissipate heat very efficiently. In the opinion of the writer, however, the geology of the Slave Province, at least as expressed in the supracrustal rocks, probably does not have a great deal to contribute towards the understanding of plate-tectonic processes in the Archaean although certain aspects of the geology are perhaps explicable as effects due to plate-tectonic processes.

What is the evidence that such processes were active in the Slave Province? As previously discussed the sediments and volcanics of the Slave Province appear to have accumulated on a sialic segment of Archaean crust that was at least as large as the present-day province. To date no simatic crustal remnants have been recognized. Folinsbee et al. (1968) have suggested that the volcanics at Yellowknife accumulated as an island arc on oceanic crust but subsequent work, particularly the discovery of much older tonalitic boulders in a diatreme within the volcanics and presumably derived from the underlying crust (Nikic et al., 1975), makes this interpretation unlikely. Thus the supracrustal rocks express the results of events occurring within sialic terrane within a hypothetical plate. It seems probable that the volcanics and sediments were deposited over a relatively short period of about 10 or 15 million years in fault-bounded basins or a basin.

If indeed sedimentation and volcanism took place in similar environments at the same time over a relatively short period of time throughout the Slave Province, then a tectonic mechanism involving regional extension of the crust is required that would be active over the entire province at about the same time. Interaction between hypothetical Archaean plates may be such a mechanism. Molnar and Tapponnier (1975), for example, have suggested that the effects of collision between India and Asia are expressed in part as intraplate deformation such as the Shansi graben system (which occupies an area similar in size to the Slave Province) some 1300 km northeast of the Himalayas. On the other hand, the Basin and Range Province of the southwestern United States, an area of Late Cenozoic regional extension several times larger than the Slave Province, consists of a complex system of normal faults that has resulted in a series of uplifted and downdropped blocks. As reviewed by Stewart (1978) this terrane has been variously described as being due to wrench faulting, back-arc spreading, subduction of the East Pacific Rise and mantle plumes.

Clearly a variety of processes can result in regional crustal extension. Recognition of the particular process involved in causing the regional extension in the Slave Province is made even more difficult as it appears at this time to have been preserved out of context with any plate system that may have been extant at the time of extension. This may be resolved to some degree in the future with a better understanding of the Archaean geology of adjacent shield areas — in particular the western Churchill Province.

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This paper could not have been written without the benefit of discussions over the years with those involved with mapping the geology of the Slave Province and others interested in its evolution. In particular I would like to acknowledge G. H. Eisbacher, R. A. Frith, M. B. Lambert, J. C. McGlynn, Mikkel Schau, and P. H. Thompson. This paper has benefited from careful reviews by D. Bridgwater, I. F. Ermanovics, T. E. Krogh, A. Kröner, J. C. McGlynn, Mikkel Schau and P. H. Thompson.

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Chapter 10

TECTONICS AND PETROGENESIS OF EARLY PRECAMBRIAN COMPLEXES OF THE ALDAN SHIELD, SIBERIA

V. M. MORALEV

ABSTRACT

The early Precambrian of the Aldan Shield can be subdivided into at least two major tectonic units: (1) a gneiss-granulite basement; (2) greenstone and schist belts (trough structures). Basement complexes were formed in two stages: (1) accumulation of ultramafic volcanic and volcano-sedimentary series (proto-ophiolite associations); and (2) transformation of basic protocrust with the emplacement of granites, growth of granite-gneiss domes or ovals (ring structures) and development of interoval granulite-charnockite-anorthosite belts. These stages of nonlinear tectonics coincided with the end of the permobile regime of the earth's crust after which the first relatively rigid microplates were formed. Greenstone belts younger than about 2.6 Ga appear to have developed from rift (trough) depressions, similar to minor oceans at the initial stages of evolution, which subsequently developed not as spreading basins, but as marginal back-arc basins. The closing of these basins took place simultaneously with the continuing growth of granite-gneiss domes and vertical accretion of lithosphere, i.e. the tectonic conditions of primitive horizontal microplate motion were coupled with partial preservation of a permobile (nonlinear) geodynamic environment. Granitization and alkaline metasomatism (with a maximum at about 2 Ga ago in the Aldan Shield) resulted in folding and thrusting in greenstone belts, thereby obliterating the unconformity between the basement and greenstone sequences. This process also resulted in a total cratonization of the shield, i.e. it led to the formation of the Siberian rigid continental lithospheric plate. The time interval of development of greenstone belts is unique in the earth's history by its double-faced character of geodynamic conditions transitional from a permobile regime to plate tectonics.

INTRODUCTION

The Aldan Shield, taken together with the Stanovoi Ridge, is the largest early Precambrian metamorphic terrain in Siberia. Katarchaean ($>$ 3.5 Ga), Archaean and lower Proterozoic rocks exposed there belong to two major tectonic units: gneiss-granulite basement and greenstone or schist belts (trough structures). Younger Upper Proterozoic and Phanerozoic rocks form a sedimentary cover (Fig. 10-1).

STRATIGRAPHY

The early Precambrian complexes of the Aldan Shield are composed of thick sequences of metamorphic rocks, subdivided into different local

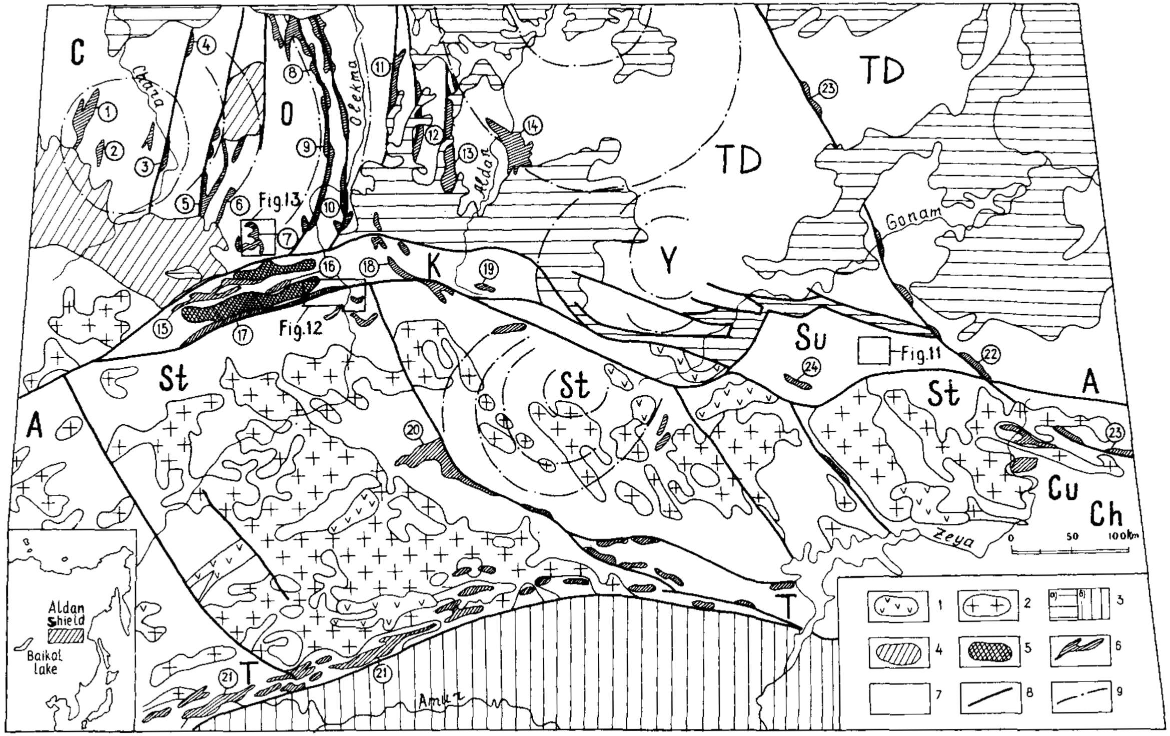

Fig. 10-1. Simplified geological map of the Aldan Shield and the Stanovoi Ridge. Explanation of legend: *1* = Cretaceous volcanics; *2* = Jurassic—Cretaceous granitoids; *3* = Phanerozoic and Upper Precambrian formations: *a* = platform cover, *b* = formations of the Mongol—Okhotsk foldbelt; *4* = Lower Proterozoic (Udokan) platform cover; *5* = Kalar Massif of gabbro-anorthosites; *6* = Upper Archaean—Lower Proterozoic greenstone and schist belts; *7* = gneiss-granulite complexes (series) of the basement (*O* = Olekma, *C* = Chara, *K* = Kurulta, *Y* = Yengra, *TD* = Timpton-Dzheltula, *Su* = Sutam, *B* = Batomga, *Ch* = Chogar, *Cu* = Cupura, *St* = Stanovoi); *8* = major faults (*A-A* = Stanovoi, *T-T* = Tukuringra); *9* = elements of ring structures as seen on satellite imagery. Encircled figures denote greenstone and schist belts: *1* = Saimagan; *2* = Torochan; *3* = Chara; *4* = Imalik—Tarinak; *5* = Khani—Olonda; *6* = Itchilyak (Evonokit); *7* = Taragai—Khain; *8* = Temulyakit; *9* = Tasmielin; *10* = Tungurchin; *11* = Borsala; *12* = Subgan; *13* = Amedichi; *14* = Ungra; *15* = Ayan—Burpala; *16* = Imangra; *17* = Kalar; *18* = Kurtakh; *19* = Chulman; *20* = Dzheltulak; *21* = Amazar—Gilyui; *22* = Avgenkur; *23* = Tirkanda; *24* = Lower Dzhelinda.

TABLE 10-I

Proposed correlation of the early Precambrian units

	Aldan Shield				Stanovoi ridge	
	West		Centre	East		
Lower Proterozoic	Udokan series			Ulkan series	Dzheltulak series	
	Tasmielin series		Subgan series	Avgenkur series	Ust-Gilyui series	
2500 Ma Upper Archaean	Borsala series					
	Olekma series	Chara complex	Timpton—Dzheltula series	Batomga series	Cupura series	Stanovoi complex
3000 Ma Lower Archaean	Kurulta series		Yengra series	Sutam series	Kurulta—Gonam (Chogar) series	
3500 Ma Katarchaean						

units (series) whose proposed correlation is given in Table 10-I. Katarchaean and Archaean units of the Aldan Shield are represented by high-grade gneisses and granulite rocks while late Archaean and early Proterozoic series comprise a number of isolated belts built of schists and greenstones, metamorphosed under amphibolite and greenschist facies conditions.

Gneiss-granulite basement

The Yengra and Timpton-Dzheltula series of the central part of the Aldan Shield as well as their analogues in other regions belong to metamorphic complexes which define a gneiss-granulite basement. The oldest horizons of these sequences are known as the Kurulta-Gonam series of the Stanovoi Ridge. The composition of the Kurulta-Gonam series and its equivalents (the Sutam and Chogar series) in different regions is strikingly similar. Basic schists (mainly hypersthene and two-pyroxene-plagioclase schists) with measured thicknesses of up to 5 km are always observed in the lower parts of the series. These are followed by a suite of various gneisses and schists, about 2—3 km thick, starting with garnet and sillimanite gneisses and granulites, beds of quartzite and lenses of corundum rock (Zverev suite). The uppermost section of the stratigraphic column is composed of pyroxene, biotite-pyroxene, biotite-amphibole and other gneisses and schists. The total thickness of the series is 9—10 km.

In the western part of the Aldan Shield hypersthene, biotite-amphibole-hypersthene, amphibole-two-pyroxene schists and gneisses occur at the base of the Kurulta series. The exposed thickness of these successions is not more than 3 km. This suite of basic schists is overlain by high-alumina gneisses and schists with intercalations of quartzites as well as biotite-amphibole and amphibole gneisses of the Olekma series, about 9—12 km thick.

In the centre of the Aldan Shield basic schists (the Gorbylyakh suite), overlain by quartzites and high-alumina gneisses (the Upper Aldan suite), make up the lower part of the Yengra series which is up to 6 km thick. Basic schists with intercalations of carbonate rocks and quartzites (the Fedorov and Idzhak suites) constitute the upper units. The total thickness of the series is about 10—13 km.

In the central and eastern parts of the Aldan Shield the rocks of the Yengra series are overlain by units of the Timpton-Dzheltula series. It is composed of biotite, biotite-garnet, biotite-amphibole gneisses, pyroxene-plagioclase schists, subordinate intercalations of marbles and calciphyres, with a total thickness amounting to about 20 km.

At the extreme east of the Aldan Shield rocks of the Batomga series are exposed. The series consists mainly of biotite-amphibole gneisses with marble lenses, the thickness being 10—12 km.

Correlation of the local Archaean units is extremely complicated because contacts between these successions are frequently faulted and all rocks are metamorphosed with various intensity. Rocks of the Yengra, Sutam and Kurulta-Gonam series are metamorphosed in the granulite facies, but in many regions (especially of the Stanovoi Ridge) they underwent retrograde metamorphism of amphibolite facies which is related to processes of later intensive granitization. It is suggested that the oldest metavolcanics form a basic basement which was generated during the early (Moon-like) stage of the evolution of the earth's crust (Pavlovsky, 1970). A thick suite of overlying high-alumina gneisses and quartzites is a good marker for regional stratigraphic correlation. This unit is considered to consist of metasedimentary rocks which formed due to weathering and desintegration of a mafic basement (Sidorenko, 1975). The overlying Timpton-Dzheltula gneisses are composed of both metasedimentary and metavolcanic rocks and possibly rest unconformably on the Yengra series.

Greenstone and schist belts

At present more than 30 isolated greenstone and schist belts (trough structures) are known within the Aldan Shield and the Stanovoi Ridge. They are mainly confined to large fault zones and are composed of volcano-sedimentary rocks and iron-chert formations, progressively metamorphosed under amphibolite or greenschist facies conditions. According to the proportion of metavolcanic and metasedimentary rocks volcanogenic (more than 50% metavolcanics), terrigene-volcanogenic (from 10 to 50% metavolcanics) and terrigene (less than 10% metavolcanics) belts can be distinguished.

Volcanics are represented mainly by rocks of basic composition. In one case (Khani-Olonda belt) a bimodal volcanic suite is recognized. Sedimentary rocks are represented by metasandstones, quartzites, mica-amphibole-cordierite and garnet schists, iron quartzites and, rarely, by graphite schists, marbles and metaconglomerates. Pebbles of gneisses, quartzites and iron quartzites are observed in the metaconglomerates.

In the volcano-sedimentary sequences metavolcanics occur mainly in the lower parts of the stratigraphy. In metasedimentary series horizons of conglomerates are often observed in the middle and upper parts of a succession. The Amedichi trough may serve as an example. Here the oldest horizons of the series are represented by biotite and biotite-garnet-cordierite microgneiss and schist. They are overlain by a pile of quartzites with intercalations of biotite-cordierite, biotite-amphibole, sillimanite and andalusite schist and metadiabase. A number of conglomerates (up to 250 m thick) occur higher in the stratigraphy, overlain by a suite of quartzites, micaceous quartzites and hematite quartzites. A pile of quartzites, amphibolite, biotite schists with horizons of marbles, metarhyolites and metadacites constitute the top of the sequence. The total thickness of the series is 5200—6400 m (Reutov, 1978).

A decrease in metamorphism from bottom to top of most sequences is characteristic of the greenstone and schist series. The lower horizons of the series are often migmatized and metamorphosed under amphibolite facies conditions while rocks of the upper members are usually altered under epidote-amphibolite and greenschist facies conditions.

In some belts the Dzheltulak series of sandstones (including copper sandstones), mica and graphite schists, gravellites, conglomerates, volcanics of intermediate and basic composition, rests unconformably on greenstone and schist series. Pebbles of granites, pegmatites, quartzites, amphibolites and quartz are observed in conglomerates of the Dzheltulak series. In lithology this series is similar to the Udokan series of the western part of the Aldan Shield, which belongs to a platform (protoplatform) cover.

GEOCHRONOLOGY

Many age determinations were carried out by various methods on Precambrian rocks of the Aldan Shield. The most representative geochronological data are given in Table 10-II. Most age measurements were obtained on young granites, pegmatites and metasomatic rocks, widely distributed both in the gneiss-granulite basement and in the greenstone belts. The data obtained show that the most intensive granitization of the early Precambrian complexes of the Aldan Shield took place between 1800—2100 Ma ago. These tectonic-magmatic processes resulted in widespread retrograde metamorphism of amphibolite facies grade in rocks of the gneiss-granulite basement as well as in progressive metamorphism in rocks of the greenstone and schist belts.

There are few isotopic data for rocks from greenstone belts, since these units have only been recognized and mapped recently as specific geological complexes. Available K-Ar ages on hornblende (2020 and 2670 Ma) are possibly not quite reliable but they still identify the probable early Proterozoic or late Archaean age of amphibolites in greenstone belts. An age determination of 2300 Ma on a pegmatite pebble from a post-greenstone conglomerate of the Dzheltulak series in the Stanovoi Ridge is of great interest since this age indicates a probable early Proterozoic—late Archaean ($>$ 2300 Ma) age for greenstone and schist series of the Stanovoi Ridge and the Aldan Shield.

Footnote to Table 10-II

Decay constants accepted in the USSR and used in calculating the ages are as follows: $^{40}K = 0.557 \times 10^{-10}$; $^{87}Rb = 1.39 \times 10^{-11}$; $^{238}U = 1.53 \times 10^{-10}$; $^{235}U = 9.72 \times 10^{-10}$; $^{232}Th = 4.88 \times 10^{-11}$. Isochron ages are marked by a cross, mineral ages based on one measurement only are marked by two crosses, other ages result from several measurements.

TABLE 10-II

Isotopic data for Precambrian rocks from the Aldan Shield (after Geochronology of the USSR, 1973; Rudnik and Sobotovich, 1969; Brandt et al., 1978)

Rock units	Rock type or mineral	Method	Age (Ma)
Granites and pegmatites	orthite	Pb/Pb	1700 ± 120^{+}
	orthite	Pb/Pb	1900 ± 85^{+}
	orthite	Pb/Pb	1950 ± 75^{+}
	granite	Pb/Pb	1880 ± 50^{+}
	granite	Pb/Pb	1970 ± 30^{+}
	granite	Pb/Pb	2130 ± 100^{+}
	granite	Pb/Pb	2180 ± 50^{+}
	zircon	Pb/U/Th	1860 ± 50^{+}
	microcline	Rb-Sr	1900 ± 120^{++}
	biotite	K-Ar	1950 ± 100
	biotite	Rb-Sr	1980 ± 100^{++}
	biotite	Rb-Sr	2110 ± 120^{++}
Metasomatic rocks	pyroxene schist	Rb-Sr	1980 ± 25^{+}
	phlogopite	Rb-Sr	2030 ± 90^{++}
	phlogopite	Rb-Sr	2145 ± 110
	phlogopite	K-Ar	1930 ± 60
	biotite	K-Ar	2050 ± 50
	diopside	K-Ar	1950 ± 50
	diopside	K-Ar	2150 ± 200
Greenstone and schist belts	pegmatites from pebble in conglomerates of Dzheltulak series	K-Ar	2300^{++}
	hornblende from amphibolite	K-Ar	2020^{++}
	hornblende from amphibolite	K-Ar	2390 ± 80
	hornblende from amphibolite	K-Ar	2670^{++}
Granites, pegmatites and metasomatic rocks	granite	Pb/Pb	2200 ± 200^{+}
	orthite	Pb/Pb	2600 ± 100^{++}
	pyroxene schist	Rb-Sr	2248 ± 39^{+}
	phlogopite	Rb/Sr	2260 ± 130^{++}
	muscovite	K-Ar	2520 ± 20
	biotite	K-Ar	2315 ± 35
Timpton-Dzheltula series	marble	Pb/Pb	2330 ± 50^{+}
	marble	Pb/Pb	2660 ± 80
	pyroxene schist	Pb/Pb	2330 ± 150^{+}
	pyroxene schist	Pb/Pb	2600 ± 200
Yengra series	marble	Pb/Pb	3200 ± 500^{+}
	marble	Pb/U	3180 ± 330^{+}
	pyroxene schist	Pb/Pb	3300 ± 200^{+}
	pyroxene schist	Rb/Sr	3960 ± 35^{+}
	charnockite	Th/Pb	4000 ± 1000^{+}
	diopside	K-Ar	4200 ± 500^{+}

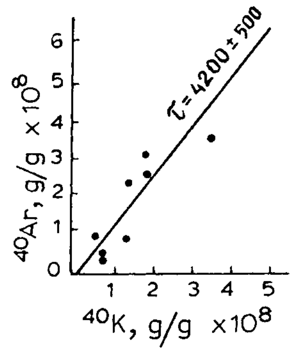

Fig. 10-2. K-Ar errorchron for diopside from Yengra series rocks (after Geochronology of the USSR, 1973).

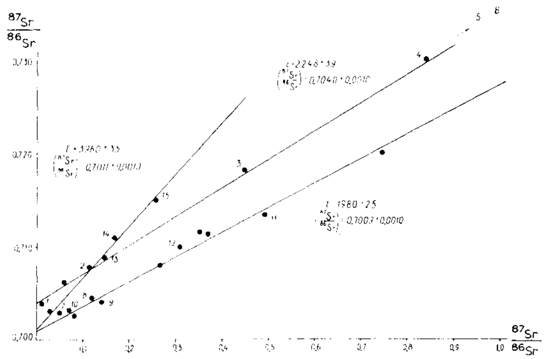

Fig. 10-3. Rb-Sr plot for Yengra series rocks (after Brandt et al., 1978). Numbers refer to data of Table 10-III.

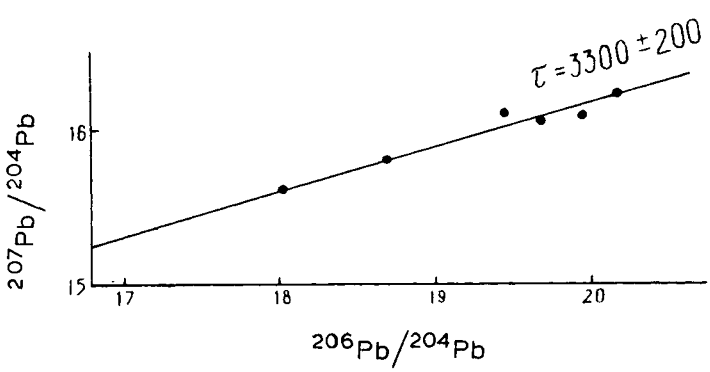

Fig. 10-4. Pb-Pb isochron for Yengra series schists (after Rudnik and Sobotovich, 1969).

It is probable that deposition of the volcanic and sedimentary rocks of the greenstone and schist series took place during the period from 2700—2600 to 2200—2100 Ma ago. There are also few age determinations for old granites whose injection terminated the formation of the gneiss-granulite basement and preceded the generation of greenstone belts. The existence of these granites is obvious as they are observed in pebbles of conglomerates of schist and greenstone series which occur together with the gneisses. Some isotopic data given in Table 10-II show a wide age range for rocks of this group (from 2200 to 2600 Ma). Geological considerations suggest these rocks to be as old as about 2400—2500 Ma.

There are many isotopic determinations for rocks of the gneiss-granulite basement (the Timpton-Dzheltula and Yengra series). The oldest ages from 3500 to $>$ 4000 Ma were obtained by the K-Ar method on diopside. The K-Ar errorchron trend for diopside indicates an age of 4200 ± 500 Ma (Fig. 10-2, after Anonymous, 1973, Geochronology of the USSR). However, Rb-Sr and Th-Pb data on charnockites and pyroxene schists of the Yengra series scatter widely as exemplified by the Rb-Sr plot of Fig. 10-3 (after Brandt et al., 1978) and the geological significance of corresponding errorchron "ages" of 3960 ± 35 Ma and 4000 ± 1000 Ma, respectively (see Table 10-II; Brandt et al., 1978) is not certain. All that can be said is that a fairly reliable Pb-Pb isochron age of 3300 ± 200 Ma (Fig. 10-4, after Rudnik and Sobotovitch, 1969) sets a younger age limit to the gneiss-granulite basement as represented by the Yengra series, but the presence of older rocks in the Aldan Shield is not ruled out (Rudnik and Sobotovitch, 1969).

Available radiometric data also suggest an age gap between the Yengra and Timpton-Dzheltula series (Table 10-II) during which regional metamorphism and the emplacement of granitoids of the charnockite and enderbite type may have occurred some 3000—3200 Ma ago.

The granitic plutonism which preceded the generation of greenstone belts is another significant tectono-magmatic event dating back to 2600 ± 100 Ma. The last and the most intensive granitization of early Precambrian complexes and progressive metamorphism of rocks of greenstone and schist series occurred during the period 1800—2100 Ma ago.

PETROGENESIS

The metamorphic rocks of the early Precambrian complexes of the Aldan Shield consist of metasediments and metavolcanics. In the gneiss-granulite basement metasedimentary rocks are represented by biotite, biotite-garnet, sillimanite-garnet gneisses, garnet-granulites, quartzites, micaceous and iron quartzites, marbles and corundites. Rocks of basic composition such as pyroxene, garnet-pyroxene and two-pyroxene-plagioclase schist originated from basic metavolcanics. Rather large (up to 1—2 m) inclusions of lherzolite as well as lenticular bodies of melanocratic sapphirine-bearing rocks are locally observed in them (Glukhovsky et al., 1977).

Metamorphic conditions of the oldest Precambrian complexes of the Aldan Shield are determined by mineral associations and are as follows: the Chogar series $P = 10$—11 kb, $T = 1000$—$1100°C$; the Sutam series $P = 9.5$—11 kb, $T = 820$—$1030°C$; the Timpton-Dzheltula series $P = 9.0$—9.5 kb, $T = 820$—$900°C$ (Kastrykina and Karsakov, 1977).

In the low-grade metavolcanics of greenstones and schist belts relic porphyritic and diabase textures are observed at places; in most cases, however, such relics are absent and the rocks are transformed into amphibolite, amphibole, chlorite-amphibole, talc-chlorite-actinolite and chlorite schists. Among these rocks there are lenses and layered bodies of altered ultramafics, represented by chlorite-serpentine, chlorite-anthophyllite, anthophyllite-talc, chlorite-talc-serpentine-carbonate rocks and by serpentinites.

Metasedimentary rocks of the greenstone and schist belts are represented by metapelites and metapsammites, transformed into mica schists and microgneisses containing garnet, sillimanite, andalusite, staurolite, as well

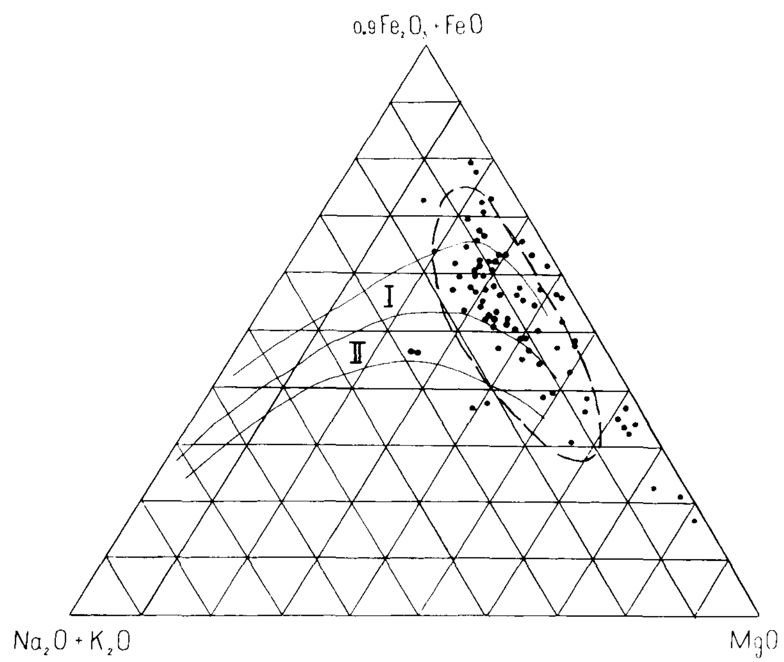

Fig. 10-5. AFM diagram for basic schists (metavolcanics) of the gneiss-granulite basement. Solid lines outline volcanic series of H. Kuno: I = tholeiitic; II = alkali olivine basalts (calc-alkali); broken line outlines the field of mid-ocean ridge basalts.

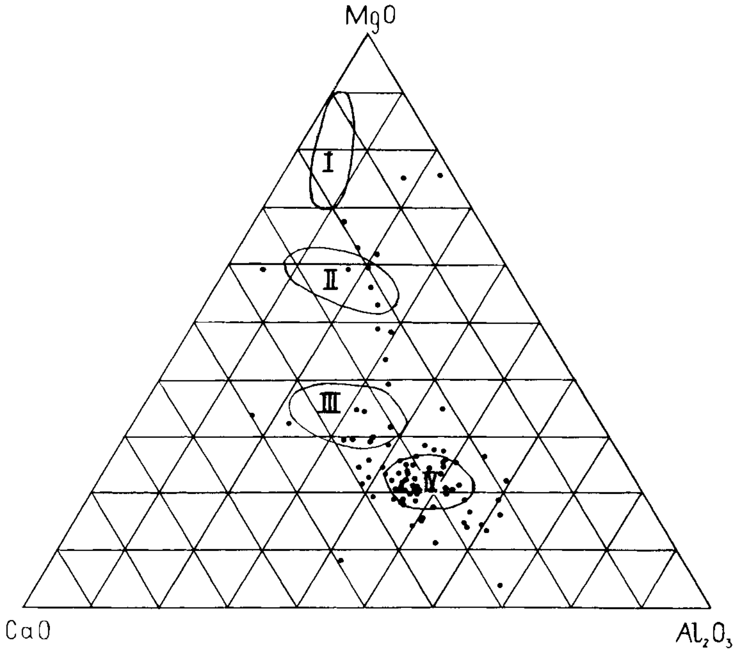

Fig. 10-6. CaO-MgO-Al_2O_3 diagram for basic schists (metavolcanics) of the gneiss-granulite basement. *I* = peridotitic komatiite; *II* = basaltic komatiite of "Geluk type"; *III* = basaltic komatiite of "Barberton and Badplaas type"; *IV* = tholeiitic basalts.

as by quartzites, iron quartzites, graphite schists, marbles and, locally, by conglomerates.

It is well known that certain chemical characteristics of the tholeiitic and calc-alkaline magmatic series are good indicators of the tectonic conditions under which the volcanogenic and volcano-sedimentary complexes may have formed. Therefore, petrochemical data on the Aldan metavolcanics are of primary interest for the purposes of this paper.

The chemical composition of basic metavolcanics from the Yengra, Timpton-Dzheltula, Kurulta-Gonam, Sutam and Chogar series (86 samples) are plotted on AFM and CaO-MgO-Al_2O_3 diagrams (Figs. 10-5 and 10-6). Most data points on the AFM diagram are in the field of modern mid-ocean ridge basalts and in the field of the tholeiitic series according to H. Kuno. A number of points are located beyond the field of tholeiitic basalts near the Mg-Fe side of a ternary diagram that indicates the presence of high magnesium basalts and ultramafics. It is obvious that these rocks do not belong to the calc-alkaline series. The CaO-MgO-Al_2O_3 diagram (according to Viljoen and Viljoen, 1969) shows that, besides tholeiitic basalts in the tested group, basaltic komatiites and ultramafics of peridotitic komatiite

TABLE 10-III

Rb-Sr isotopic data and isochron ages of selected rock types from the Timpton-Dzheltula series (after Brandt et al., 1978)

Rock type	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Calculated age and initial ratio
1. Pyroxene-plagioclase schist	0.0126	0.7042	
2. Diopside schist with scapolite	0.1476	0.7088	
3. Hornblende-biotite schist	0.4570	0.7181	2248 ± 39 Ma
4. Hornblende-biotite schist	0.8426	0.7307	0.7040 ± 0.0010
5. Hornblende-biotite schist	1.2732	0.7436	
6. Hornblende-biotite schist	2.9636	0.7972	
7. Diopside-hornblende schist	0.0535	0.7030	
8. Pyroxene-hornblende schist	0.1208	0.7045	
9. Pyroxene-hornblende schist	0.1409	0.7043	1980 ± 25 Ma
10. Pyroxene-hornblende schist	0.0705	0.7031	0.7009 ± 0.0010
11. Phlogopite-diopside schist	0.3121	0.7100	
12. Hornblende-biotite schist	0.4950	0.7136	
13. Hypersthene plagiogneiss	0.1180	0.7078	3960 ± 35 Ma
14. Pyroxene-hornblende schist	0.1705	0.7110	0.7011 ± 0.0010
15. Pyroxene-biotite schist	0.2486	0.7152	

The numbers refer to samples shown in Fig. 10-3 and all data presented here were assigned by Brandt et al. (1978) to the regression lines shown in Fig. 10-3.

chemistry are also observed. The presence of lherzolite inclusions in pyroxene schists characterized by their affinity with basaltic and peridotitic komatiites confirms an assumption that komatiite melts were derived from a mantle composed of pyroxene lherzolite (Cawthorn and Strong, 1974). For the rocks of the Sutam series derivation of komatiites and lherzolites from a common source have been noted (Glukhovsky et al., 1977).

Low (0.7009—0.7040) initial $^{87}Sr/^{86}Sr$ ratios (Table 10-III, after Brandt et al., 1978) are characteristic of basic metavolcanics which are weakly altered by processes of granitization and metasomatism. The $^{87}Sr/^{86}Sr$ ratios rise considerably (0.7152—0.7972) in rocks subjected to alkaline metasomatism when biotite or phlogopite (Table 10-III) appear in noticeable amounts. With granitization the anorthite content in plagioclase decreases while that of iron in hypersthene increases (Kastrykina and Karsakov; 1977).

Data on strontium isotope ratios show that the schists of basic composition most likely represent mantle-derived magmatic rocks (volcanics) which were later subjected to alkaline metasomatism (Brandt et al., 1978).

A comparison of average compositions of the Sutam series rocks of the Aldan Shield with rocks of recent oceanic crust (Table 10-IV, after Glukhovsky et al., 1977) shows considerable agreement in major element patterns as well as in concentrations of some trace elements such as Cr, Ni,

TABLE 10-IV

Average composition for rock types of the Sutam series (after Glukhovsky et al., 1977).

	1	2	3	4	5
SiO_2	44.2	44.69	48.90	43.95	49.34
TiO_2	0.3	0.78	1.25	0.10	1.49
Al_2O_3	4.5	10.23	14.25	4.82	17.04
Fe_2O_3	4.1	2.91	2.80	2.20	1.99
FeO	7.1	9.34	9.0	6.34	6.82
MnO	0.2	0.28	0.30	0.19	0.17
MgO	36.7	17.57	7.70	36.81	7.19
CaO	2.3	10.27	10.30	3.57	11.72
Na_2O	0.4	0.80	2.00	0.63	2.73
K_2O	0.2	0.49	0.50	0.21	0.16
Li	9	4	7	4	6
Rb	4	6	14	0.5	1.6
Ba	25	48	114	8	23
Sr	11	26	178	–	130
Cr	3450	977	131	4400	303
Ni	1450	942	106	2500	114
V	112	152	346	47	314
Co	225	119	75	117	42
Number of samples	4	7	14		

1—3; the Sutam series rock types: 1 = lherzolites, 2 = metakomatiites (pyroxene schists and pyroxenites), 3 = metatholeiites (pyroxene-plagioclase schists); 4 = lherzolite (after Dely, 1910); 5 = average composition of tholeiite of the oceanic crust (after Engel et al., 1965). Values in percentage (SiO_2—K_2O) and in ppm (Li—Co).

V, Co. The basic metavolcanics of the early Precambrian of the Aldan Shield are slightly enriched in lithophyle elements, especially in Rb, Sr and Ba, due to either a high content of these elements in the parental magma or their influx during the process of alkaline metasomatism.

Chemical analyses of basic metavolcanics (45 samples) from greenstone belts of the Aldan Shield are plotted on MgO-Na_2O, AFM and CaO-MgO-Al_2O_3 diagrams (Figs. 10-7, 10-8 and 10-9). The MgO-Na_2O binary diagram (Fig. 10-7) reveals that the analyses have clear affinities with basalt (III) and basaltic komatiite (II) in many cases. A few volcanics plot within the field of peridotitic komatiites (I) and andesites (IV). AFM and CaO-MgO-Al_2O_3 diagrams (Figs. 10-8 and 10-9) show that the greenstone metavolcanics mostly plot as mid-ocean ridge basalts or as tholeiitic basalts. The presence of basaltic komatiites as well as ultramafics with peridotitic komatiite chemistry can also be seen. The AFM diagram clearly reveals that the basic metavolcanics of greenstone belts partly belong to the calc-alkaline series of H. Kuno. Using the discriminate function diagram (Pearce, 1976) allows to conclude (Fig. 10-10) that the greenstone metavolcanics

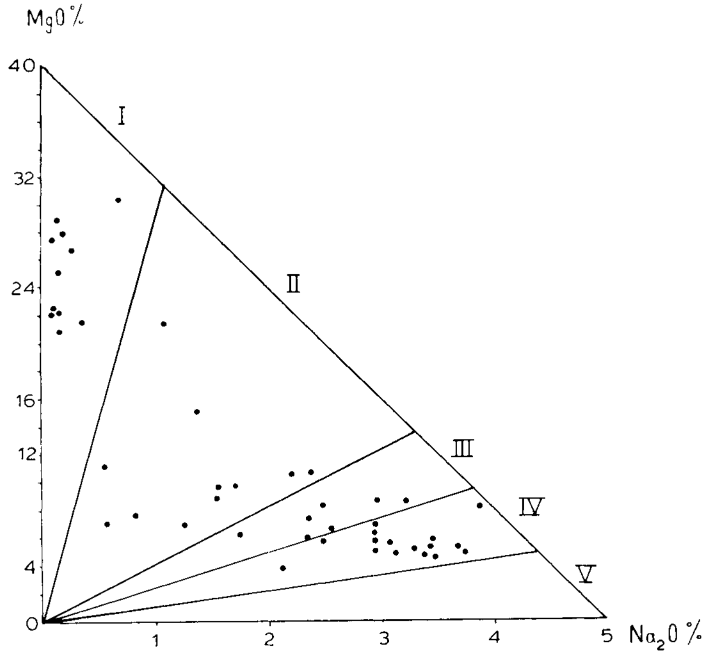

Fig. 10-7. MgO-Na$_2$O diagram for basic metavolcanics of various greenstone belts. *I* = peridotitic komatiite and basaltic komatiite of "Geluk type"; *II* = basaltic komatiite of "Barberton and Badplaas type"; *III* = basalts and gabbro; *IV* = andesite and diorite; *V* = rhyolite, rhyodacite, dacite, tonalite, granodiorite, granite.

have chemical affinities to both modern oceanic-floor basalts and volcanic island-arc series (calc-alkaline basalts, low-potassium tholeiites and shoshonite series). Therefore, the majority of the basic metavolcanics of greenstone belts could not have been generated directly from mantle-derived tholeiitic and komatiitic melts, in contrast to mafic rocks of the gneiss-granulite basement. The petrogenetic model for the generation of greenstone sequences must include magma generation mechanisms which first allow the formation of parental primitive rocks from which the calc-alkaline rock series can be generated, most likely under high pressure conditions from sources containing eclogite.

TECTONICS

Three cycles of deformation are recognized in the early Precambrian complexes of the Aldan Shield (Duk and Kitsul, 1975). Folds of the oldest cycle are recognized only in rocks of the Yengra series and its analogues. Foliation is related to this cycle.

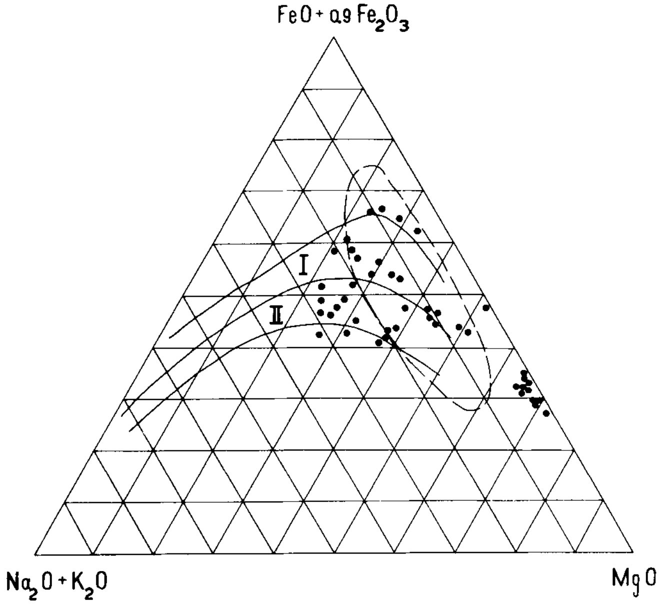

Fig. 10-8. AFM diagram for basic metavolcanics of various greenstone belts. Symbols as defined in Fig. 10-5.

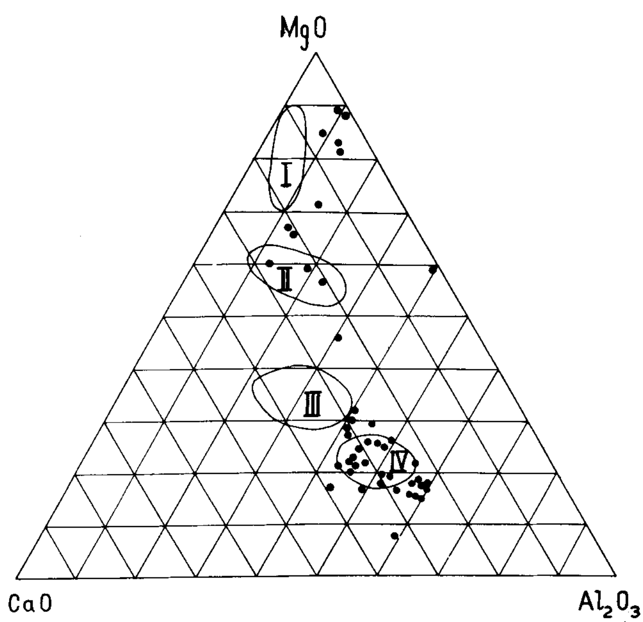

Fig. 10-9. CaO-MgO-Al_2O_3 diagram for basic metavolcanics of several greenstone belts. Symbols as defined in Fig. 10-6.

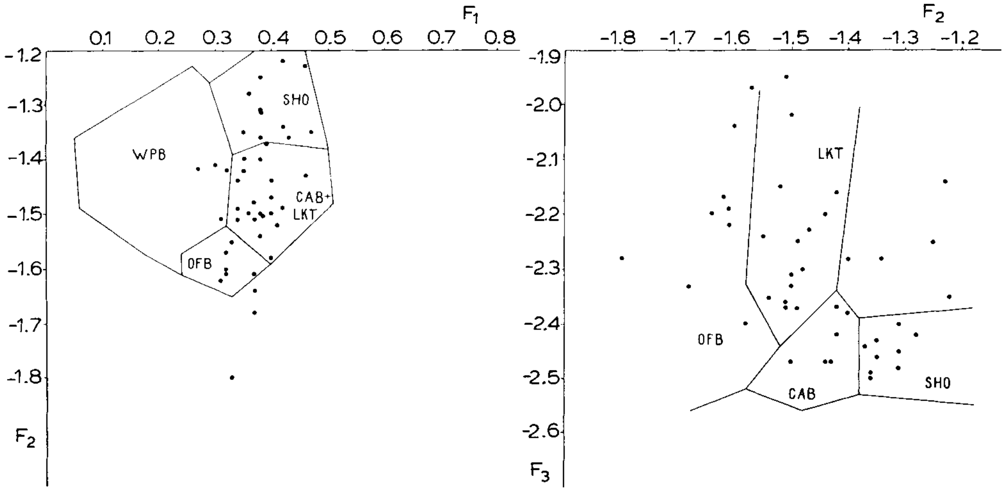

Fig. 10-10. Discrimination diagram according to Pearce (1976) for basic metavolcanics of Aldan greenstone belts. *OFB* = ocean-floor basalts; *LKT* = low-potassium tholeiites; *CAB* = calc-alkali basalts; *SHO* = shoshonites; *WPB* = within-plate basalts.

Folds of the first cycle are mostly obliterated by later deformation but in the Sutam and Kurulta-Gonam series broad synclinoria of various dimensions are locally observed. Large structures of such type (up to 130 × 80 km) are complicated by smaller variously oriented folds. Relatively gently dipping (25—45°) flanks of synclines consist of a system of recumbent folds, complicated by linear folds of later stages of deformation. Such a structure as exemplified by the Sutam syncline (Fig. 10-11) is described by Glukhovsky and Pavlovsky (1973). It is supposed that synclinal folds may be considered as remnants of older structural forms, generated during the early evolutionary stages under conditions of lacking horizontal tensions (non-linear tectonic style) in a thin and unconsolidated basic protocrust (Pavlovsky, 1970, 1975).

The general structure of the gneiss-granulite basement of the Aldan Shield is mostly controlled by folds of the second deformation cycle, which is characterized by the development of granite-gneiss domes or ovals, surrounded by ring structural interoval zones of compressed linear folds. The largest concentric ring structures were interpreted on satellite imagery (Glukhovsky, 1978). The most prominent elements of these structures are marked in Fig. 10-1. In the central portion of the Shield 10 such structures, 50—400 km in diameter, were recognized. Foliation in rocks of the gneiss-granulite basement is mostly subparallel to major lineaments which are observed on satellite imagery, and partly coincides with major faults and greenstone belts.

It is suggested that the formation of granite-gneiss domes or ovals may

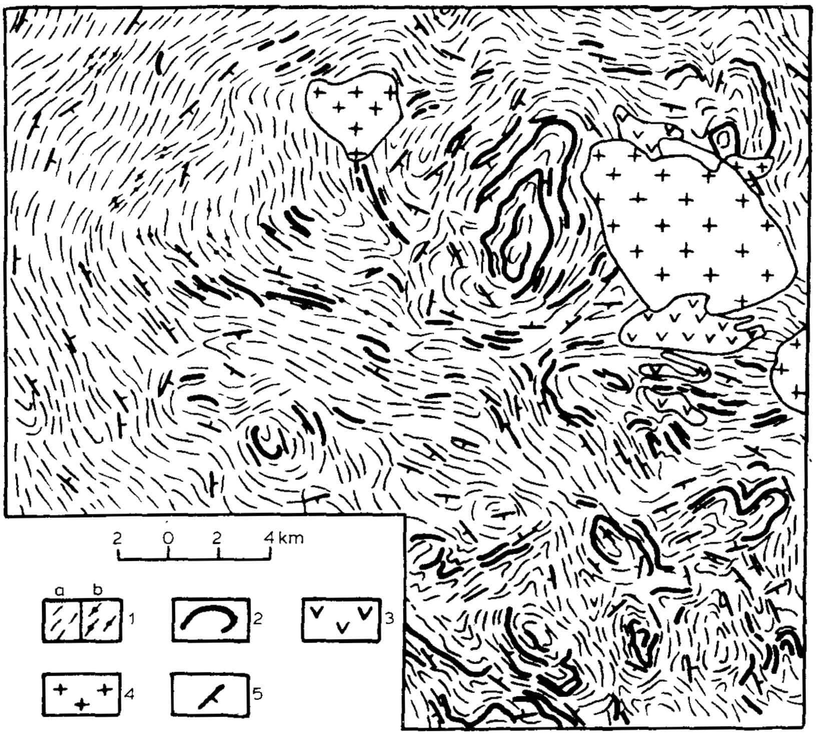

Fig. 10-11. Sketch-map of the Sutam Syncline (after Glukhovsky and Pavlovsky, 1973): *1* = foliation of the rocks of granulite facies: *a* = pyroxene schists (metavolcanics), *b* = alumina schists (metapelites); *2* = iron quartzites; *3* = gabbro-anorthosites; *4* = Jurassic—Cretaceous granites; *5* = foliation dip.

be related to the processes of granitization and a concomitant decrease in the density of the basic protocrust. This suggestion agrees with the evaluation of the distribution of upper-crustal rock densities, which have been obtained from residual gravity anomalies calculated up to 15 km depth, the assumed density of the intermediate layer being 2.3 g/cm^3. Weakly granitized high density rocks of the gneiss-granulite basement (with a density of 2.95—3.5 g/cm^3) were defined by the dimension of positive residual gravity anomalies contoured by + 5 mgal. The areas of ring structures were divided by concentrical circles into inner (from the centre up to 2/3 of radius) and outer (from 2/3 up to 4/3 of radius) zones. Calculated areas of residual gravity anomalies more than 5 mgal are given in Table 10-V.

As is obvious from Table 10-V the high density rocks are not found in the central zones of ring structures. Here only 2.5% of the total area of residual anomalies are located. More than 66% of the areas with positive anomalies are distributed within the outer zones of the ring structures.

TABLE 10-V

Positive residual gravity anomalies more than + 5 mgal (km^2) with respect to ring structure elements.

Ring structures	Inner zones	Outer zones	Zone of the Stanovoi Fault	Other areas
1. Chara	230	2900		
2. Aldan	—	1650		
3. Aldan-Uchur	220	2800		
4. Gonam	—	2500		
5. Uchur	150	1650		
6. Algoma	—	550		
7. Timpton	150	150		
8. Kalar	—	2400		
9. Njukzha	—	1200		
10. Zeya	—	4200		
Total (km^2)	750	20000	2400	6900
Total (%)	2.5	66.5	8	23

The Stanovoi fault undoubtedly follows the outer zones of these ring structures. It can be assumed that this largest fault of the Aldan Shield was generated in the early Precambrian as a system of joined external zones of ring structures. The total percentage of positive residual anomalies found in the outer zones of ring structures, taken together with the Stanovoi fault, amounts to 75%. These results favour the interpretation of ring structures as centers of granitization where the density of the basic protocrust is considerably decreased.

It can be assumed that the development of granite-gneiss ovals generated local lateral stresses in the outer zones of the ring structures. This process apparently resulted in high-pressure granulite facies metamorphism, basification and partial melting of protocrust rocks as well as in the generation of calc-alkaline magmas and the emplacement of gabbro-anorthosite complexes.

Fold deformation of the third cycle was the most important for the consolidation of the Aldan Shield. Greenstone and schist belts (trough structures) as well as numerous granite domes grouped in complex anticlinoria are attributed to structures of this cycle. Greenstone and schist belts are confined to regional faults of three systems. In the central and western parts of the shield the belts are of meridional trend, in the southern portion they trend northwest, while in the Stanovoi and Tukuringra fault zones their orientation is east—west. Some schist belts are located in curved faults of outer zones of ring structures.

The dimensions of greenstone and schist belts are usually 3—7 × 30—50 km, but larger belts (up to 10—20 × 150 km), as well as extended systems of belts, are also observed (Fig. 10-1). The volcano-sedimentary sequences of several belts differ considerably from each other, thereby suggesting that they were not linked during sedimentation. The greenstone and schist series are usually folded into narrow synclines cut by faults and turned into steeply dipping, monoclinal nappes.

Contacts of the greenstone and schist series with rocks of the gneiss-granulite basement are either faulted or migmatized by granitoids so that nobody has yet succeeded to observe any of the greenstone and schist sequences stratigraphically resting on gneiss-granulite basement rocks.

Greenstone and schist belts are mostly linear elongate structures, gradually thinning out along strike or sharply truncated by granite domes. On the eastern termination of the Kalar belt one can see such type of contact (Fig. 10-12). When trough structures are located in the outer zones of large

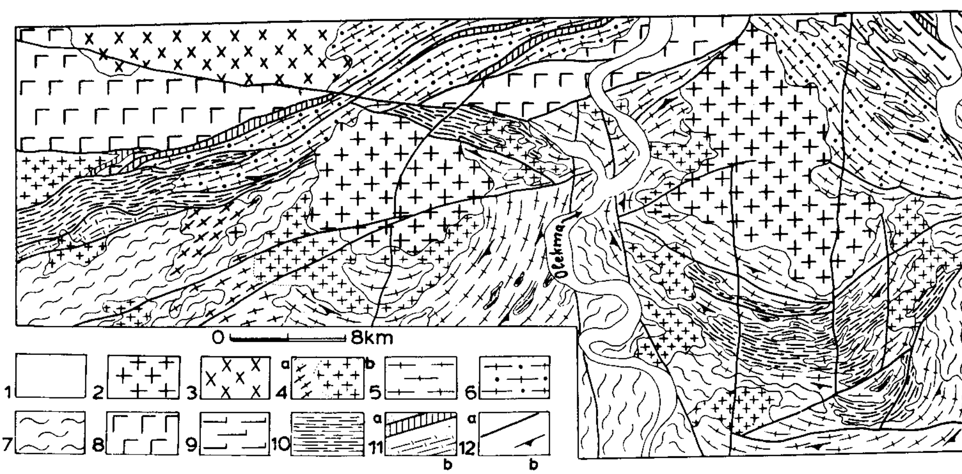

Fig. 10-12. Simplified structural map of the eastern portion of the Kalar greenstone belt: *1* = Alluvium; *2* = Jurassic—Cretaceous granites; *3—6* = Lower Proterozoic intrusive rocks: *3* = syenites, *4* = gneissic granites (*a*) and massive granites (*b*), *5* = granite-gneisses; *6* = garnet plagiogranites; *7* = gneissic complex of a basement; *8* = gabbro-anorthosites; *9* = gabbro; *10* = greenstone belts; *11* = deformed rocks: diaphthorites (*a*), blastomylonites (*b*); *12* = faults (*a*) and foliation dip (*b*).

granite-gneiss domes the greenstone and schist series often consitute lens-like monoclinal inliers within the granite-gneisses, foliation trends are bent according to a regional structure. Synclinal structures are preserved in some of these inliers, for instance in the Taragay-Khin belt (Fig. 10-13).

Numerous dykes and gabbro, gabbro-amphibolite and ultramafic intrusions are observed in the greenstone belts and in the granite-gneiss terrains bordering these belts. In some cases chains of mafic and ultramafic bodies located within zones of high strain are apparently relics of deeply eroded greenstone belts.

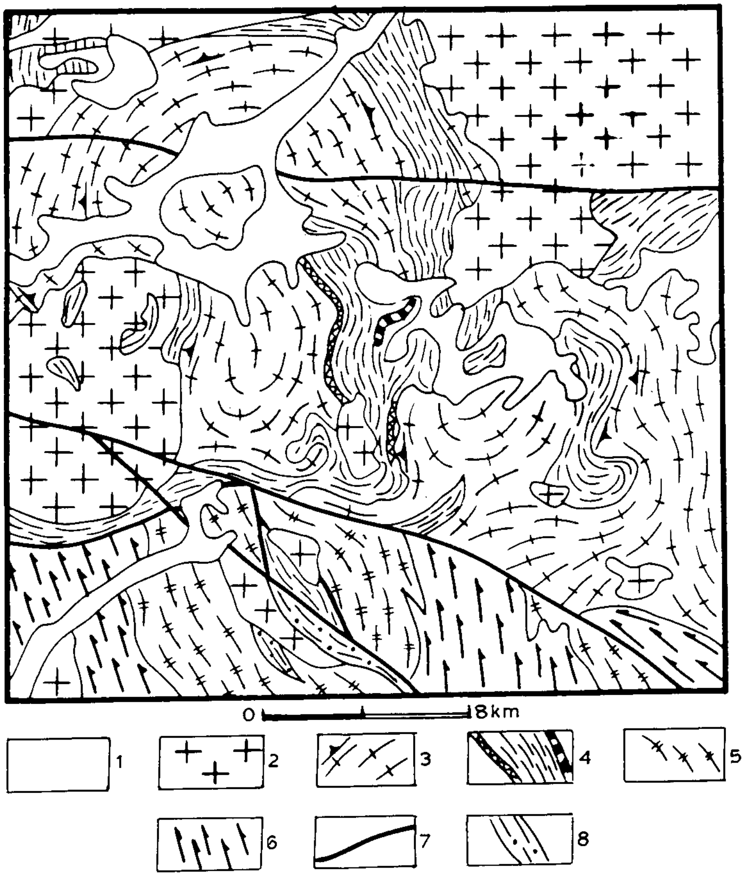

Fig. 10-13. Simplified structural map of the southern portion of the Taragai-Khain greenstone belt: *1* = Alluvium; *2* = Lower Proterozoic granites; *3* = granite-gneisses; *4* = greenstone belts (quartzite beds to the left, iron quartzite beds to the right); *5* = pyroxene plagiogranites; *6* = pyroxene schists (metavolcanics); *7* = faults; *8* = diaphthorites.

Late Precambrian and subsequent phases of deformation in the Aldan Shield and Stanovoi Ridge are evident in movements along the major faults, in injection of basic dykes, alkaline and granite intrusions, as well as in the development of mylonite and diaphthorite zones and in the formation of depressions filled with sediments.

CRUSTAL DEVELOPMENT

Available data on the geology and geochronology of the early Precambrian complexes of the Aldan Shield suggest that a gneiss-granulite sialic basement has already existed prior to the development of greenstone belts. The evolution of this basement started with the generation of mafic-ultramafic volcanic and volcano-sedimentary series. The presence of tholeiitic basalts, basaltic komatiites, ultramafics (lherzolites), together with cherty and pelitic metasediments (clean quartzites, iron quartzites, garnet-biotite-sillimanite schists), make it possible to compare these series with modern ophiolites and to consider them as proto-ophiolitic associations. However, these associations are not believed to be precise analogues of ophiolitic complexes of Phanerozoic oceans since constituent rocks differ from modern oceanic ones with respect to certain geochemical features and since they also originated in a different tectonic environment.

The petrogenesis of these ultramafic to mafic volcanic rocks, including basaltic and peridotitic komatiites, apparently reveals peculiarities of magma generation and tectonic conditions in the early Precambrian (Brooks and Hart, 1974). Partial melting of inhomogenous mantle occurred at that time at various depth simultaneously with multistage remelting of restites in ascending mantle diapirs (Cawthorn and Strong, 1974; Arth et al., 1977). The model of multistage melting (Arth et al., 1977) seems to provide a plausible explanation for the genesis of both the mafic to ultramafic and the silicic magmas (tonalites, granodiorites).

The presence of metasediments in the gneiss-granulite complexes testifies to breaks in volcanic activity, to intense weathering and desintegration of basic rocks and to accumulation of clastic debris in shallow sedimentary basins (Sidorenko, 1975).

The mafic to ultramafic rocks of the high-grade metamorphic complexes are comparable to a certain extent to the volcanic series of greenstone belts since both contain associations of primary magmatic rocks. Nevertheless, a very important distinction between them, reflecting the change in tectonic conditions, is the absence of rocks of the calc-alkaline series in the oldest (Katarchaean) complexes. They are characteristic of greenstone belt complexes, especially of the upper horizons of greenstone sequences.

Data available for the Aldan Shield do not allow to consider the evolution of the earth's crust of this region in terms of repeated generation of greenstone belts (Glikson, 1976). A thin continental gneiss-granulite crust was in existence prior to 2600—3000 Ma and passed at least through two stages of tectonic evolution:

(1) Generation of mafic to ultramafic volcanic and plutonic complexes with subsequent substantial addition of metasedimentary material.

(2) Thickening of this predominantly basic protocrust by addition of granites, growth of granite-gneiss domes or ovals (ring structures) and development of interoval granulite-charnockite-anorthosite belts.

These stages of nonlinear tectonic evolution correspond to the end of the permobile regime of the earth's crust when first comparatively rigid microplates were formed (Zonenshain et al., 1976; Shaw, 1976).

It can be assumed that granulite and amphibolite facies metamorphism developed simultaneously. High-pressure mineral assemblages of the granulite facies have been created in the outer zones of granite-gneiss ovals, whilst in the inner zones the influx of silicic magmas into basic protocrust material took place with the formation of tonalitic granite-gneisses. The growth of granitization centres was accompanied by amphibolite facies metamorphism which was superimposed on the granulite facies rocks.

The evolution of the late Archaean greenstone belts of the Aldan Shield began after a relatively rigid sialic crust had already formed. Trough structures probably developed as rift (trough) depressions similar to minor oceans (Windley, 1973; Hawkesworth and O'Nions, 1977). They did not develop as spreading basins, but as marginal back-arc basins (Zonenshain et al., 1976; Burke et al., 1976; Windley, 1976). During evolution of the trough structures lateral interactions of lithosphere microplates were coupled with processes of granite-gneiss dome generation and vertical accretion of the lithosphere, i.e. there was partial preservation of nonlinear geodynamic conditions (Zonenshain et al., 1976; Moralev, 1978). Processes of granitization and alkaline metasomatism continued simultaneously with the accumulation of volcano-sedimentary trough sequences in adjoining zones of the gneiss-granulite basement (i.e. in zones similar to active continental margins according to actualistic concepts). Such processes caused partial remobilization of the basement and injection of tonalite-granodiorite magmas. The most intensive granitization took place at the end of the early Proterozoic (about 2000 Ma ago). It caused folding and thrusting in greenstone belts, thereby obliterating the original unconformities between the basement and the greenstone sequences and also led to a total cratonization of the shield, i.e. the formation of the Siberian rigid continental lithospheric plate.

The period of greenstone belt development in the geological history of the Aldan Shield therefore represents a transition from a permobile regime to plate tectonic conditions. This double-faced character of the geodynamic environment of greenstone belts in the Aldan Shield and perhaps for other Archaean shields is unique in the earth's history and gave rise to peculiar features of tectonics, petrogenesis and metallogeny of trough structures.

CONCLUSIONS

The study of early Precambrian complexes of the Aldan Shield and the Stanovoi Ridge suggests that Katarchaean and early Archaean ultramafic to mafic rocks and paragneisses, enclosed in granite-gneiss and granulite complexes, may well be the oldest rocks of a basic protocrust which probably formed during the earliest stages of the history of the earth. This protocrust was transformed into the oldest continental crust by the growth of granite-gneiss domes or ovals and by the formation of interoval granulite-charnockite-anorthosite belts that were the oldest linear mobile belts. In the later geological history these belts appeared to be zones of high magmatic permeability and tectonic activity. The Stanovoi Ridge was such a zone where tectonic reactivation of Mesozoic age led to the emplacement of numerous granitoid batholiths.

The application of actualistic tectonic concepts to the epoch of nonlinear tectonics is practically impossible in spite of the fact that processes of petrogenesis of magmatic (proto-ophiolite) rock complexes must have been similar to a certain extent to those that led to the basalt generation in recent oceans. From the stage when the development of greenstone belts within the first relatively rigid microplates started, the application of modern plate tectonic concepts seems to be plausible. At that time processes of lateral and vertical accretion of microplates and their destruction with generation of greenstone belts occurred simultaneously. The double-faced geodynamic conditions of this stage of crustal evolution define its unique character in the earth's history as transitional from a permobile regime to modern plate tectonics.

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Chapter 11

PROTEROZOIC CHRONOLOGY AND EVOLUTION OF THE MIDCONTINENT REGION, NORTH AMERICA

W. R. VAN SCHMUS and M. E. BICKFORD

ABSTRACT

Early Proterozoic rocks of the midcontinent region consist of mafic to felsic volcanic rocks, thick sequences of metasedimentary rocks, and calc-alkaline plutons, all of varying age. In the Penokean Fold Belt the principal orogenic activity occurred about 1820—1860 Ma ago; about 80 Ma later a terrane of rhyolite and epizonal granite formed to the south of the Penokean rocks. The early Proterozoic basement of the western United States represents two distinct periods of orogenic development, 1690—1780 and 1610—1680 Ma ago, but units of Penokean age are apparently absent. In the central midcontinent area, the basement is composed of gneissoid granitic rocks and small volumes of metasedimentary and metavolcanic rocks. In the northern part of this region some of the rocks may be coeval, and possibly correlative, with either the 1690—1780 Ma old terrane of the Rocky Mountains or the 1820—1900 Ma old Penokean terrane (or both), but precise age confirmation is lacking. At least some of the granitic rocks in northern Kansas and Missouri formed about 1625 Ma ago and are thus coeval with plutonic and volcanic rocks in Arizona and New Mexico. Metamorphic effects of that age occur in the Penokean terrane, suggesting that the 1610—1680 Ma old belt extends as far east as the southern Great Lakes region.

A striking feature of the southern and eastern midcontinent region is a great terrane of rhyolite and epizonal granite that stretches from northern Ohio across Indiana, Illinois, Missouri, southern Kansas, and Oklahoma at least into the Texas panhandle. These rocks were formed in middle Proterozoic time, mostly in the interval 1380—1480 Ma ago.

The most important characteristics of the Proterozoic rocks of the midcontinent and their distribution with regard to the possible operation of plate-tectonic mechanisms in the Proterozoic in this region are: (1) the steady progression of younger and younger rocks southward from the Archaean craton of the Canadian Shield; and (2) the absence of typical island-arc rock assemblages in the northern midcontinent and the great abundance of granite and rhyolite in the southern midcontinent. We believe that these terranes were probably formed by convergent processes on the margin of the continent despite their lack of similarity with modern circum-Pacific rock assemblages.

INTRODUCTION

The midcontinent region of North America presents several interesting aspects of continental evolution and also presents particular questions regarding the role of plate tectonics during the Proterozoic. One distinctive feature of this region is the general progression from the Archaean craton of the Canadian Shield southward into progressively younger terranes. This apparent chronologic zonation has been recognized since the advent of routine geochronology in the 1950's and has been responsible for continental evolution models that involve some form of continental accretion (e.g. Engel, 1963; Tilton and Hart, 1963; Hurley and Rand, 1969). Although the general concept of continental accretion predates the plate-tectonic revolution of the late 1960's, it is easily adaptable to scenarios in which recurring subduction zone activity creates successively younger arc assemblages along a continental margin (Hurley and Rand, 1969; Engel et al., 1974).

In this paper we will evaluate the data available for the Precambrian basement of the midcontinent region of North America and present an updated summary of its major crustal terranes. Considerable progress has been made in understanding midcontinent basement terranes since the pioneering work of Muehlberger et al. (1966, 1967), Goldich et al. (1966a, b) and Lidiak et al. (1966), but the problem is still compounded by lack of exposures, lack of suitable basement samples, and lack of *precise* and *accurate* radiometric ages. In the latter case the problem exists even in several areas for which exposure is quite good. As a result, many of the conclusions or suggestions made in this paper must be considered tentative and subject to revision or rejection as additional data become available.

GEOCHRONOLOGIC DATA

One major problem in studying the Precambrian history of the midcontinent region is that of trying to correlate precisely (or show non-correlation of) major events. In the absence of biostratigraphic criteria, the main tool that must be used is radiometric dating, but not all radiometric data are suitable for these purposes. It is desirable to resolve age differences of 10 Ma or less for early and middle Proterozoic rocks in order to examine potential correlation adequately, and this requires precision as well as accuracy. The precision inherent in routine Rb-Sr and K-Ar age determinations is of the order of about 2 to 3%, or ± 30 to 60 Ma for the age range in question, which is considerably less than desired. Furthermore, it has been shown in numerous studies that K-Ar ages (especially) and Rb-Sr ages (even for "good" isochrons) can yield apparent ages that are significantly lower (younger) than true ages as inferred from U-Pb dating of zircons in the same region. Two prime examples where such discordance (Rb-Sr vs. U-Pb) is prevalent occur in the midcontinent region. These are the Lake Superior area (Van Schmus, 1976) and the St. Francois Mountains of southeast Missouri (Bickford and Mose, 1975). General consideration of the ages reported by Goldich et al. (1966a, b), Lidiak et al. (1966) and Muehlberger et al. (1966) indicates that many, if not most, of them are too low and should only be considered as minimum ages.

Extreme caution must be used in compiling chronologic information for use in regional studies; many recent attempts at regional synthesis (e.g. Emslie, 1978; Warner, 1978, 1979; Dutch, 1979) have, to a greater or lesser degree, grouped together lithologic units or terranes that represent distinctly (sometimes substantially) different ages. We have adopted the position in this report that the most reliable age information currently available is that obtained from U-Pb concordia intercepts for suites of cogenetic zircon samples (e.g. Silver and Deutsch, 1963). Not only is this method inherently more accurate, but it has the potential for yielding precision better than ~ 0.5% (± 5–15 Ma for the age range being studied). Consequently, age inferences in this paper will be based primarily on U-Pb ages of zircons. In a few instances, particularly for Colorado where U-Pb and Rb-Sr ages are commonly equivalent (e.g. Bickford et al., 1969; Barker et al., 1969), we will rely on Rb-Sr results to provide minimum ages. In other instances Rb-Sr and K-Ar data can provide useful insight into the metamorphic chronology of a region.

All ages used in this paper are based on the decay constants recently proposed for international adoption by the IUGS Subcommission on Geochronology (Steiger and Jäger, 1977) and have been recalculated from original references as necessary.

CONTINENTAL AND TECTONIC SETTING

The midcontinent region in the United States consists primarily of Phanerozoic sedimentary rocks that are underlain by early to middle Proterozoic igneous, sedimentary, and metamorphic rocks. These Proterozoic rocks crop out extensively in the western U.S. and the Great Lakes area, and they are sporadically exposed in the south-central U.S., primarily in southeastern Missouri (Fig. 11-1). Because most of the region is buried by younger rocks, it is necessary to rely heavily on information from surrounding exposed areas which can be combined with limited data from subsurface samples and isolated outcrops in order to interpolate geologic provinces across the entire region. A complete review of the geology of all these regions is beyond the scope of this paper, but some of the more important regional features are shown in Fig. 11-1.

The principal older crustal blocks to the north of the midcontinent region are the Superior Province of the Canadian Shield (Goodwin et al., 1972), its extension into the United States in the Lake Superior region (Sims, 1976), and the Wyoming block (Condie, 1976). These consist of a variety of rock types ranging from 2.5 to 3.5 Ga (or more) in age. For the most part these regions have acted as stable cratons during the Proterozoic, although the Archaean basement in Wisconsin, in the southern part of Upper Michigan, and in east-central Minnesota has been extensively involved in Proterozoic deformational and metamorphic events (Sims, 1976; Van Schmus, 1976; Van Schmus and Anderson, 1977). Morey and Sims (1976) recognized that this terrane is lithologically different from more common granite-greenstone assemblages of the Superior Province to the north and have delineated a

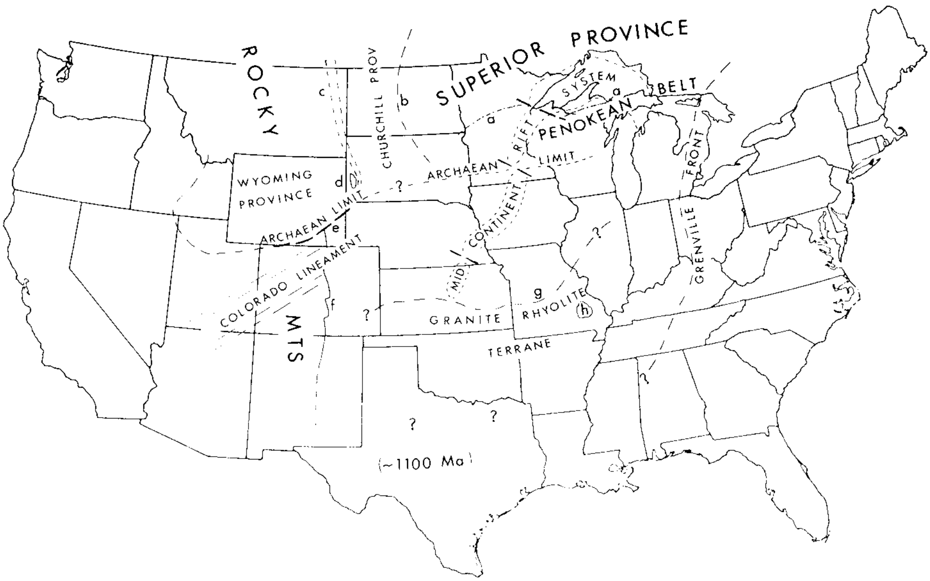

Fig. 11-1. Outline map of the United States showing major geologic features that relate to interpretation of the geology of the midcontinent region. Lettered features are: a = gneiss-migmatite, granite-greenstone terrane boundary of Morey and Sims (1976); b = Churchill—Superior Province boundary; c = North American Central Plains conductive anomaly of Alabi et al., 1975; d = Black Hills uplift, South Dakota; e = shear zone at southern margin of Wyoming Province; f = eastern edge of Rocky Mountains uplift; g = northern edge of southern midcontinent granite-rhyolite terrane; h = St. Francois Mts. uplift in southeast Missouri. Question marks (?) refer to uncertain locations or extensions of boundaries.

boundary between the latter and a "gneiss-migmatite" terrane to the south (a, Fig. 11-1).

Recent studies have shown that much of the Churchill Province of the Canadian Shield is underlain in part by Archaean rocks that were reworked during the "Hudsonian Orogeny" about 1.8 Ga ago (Davidson, 1972). The Churchill-Superior province boundary is inferred to extend into the subsurface of north-central North Dakota and southward to southeastern South Dakota (Lidiak, 1971; Green et al., 1979; feature b in Fig. 11-1) so that the Precambrian basement of western North and South Dakota may also consist of remobilized Archaean crust.

The southernmost occurrences of confirmed (or inferred) Archaean rocks, reworked or not, are in central Wisconsin (Van Schmus and Anderson, 1977), the Minnesota River Valley (Goldich et al., 1970) and the southern edge of the Wyoming block (Hills and Houston, 1979). These mark the minimum extent of the Archaean (Fig. 11-1) and a potential locus of a continental margin along which plate-tectonic processes could have operated. It is not known, however, whether there was a continental margin there at the beginning of Proterozoic time, 2.5 Ga ago, whether one was subsequently created by a later rifting event, or even whether Archaean crustal rocks extended (and may still extend) far to the (present) south.

A prominent structural feature along the southern edge of the Wyoming block is the Cheyenne Belt (Houston et al., 1979), a major shear zone that separates the Archaean craton in the north from the Proterozoic terrane to the south (Hills et al., 1968; Hills and Houston, 1979). This is also the northern edge of the Colorado Lineament, an extensive northeast-trending feature described by Warner (1978), which is approximately on strike to the northeast with the gneiss-migmatite/granite-greenstone boundary of Morey and Sims (1976). Although Warner (1978) suggested correlation of these features, there are few data that require it, and other interpretations are quite possible. Other northeast to east-northeast structural features, especially shear zones, are common in the West (Chapin et al., 1978) and in the Great Lakes area (LaBerge, 1972; Sims, 1976); Silver (1968, 1978) has reported northeasterly trends for Proterozoic chronologic provinces in Arizona and New Mexico. Thus, although many details remain to be worked out, it is clear that one principal crustal fabric through the midcontinent region is northeasterly.

The western edge of the buried Precambrian of the central United States is marked by the north—south-trending uplift of the Rocky Mountains in Colorado and smaller ranges in New Mexico (*f* in Fig. 11-1). The eastern edge of this uplift is approximately on strike with the North American Central Plains (NACP) conductive anomaly (*c*) delineated by Alabi et al. (1975). These features could be an indication of a major dislocation of basement structures in the west-central United States and attempts to correlate across them must be regarded with caution.

To the east and south, early and middle Proterozoic basement of the midcontinent region is truncated by the Grenville Province (Wynne-Edwards, 1972) and its southern extension (Lidiak et al., 1966). Rocks of similar age, perhaps an extension of this province, also occur in central Texas (Zartman, 1964; Muehlberger et al., 1966). A major feature of comparable age that cuts through the midcontinent region is the Midcontinent Geophysical Anomaly (King and Zietz, 1971) and associated Keweenawan igneous and sedimentary rocks of the Midcontinent Rift System (Green, 1977). This latter feature does not create a major problem in deciphering early Proterozoic history, although it is necessary to take into account the minor disruption created by the rifting. The orientation of the rift system may, however, be indicative of deep crustal structure in the midcontinent region.

EARLY PROTEROZOIC TERRANES (2500—1600 Ma)

The Precambrian of the midcontinent region can be roughly subdivided into those terranes that appear to be orogenic in nature (extensive deformation, metamorphism, and calc-alkaline igneous activity) and those that

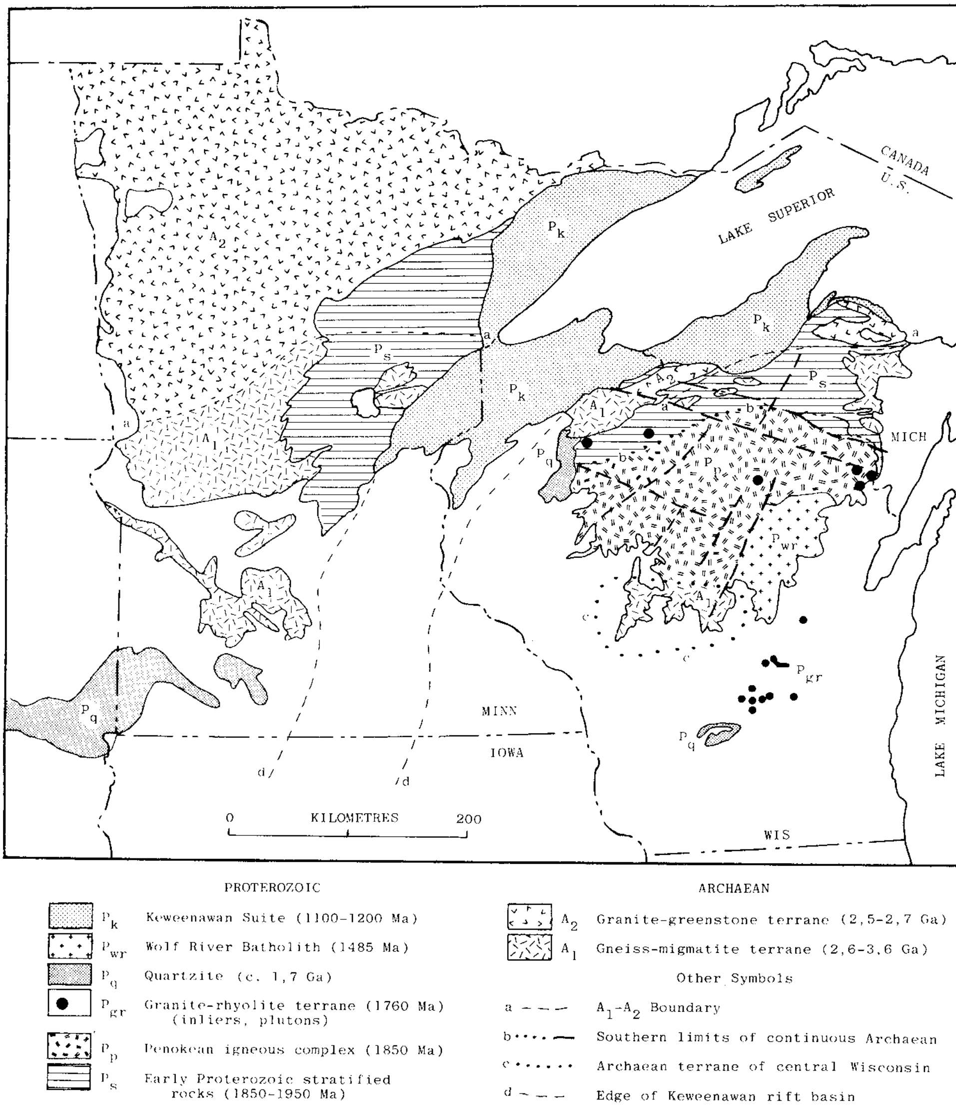

Fig. 11-2. Generalized geologic map of the western part of the Penokean Fold Belt of the Great Lakes area. Adapted from Morey (1978) and recent work by the senior author.

do not have readily apparent orogenic affinity (little pervasive deformation; limited ranges of composition — dominantly granitic — for igneous rocks). The terranes with apparent orogenic character are of early Proterozoic age and include the Penokean Fold Belt of the Great Lakes Region, the Precambrian of Colorado, Arizona and New Mexico and the intervening buried Proterozoic of the northern Plains states.

Penokean region

The Penokean Fold Belt of the Canadian Shield extends from the Sudbury, Ontario, area westward along the north shore of Lake Huron and northern Michigan, Minnesota, and northern Wisconsin (Card et al., 1972). It is in the western half of the fold belt (Fig. 11-2) that the greatest diversity of rock types is exposed and the part for which we have the most information regarding geologic history. The overall geology has been summarized in several papers (Card et al., 1972; Van Schmus, 1976; Sims, 1976), and other papers by Van Schmus and Anderson (1977), Van Schmus (1980), Smith (1978), LaBerge and Mudrey (1979), Sims (1980) Larue (1979), and Larue and Sloss (1980) have covered recent developments.

Archaean basement

Archaean rocks in the Lake Superior region range in age from about 3.5 to 2.5 Ga. The oldest rocks are found in Minnesota (Goldich and Hedge, 1974) and in the western part of Upper Michigan (Peterman et al., 1980), near the southern edge of the shield. These rocks occur south of the tectonic boundary recognized by Morey and Sims (1976; Fig. 11-2). Rocks of similar lithology occur to the east in the Southern Complex (Cannon and Simmons, 1973) and Carney Lake Gneiss (Bayley et al., 1966) of northern Michigan and in central and western Wisconsin (Van Schmus and Anderson, 1977). In these latter areas, however, no crystallization ages in excess of 2800 Ma have been reported. One distinctive aspect of much of the eastern part of this gneiss-migmatite terrane is that it has been subjected to extensive remobilization during Early Proterozoic time (Sims, 1976, 1980; Van Schmus and Anderson, 1977; Morey, 1978).

It is not known at present whether Archaean gneiss and migmatite exposed in central Wisconsin are structurally continuous with those in Michigan and northern Wisconsin. Boundary *b* in Fig. 11-2 denotes the southern limit of *confirmed* Archaean rocks (unequivocal age determinations) in the northern part of the area. Between that line and the area of confirmed Archaean rocks in central Wisconsin (boundary *c*) geochronologic studies have so far only yielded Proterozoic ages.

Pre-Penokean Proterozoic rocks

Pre-Penokean (1900 to 2500 Ma old) rocks in the Lake Superior region (Table 11-I) are found along the 1400 km long Penokean Fold Belt. This fold belt lies astride the tectonic boundary of Morey and Sims (1976); north of the boundary early Proterozoic strata are gently folded, whereas south of the boundary, overlying the Archaean gneiss-migmatite terrane, the strata are more intensely folded and metamorphosed. These sediments were

TABLE 11-I

Early Proterozoic strata in the Penokean Fold Belt

Minnesota	Michigan	Ontario (Lake Huron Area)
	Marquette Range Supergroup	
(equivalent not known)	Paint River Group: siltstone, graywacke, slate, iron-fm.	(equivalent not known)
Animikie Group (Upper): graywacke, siltstone	Baraga Group: slate, quartzite, iron-fm., metavolcanics	(equivalent not known)
	— Unconformity —	
(Middle): iron-formation (Lower): quartzite	Menominee Group: iron-fm., slate, quartzite	(equivalent not known)
— Unconformity —	— Unconformity —	
(No Group name) dolomite, quartzite	Chocolay Group: dolomite, quartzite, conglomerate	(equivalent not known)
— Unconformity —	— Unconformity —	— ? ? ? ? ? ? —
		Huronian Supergroup
(equivalent not known)	(equivalent not known)	argillite, siltstone, quartzite, conglomerate, limestone, tillite
		— Unconformity —
Archaean Basement	Archaean Basement	Archaean Basement

For further details see Sims (1976) and Robertson (1976).

not deposited along the length of the belt during the same time interval. In Ontario Early Proterozoic sedimentary rocks, designated Huronian Supergroup by Robertson (1976), are intruded by Nipissing diabase dikes which are about 2100 Ma old (Van Schmus, 1965; Fairbairn et al., 1969), thus making the Huronian Supergroup older than 2.1 Ga. In Michigan Archaean gneissic rocks underlying the Marquette Range Supergroup in the Felch Trough area (James et al., 1961) have been strongly metamorphosed as recently as 2000 Ma ago (Van Schmus et al., 1978), indicating that the less metamorphosed Proterozoic strata are younger than 2000 Ma and *not* correlative with the Huronian Supergroup. The Michigan strata are cut by 1850 Ma old plutonic rocks, thus establishing a younger age limit. Banks and Van Schmus (1972) have reported a U-Pb zircon age of 1910 ± 10 Ma for the Hemlock Volcanics of the Menominee Group within this section, consistent with a preferred age bracket of 1850—1950 Ma for the Marquette Range Supergroup.

The sedimentary rocks (Table 11-I) show two major trends. First, the older units, namely the Huronian Supergroup in Ontario and the lower part of the Marquette Range Supergroup in Michigan, consist of conglomerates, quartzites, argillites, carbonates, etc. (James, 1958; Robertson, 1976) that indicate a shallow-water environment. The younger units, particularly the Animikie Group in Minnesota and the upper part of the Marquette Range Supergroup, consist of impure quartzites, slates, greywackes, iron formations and volcanic rocks (James, 1958; Sims, 1976), indicative of a tectonically more active environment. Thus, there is a trend through time in which the depositional environments evolved from stable shelf (or miogeoclinal) in the Huronian, through slightly active shelf environments during deposition of the lower part of the Marquette Range Supergroup (Larue, 1979; Larue and Sloss, 1980) to tectonically active (eugeoclinal?) basins during deposition of the upper part of the Marquette Range Supergroup.

Second, the sedimentary formations of the region tend to thicken southward, with local variations in developing basins, suggesting that these rocks represent a sedimentary wedge formed off the evolving margin of the Archaean craton. James (1954, fig. 6) attributed the lithologic changes to basins that were forming in response to a developing volcanic highland to the south, and Van Schmus (1976, fig. 8) suggested that James' model could be related to volcanic arc development associated with a north-dipping subduction zone. A continental plate-margin model has also been proposed by Cambray (1978) in which the depositional basins for the Marquette Range Supergroup were formed during rifting and spreading along the southern margin of the Archaean craton. Larue (1979) and Larue and Sloss (1980) have studied the sedimentary environments in detail and also favour the developing continental margin model of Cambray (1978).

There were several igneous, metamorphic and tectonic events in the Lake Superior region during the early Proterozoic. The oldest recognized events include Huronian volcanism in Ontario (Robertson, 1976) prior to 2200 Ma ago and metamorphism of Archaean basement with formation of anatectic granite about 2350 Ma ago in northern Michigan (Hammond, 1978). This was followed by intrusion of post-Huronian Nipissing Diabase about 2100 Ma ago in Ontario and high-grade metamorphism of Archaean basement with local granitic intrusions about 2000 Ma ago in the Felch trough area of upper Michigan (Banks and Van Schmus, 1971, 1972; Van Schmus et al., 1978), prior to deposition of the Marquette Range Supergroup. Volcanic rocks are locally abundant in the upper part of the Marquette Range Supergroup, particularly in the Baraga Group in the southern part of upper Michigan. Chemical data (Cudzilo, 1978) suggest that these rocks are different from those in Wisconsin (see below), and geochronologic data suggest that they may be as much as 50 Ma older than the Wisconsin igneous complex.

Penokean orogenic rocks

The dominant early Proterozoic structural, igneous, and metamorphic event of the region is the Penokean Orogeny (Goldich et al., 1961; Card et al., 1972). Early geochronologic studies showed that the main peak of Penokean activity occurred about 1.8—1.9 Ga ago (cf. Goldich, 1968, 1972; Van Schmus, 1976). In the eastern Lake Superior region, north of Lake Huron, the Penokean Orogeny is represented primarily by deformation and metamorphism, with isolated plutons (Card et al., 1972; Card, 1978). In the western Lake Superior region, in Minnesota, the Penokean Orogeny is represented by deformation, metamorphism, and plutonism (Goldich et al., 1961; Goldich, 1972; Keighin et al., 1972). The Penokean orogenic terrane is most extensively developed in upper Michigan and the northern half of Wisconsin, where it includes extensive igneous complexes (P_p, Fig. 11-2), deformation, and several high-grade metamorphic nodes (Sims, 1976). Recent geochronologic and field studies (Van Schmus, 1976, 1980; Maass et al., 1980) have demonstrated that the peak of magmatism and deformation occurred about 1840 Ma ago (over the approximate range 1820—1860 Ma ago).

The volcanic rocks range from basalt to rhyolite. Their geochemistry has not been studied extensively, but available data indicate that basaltic varieties dominate (Cudzilo, 1978; Dann, 1978; LaBerge and Mudrey, 1979). Andesitic varieties, which are characteristic of modern island arcs, are apparently not a common rock type, but there are insufficient data to determine whether they are rare, as in a bimodal rhyolite-basalt suite, or whether they are just subordinate to basaltic varieties in a more or less calc-alkaline suite. Associated plutons range from granite (adamellite) to tonalite and quartz diorite (Cudzilo, 1978; Maass et al., 1980). Geochemical data for these rocks are also quite limited but they appear to represent calc-alkaline plutonism. Initial $^{87}Sr/^{86}Sr$ ratios for these rocks are typically low, c. 0.702 (Van Schmus et al., 1975c; Sims and Peterman, 1980). Thus, the igneous complex of northern Wisconsin could represent the core of an arc that developed along a convergent plate boundary and was then accreted to the southern margin of the Archaean craton.

Post-Penokean rocks

There are four principal post-Penokean lithologic suites in the Proterozoic of the Great Lakes area:

(a) A 1760 Ma old granite-rhyolite terrane in southern Wisconsin (Smith, 1978; Van Schmus, 1978, 1980) which probably extends into the subsurface of Iowa and Illinois (Goldich et al., 1966b).

(b) A slightly younger group of orthoquartzites that occurs in Wisconsin, Iowa, Minnesota, Nebraska, and South Dakota (Dott and Dalziel, 1972; Austin, 1972).

(c) Undeformed plutons c. 1480 Ma old (Van Schmus et al., 1975a, b; Anderson and Cullers, 1978).

(d) The 1100 Ma old Keweenawan igneous and sedimentary suite (e.g. Green, 1977; Craddock, 1972).

In addition, there is one further event in the region for which no newly generated rocks are known: Rb-Sr whole-rock systems throughout the western Great Lakes region have recorded a resetting (presumably metamorphic) event at c. 1630 Ma ago.

Discussion of all of these is beyond the scope of this paper but the 1760 Ma old felsic igneous suite and the 1630 Ma old metamorphic event relate directly to the problems under consideration. The 1480 Ma old plutonism is also germane and is included in subsequent sections.

1760 Ma old rhyolite-granite suite

In southern Wisconsin are several inliers of Precambrian rhyolite and granite (Fig. 11-2). The volcanic rocks consist almost exclusively of rhyolite which in many instances is ignimbritic. The granite is primarily leucocratic, epizonal granophyre, and recent chemical data (Smith, 1978) indicate that the rhyolite and granite are genetically related, if not co-magmatic. Zircons from several localities define an age of 1760 ± 10 Ma (Van Schmus, 1978). Regional geochronologic studies in northern Wisconsin have revealed at least four granitic plutons that are coeval with the southern Wisconsin samples (Van Schmus, 1980, and unpubl. data; Fig. 11-2). On the basis of limited chemical data these plutons from northern Wisconsin appear to be similar to those in southern Wisconsin (J. L. Anderson, pers. commun., 1979), but they are generally coarser-grained, granular and overall more mesozonal in character. The northern 1760 Ma old plutons thus appear to be deep-seated equivalents of the southern units and indicate that the 1760 Ma event was widespread.

This event followed the peak of Penokean igneous activity by about 80 Ma, and so far no rocks with ages in the interval 1760—1820 Ma have been found. The 1760 Ma old rocks do not appear to represent a major orogenic assemblage and do not appear to be associated with a major tectonic event. It is not clear whether they should be considered a late phase of Penokean activity, but in view of the distinct time gap and the distinctly different chemical character of the rocks (primarily granite; versus intermediate granite through tonalite for Penokean plutons, Cudzilo, 1978) we tentatively consider the 1760 Ma event post-Penokean.

1630 Ma old event

The rhyolitic volcanism was followed by extensive deposition of blanket sandstones, now quartzite (Dott and Dalziel, 1972; Austin, 1972), and the entire southern half of the Lake Superior region was subjected to some event that effectively reset a wide variety of mineral and whole-rock Rb-Sr systems 1630 Ma ago. This resetting has been found in several studies (e.g. Peterman, 1966; Van Schmus et al., 1975c; Sims and Peterman, 1980) and is remarkable in that there is surprisingly little scatter in these reset ages even though an area in excess of 150,000 km^2 was affected. At present no igneous rocks of any kind have been found in the region having an age even close to 1630 Ma; there appears to be a complete absence of igneous rocks in the interval 1530—1760 Ma.

The rhyolite and quartzite have been gently to tightly folded (Dott and Dalziel, 1972; Smith, 1978), and it is possible that this deformation occurred during the event that reset the Rb-Sr systems (or vice-versa). We believe, for reasons outlined below, that the 1630 Ma event in the Lake Superior region is a foreland manifestation of a more extensive orogenic event that occurred to the south.

Western United States

If there are any major orogenic belts that traverse the midcontinent region parallel to the general NE—SW structural trend we would also expect them to be present in the Precambrian basement of the western and southwestern United States. There is a scarcity of published U-Pb data for zircons from Proterozoic igneous rocks of the western U.S., making regional synthesis difficult. Several zircon ages have been presented in summary form in a variety of papers and abstracts, however, and we have attempted to compile a reasonably complete set of these ages (Table 11-II). One of the most obvious conclusions that can be drawn is that there are no ages older than 1800 Ma, and hence no known igneous rocks in the western U.S. that can be considered coeval with igneous rocks of the Penokean Orogeny in the Great Lakes region. On the other hand, Silver (1968) and Silver et al. (1977a) have proposed that two parallel, NE-trending orogenic terranes exist in the

TABLE 11—II

U-Pb ages on early Proterozoic zircons from the western U.S.

Unit and Location[a]	Age (Ma)[b]	Ref.[c]
Deception Rhyolite, Arizona (v)	1780 ± 10	1
Twilight Gneiss, Colorado (v)	1745 ± 20	2
Big Bug Group, Arizona (v)	1740 ± 10	1
Brady Butte Granodiorite, Arizona (p)	1735 ± 10	1
Government Canyon Granodiorite, Arizona (p)	1735 ± 10	1
Quartz diorite of Mingus Mountain, Arizona (p)	1725 ± 10	1
Payson Granite, Arizona (p)	1710 ± 20	3
Alder Series, Arizona (v)	1700 ± 15	4
Zoraster granite gneiss, Arizona (p)	1695 ± 15	5
Boulder Creek Granite, Colorado (p)	1695 ± 25	6
Baker's Bridge/Ten Mile granites, Colorado (p)	1690 ± 20	2
Pinal Schist, Arizona (v)	1690 ± 10	7
Red Rock Rhyolite, Arizona (v)	1685 ± 15	8
Tres Piedras Granite, New Mexico	1625 ± 17	9
Johnny Lyon Granodiorite, Arizona (p)	1625 ± 10	7
Granite near Sunflower, Arizona (p)	1625 ± 15	8

[a] (v) denotes volcanic; (p) denotes plutonic.
[b] Ages are based on decay constants recommended by Steiger and Jäger (1977) and have been recalculated from original references where necessary by decreasing the reported ages by approximately 2%.
[c] References: 1, Anderson et al., 1971; 2, Silver and Barker, 1967; 3, Ludwig and Silver, 1977; 4, Ludwig, 1974; 5, Pasteels and Silver, 1965; 6, Stern et al., 1971; 7, Silver and Deutsch, 1963; Silver, 1963, 1978; 8, Silver, 1964; 9, Maxon, 1976.

southwest, the northern one consisting of rocks 1690—1780 Ma old and the southern one consisting of rocks 1610—1680 Ma old. These two terranes strike into the midcontinent region and thus need to be considered as potential exposed ends of terranes in the buried basement.

1690—1780 Ma old belt

Exposed rocks of this belt occur in Colorado (Peterman and Hedge, 1968; King, 1976), south of the Nash Fork—Mullen Creek shear zone in Wyoming (Hills and Houston, 1979), in central and northern Arizona (Anderson et al., 1971; Brown et al., 1979; Babcock et al., 1979; Clark, 1979), and in northern New Mexico (Barker et al., 1976; Condie, 1979). Metasedimentary rocks in this belt include phyllite, schist, gneiss, quartzite, migmatite and marble. Recent work in Arizona (Brown et al., 1979) and in New Mexico (Condie and Budding, 1979) indicates that quartzite and pelitic schists are the major metasedimentary rocks, but Hills and Houston (1979) indicate that meta-greywackes may be more common in the northern part of the belt, though still not dominant. Metavolcanic rocks range from original tholeiitic through rhyolitic compositions and are typically interlayered with the metasedimentary rocks as greenstone, amphibole, mafic or felsic schists and granitic gneiss. Recent studies (Condie and Budding, 1979; Clark, 1979) indicate that basaltic varieties dominate and felsic varieties may be locally significant; andesitic varieties are apparently rare to absent. Plutonic rocks range from tonalite to granite but the more felsic varieties are more abundant.

Rb-Sr dating of high-grade rocks from this metamorphic complex in the Front Range of Colorado yielded an isochron age of 1715 ± 30 Ma (Idaho Springs Formation, Hedge et al., 1967) which is interpreted as the time of metamorphism. The upper age limit for these rocks is not tightly constrained, although the $^{87}Sr/^{86}Sr$ intercept of 0.708 for the isochron indicates that these rocks had a relatively short crustal history, since their $^{87}Rb/^{86}Sr$ ratios range from 2 to more than 20. There are no zircon ages from the volcanic rocks in excess of 1780 Ma. It is quite possible that the entire supracrustal complex is less than 1800 Ma old, although recent Rb-Sr dating by Divis (1977) along the northern edge of the belt indicates some of the units may be older than 1800 Ma.

Plutons that intrude the supracrustal rocks have ages from about 1690 to 1740 Ma (Table 11-II) which, taken with the Rb-Sr results above, indicate that the intrusive phase of orogeny in this belt occurred 1715 ± 25 Ma ago. Regional synthesis in this region has not proceeded to the point where this orogeny has been named, although some authors have suggested "Hudsonian" (King, 1976) or "Penokean" correlations (Hills and Armstrong, 1974). We do not believe either is appropriate and will leave this event unlabeled.

In many recent papers this orogenic belt is regarded as a result of continental margin activity with implied or suggested subduction zone environments (Hills et al., 1968; Hills and Armstrong, 1974; Barker et al., 1976; Silver et al., 1977a; Hills and Houston, 1979; Houston et al., 1979; Brown et al., 1979; Babcock et al., 1979; Clark, 1979). Such a conclusion is not required by available data as pointed out by Barker et al. (1976) who have also suggested vertical tectonics over a sialic crust, and by Condie (1979) and Condie and Budding (1979) who have suggested continental rift tectonics as alternative models. A major inference in the two latter hypotheses is that Archaean continental crust may underlie the early Proterozoic orogenic rocks. So far, however, there are no isotopic or geochemical data that suggest such older rocks at depth.

1610—1680 Ma old belt

Rocks belonging to this younger belt occur throughout southern Arizona and central and southern New Mexico (Silver et al., 1977a). There is much less information generally available for this belt than for the northern belt, and the principal summaries are those of Silver (1978) for southern Arizona and Condie and Budding (1979) and Robertson and Moench (1979) for New Mexico. In many respects this younger belt is quite similar to the older one. The terrane consists of metavolcanic and metasedimentary rocks that have been deformed, metamorphosed and intruded by a series of calc-alkaline plutons. The principal differences between the two belts are geographic location and age.

Silver (1978) has concluded from U-Pb dating of zircons that the supracrustal rocks of this terrane are about 1680—1700 Ma old, whereas the post-tectonic plutons are 1625 ± 10 Ma old (Table 11-II). Thus, the orogeny is bracketed between 1625 and 1680 Ma. As Silver (1964) has pointed out this event is probably the Mazatzal Orogeny identified by Wilson (1939) and can be traced northeastward into New Mexico (Silver et al., 1977a; Silver, 1978). We will refer to this terrane and its possible eastward extension as the Mazatzal Belt although we recognize the potential problems in doing so. Silver (1978) has likened the rock assemblages in this belt to those formed in typical geosynclines and favours a continental margin—subduction zone environment. Condie and Budding (1979), on the other hand, have proposed a tensional intra-cratonic setting for this belt, largely because of the similarity in rock types with those of modern continental rift environments.

Midcontinent region

The principal sources of information on the geology of early Proterozoic basement rocks in the midcontinent are the pioneering studies of Goldich et al. (1966a, b), Lidiak et al. (1966) and Muehlberger et al. (1966, 1967) and more recent work by Lidiak (1971, 1972), Bickford et al. (1975, 1979), Harrower (1976) and Kisvarsanyi (1972, 1974, 1979). A compilation of data and a summary of the Precambrian geology and geochronology of the midcontinent area from the Great Lakes south through Texas and from the Appalachian Mountains west to the Rocky Mountains has recently been completed by Denison et al. (in press) as part of a project of the North American Working Group for the Precambrian of the International Union of Geological Sciences. Figure 11-3, adapted from that work, shows the distribution of rock types in the southern midcontinent region.

The basement rocks of North and South Dakota are mainly known from interpretation of geophysical data and from several hundred basement well samples (Muehlberger et al., 1967; Lidiak, 1971). These data indicate that the eastern portions of both North and South Dakota are underlain by extensions of the Archaean terranes of the Canadian Shield. Western North and South Dakota are underlain by extensive terranes of metamorphic rocks including mafic and silicic schist and gneiss, and granite. The only age measurements available for these rocks are Rb-Sr analyses for single samples of whole rocks or minerals from well cuttings. These indicate that the rocks were formed in the period 1700—1900 Ma ago (Goldich et al., 1966a, b; Lidiak, 1971; Denison et al., in press). The presence of Archaean rocks overlain by a thick sequence of metasedimentary rocks in the Black Hills of South Dakota (Zartman and Stern, 1967) suggests that much of the western Dakotas may be underlain by metasedimentary sequences originally deposited

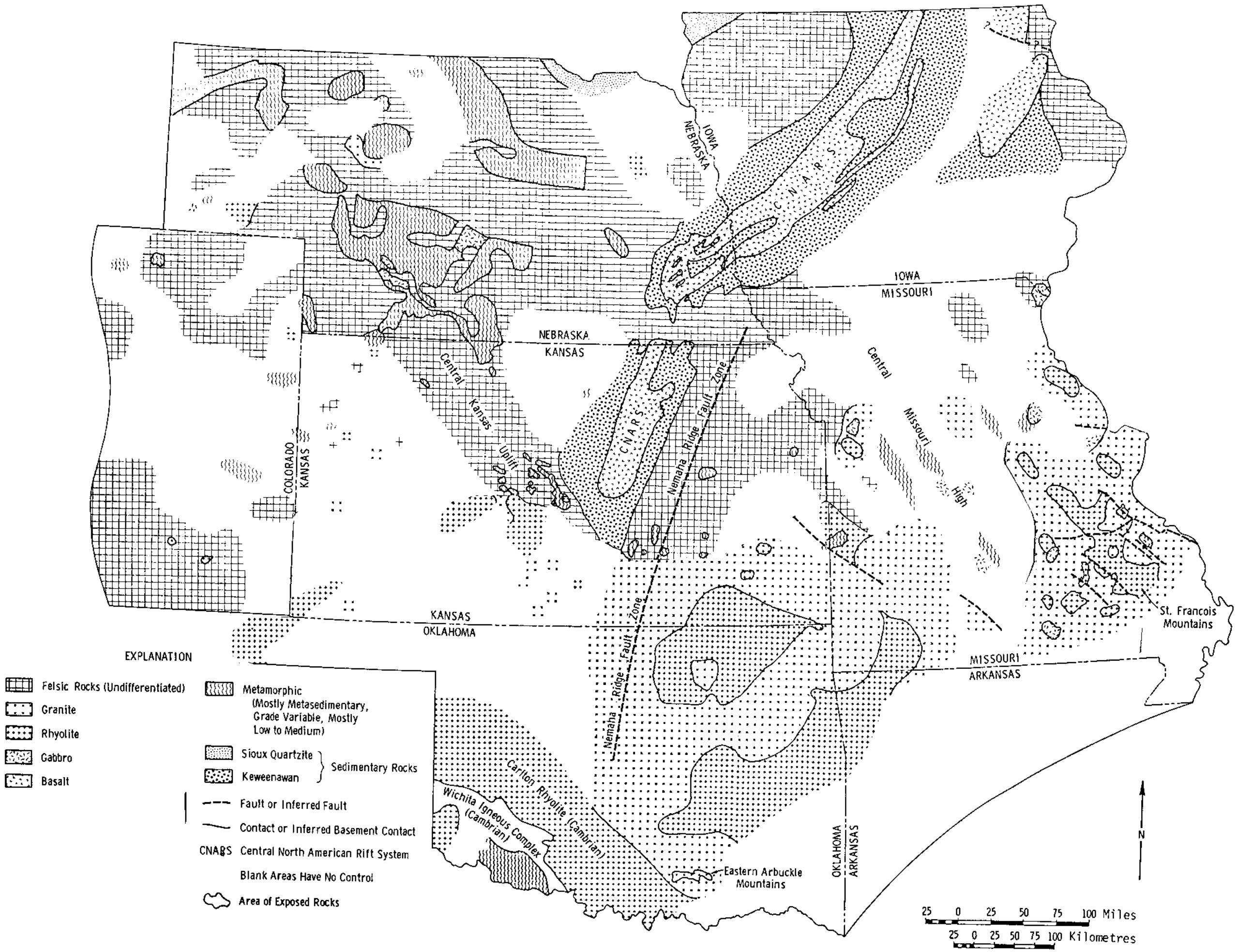

Fig. 11-3. Generalized geologic map of basement rocks in the southern midcontinent region of the United States. Adapted from report by Denison et al. (in press).

upon older Archaean crust with intrusion of granitic rocks and metamorphism occurring in the early Proterozoic.

The basement rocks of Nebraska, Kansas and Missouri are known from a large number of wells that have penetrated the crystalline crust, but very little is known of the basement rocks of Iowa. Basement rocks in Nebraska have been described by Lidiak (1972) and include large terranes of gneissic granitic rocks, metasedimentary rocks including abundant quartzite, muscovitic schist and biotitic schist, silicic metavolcanic rocks and minor amphibolite. Many of the granitic gneisses appear to be of igneous origin with foliation induced by pervasive shearing. Small bodies of gabbroic and anorthositic rocks are also known. The basement rocks of northern Kansas and northern Missouri are quite similar to those of Nebraska according to descriptions by Bickford et al. (1979) and Kisvarsanyi (1972, 1974, 1979) except that metamorphic rocks are apparently less abundant. In all of these areas sheared granitic to granodioritic rocks make up 60% or more of the basement.

Conspicuously absent in the early Proterozoic basement of the midcontinent are extensive terranes of mafic rocks, intermediate volcanic or plutonic rocks or of their metamorphosed equivalents. The metavolcanic rocks are predominantly rhyolitic to dacitic in composition and the metasedimentary rocks are mostly quartzite and metapelites.

For the most part geochronologic information for the midcontinent region is based on single Rb-Sr total rock or mineral ages and K-Ar ages from well cuttings and a few core samples from the basement. These ages are susceptible to many uncertainties including weathering and metamorphism of the rocks and uncertainties in initial Sr compositions, and in general such "ages" should only be taken as minimum ages. Work at the University of Kansas has included a major effort to obtain zircons from rocks of the subsurface and we have been successful in several cases. So far all but two of the zircon ages obtained have been from rocks of the 1380—1480 Ma event (below), but zircon samples from a granitic pluton in northeastern Kansas and one in northwestern Missouri have yielded ages of 1625 ± 25 Ma (Bickford et al., 1979, and unpubl. data). These are the only two localities in the buried early Proterozoic portion of the midcontinent for which we have high-quality age information. The Harney Granite in the Black Hills of South Dakota (Riley, 1970) yielded an excellent Rb-Sr isochron age of 1710 ± 15 Ma. However, the initial $^{87}Sr/^{86}Sr$ for that isochron is 0.7143 ± 0.0006, indicating either that the granite was derived from partial melting of older crustal rocks or that the Rb-Sr system has been severely disturbed. The first alternative is consistent with the presence of Archaean rocks in the Black Hills (Zartman and Stern, 1967) and the location of the Black Hills north of the probable southern boundary of the Archaean craton (Fig. 11-1).

MIDDLE PROTEROZOIC ROCKS OF THE MIDCONTINENT

The central and southern parts of the midcontinent region are underlain by a unique terrane of undeformed and unmetamorphosed silicic volcanic rocks and related epizonal to mesozonal granitic plutons that were formed 1370—1485 Ma ago (Fig. 11-1). These rocks were first described by Muehlberger et al. (1966) and Lidiak et al. (1966) in their pioneering petrographic and geochronological studies of the buried basement of North America. More recent work includes that of Bickford and Mose (1975) and Kisvarsanyi (1979) in southeastern Missouri, of Denison (1966), Denison et al. (1969) and Bickford and Lewis (1979) in Oklahoma, and of Bickford et al. (1979) in Kansas.

This terrane is unusual in that it includes almost no igneous rocks of mafic or intermediate composition, no sedimentary rocks, and no metamorphic rocks; no rocks with ages greater than 1485 Ma are known from this terrane. Rocks of this general composition and age are also found as intrusives within the older terranes to the north. Both the plutonic suite and the granite-rhyolite terrane are described below.

Undeformed plutonic rocks

A number of plutons, mostly granitic in composition, occur within the 1600—1900 Ma old orogenic terranes described in the preceding section of this paper (they are rare to absent in the Archaean terranes). Table 11-III gives the names, locations and reported ages of a number of these plutons. All of these rocks have yielded ages in the range 1380 to 1480 Ma. The undeformed nature of these rocks indicates that they were not emplaced in an orogenic stress regime. In the midcontinent region poor exposures preclude study of contact relations (indeed, most of the plutons are known from the subsurface), but where representatives of this suite are exposed in the Rocky Mountains (e.g. St. Kevin Granite, Colorado, Doe and Pearson, 1969; Vernal Mesa-type Quartz Monzonite, Colorado, Bickford and Cudzilo, 1975) they can be seen to cut older foliated rocks. Silver et al. (1977b) have pointed out that these plutons are not associated with sedimentary sequences of similar age and were thus probably emplaced within the craton rather than marginal to it.

As discussed more fully below, these plutonic rocks have essentially the same age as an extensive terrane of rhyolitic volcanic rocks and epizonal granitic plutons that lie to the south. However, whereas the plutons of the granite-rhyolite terrane are typically granophyric and have other textural features indicating emplacement at shallow crustal levels, most of the undeformed plutons have petrographic features (abundant perthite, mafic minerals such as biotite and hornblende, lack of granophyric texture) suggesting emplacement at moderate (8—12 km) depths.

An interesting feature of these plutons, one that potentially has genetic and tectonic significance, is that they tend to become younger to the west and southwest (Table 11-III). The age progression of these plutons apparently spans about 60 Ma but their genetic association seems inescapable. Possible models for the formation of these bodies are discussed below.

Granite-rhyolite terrane of the south-central United States

A major feature of the midcontinental region of the United States is a great belt of rhyolitic volcanic rocks and associated epizonal granitic plutons. This terrane, which can be traced in the subsurface from northwestern Ohio across Indiana, Illinois, Missouri, Arkansas, Kansas and Oklahoma into the Texas Panhandle region, is about 2000 km long and 500 km wide. If these rocks make up only the upper 1 km of the crust in this region they still represent 10^6 km^3 of very silicic (SiO_2 commonly > 70 weight%) and potassic (K_2O commonly 5—6%) material. Mafic and intermediate rocks are rare as are associated sedimentary and metamorphic rocks. Most of the volcanic rocks have features (e.g. shard structures) indicating that they are ash-flow tuff. None of these rocks show penetrative deformation; metamorphism is limited to the presence of minor epidote and chlorite, granular recrystallization of the ground mass in some rocks and the effects of devitrification.

The major exposure of this terrane is in the St. Francois Mountains of southeastern Missouri where some 900 km^2 are underlain by epizonal granite and an extensive terrane of rhyolitic ash-flow tuff and minor rhyolitic and dacitic flows. Field studies in this area (Sides and Bickford, 1978) have shown that the ash flows were probably erupted from one or more calderas and that some of the major plutons are exposed because of deep erosion of one of these calderas. SiO_2 in these rocks ranges between 68 and 76 weight% but the more silicic varieties predominate. K_2O/Na_2O ranges from about 0.75 in the lower silica rocks to as great as 1.4 in the high-silica varieties.

Bickford and Mose (1975) showed that a number of granitic plutons and at least one of the major rhyolite flows formed 1470 ± 20 Ma ago (Table 11-III). To the southwest, however, epizonal granite bodies in southern Kansas (Bickford et al., 1979) have yielded zircon ages of 1380 ± 20 Ma as have similar rocks exposed at the surface in northeastern Oklahoma and mesozonal granites exposed in the eastern Arbuckle Mountains of southern Oklahoma (Denison, 1973; Bickford and Lewis, 1979; Table 11-III). Still farther to the west, in the Texas and Oklahoma Panhandle area, rhyolite and granite known from drill cuttings and rare cores have yielded Rb-Sr ages in the range 1100 to 1350 Ma (Muehlberger et al., 1966); these ages may, however, be minimum values only. The ages of rhyolite and granite samples from the basement of Illinois, Indiana, and Ohio are poorly known, for only Rb-Sr measurements have been made on total rock or mineral samples from scattered cuttings. The data available indicate that many of these rocks may have formed between 1200 and 1500 Ma ago whereas some, particularly to the north, may be part of the 1760 Ma old terrane exposed in southern Wisconsin.

TABLE 11-III

U-Pb zircon ages of undeformed plutons and plutons of the granite rhyolite terrane

Unit and location	Age (Ma)[a]	Ref.[b]
Undeformed plutons		
Granite, Manitoulin Island, Ontario	1467 ± 10	1
Wolf River Batholith, Wisconsin	1485 ± 15	2
St. Francois Mountains, Missouri (Includes granitic plutons and rhyolitic volcanic rocks; may better be considered part of granite rhyolite terrane)	1470 ± 20	3
Diorite, LaClede County, Missouri	1458 ± 15	4
Gneissic Granite, LaClede County, Missouri	1456 ± 15	4
Granite, Russell County, Kansas	1445 ± 15	5
Granite, Red Willow Batholith, Furnas County, Nebraska	1445 ± 15	5
St. Kevin Granite, Colorado	1394 ± 17	6
Vernal Mesa Type Quartz Monzonite, Colorado	1440 ± 16	7
Eolus Granite, Colorado	1440 ± 20	8
Sandia Granite, New Mexico	1437 ± 17	9
Tungsten King Granite, Arizona	1420 ± 10	10
Plutons of the Granite Rhyolite Terrane		
Granite, Greenwood County, Kansas	1380 ± 33	4
Granite, Rose Dome, Woodson County, Kansas	1408 ± 21	4
Granite, Stevens County, Kansas	1372 ± 30	4
Spavinaw Granite, Oklahoma	1370 ± 20	11
Tishomingo Granite, Oklahoma	1374 ± 15	11
Troy Granite, Oklahoma	1399 ± 95	11
Blue River "Gneiss", Oklahoma	1396 ± 40	11

[a] Ages based upon the decay constants recommended by Steiger and Jäger (1977).
[b] References cited: 1, Van Schmus et al., 1975b; 2, Van Schmus et al., 1975c; 3, Bickford and Mose, 1975; 4, Bickford et al., 1979; 5, Harrower, 1976; 6, Doe and Pearson, 1969; 7, Bickford and Cudzilo, 1975; 8, Silver and Barker, 1967; 9, Steiger and Wasserburg, 1966; 10, Silver, 1978; 11, Bickford and Lewis, 1979.

An important aspect of the regional distribution of ages is that the 1380—1480 Ma old granite-rhyolite terrane is bounded in the north by rocks that are 1600 Ma old or older, in the east by rocks of the Grenville Province whose ages are about 1100 Ma, and in the south by the rocks of the Llano Province of Texas whose ages are also about 1100 Ma. The granite-rhyolite terrane is surrounded by either older and younger rocks (Fig. 11-1) but no older or younger crystalline rocks are known within it.

DISCUSSION

Regional correlations of early Proterozoic terranes

In order to examine potential correlations into and across the midcontinent region we have summarized known age relationships in Fig. 11-4 which forms the basis of our general interpretation of the buried basement in the midcontinent (Fig. 11-5). An essential feature of both Figs. 11-4 and 11-5 is the presence of several large question marks, clearly indicating where fact leaves off and conjecture begins.

For the southern part of the continent there appear to be at least three distinct early Proterozoic orogenic events: Penokean in the Great Lakes region (1820—1900 Ma); un-named in Colorado—Arizona (1690—1780 Ma); and Mazatzal in Arizona (1610—1680 Ma). Of these only the latter appears traceable across the midcontinent, with igneous rocks located as far east as northwestern Missouri (Fig. 11-5). We attribute the 1630 Ma metamorphic overprinting in the Great Lakes region to thermal activity (and possible deformation) associated with eastward extension of the Mazatzal Belt at

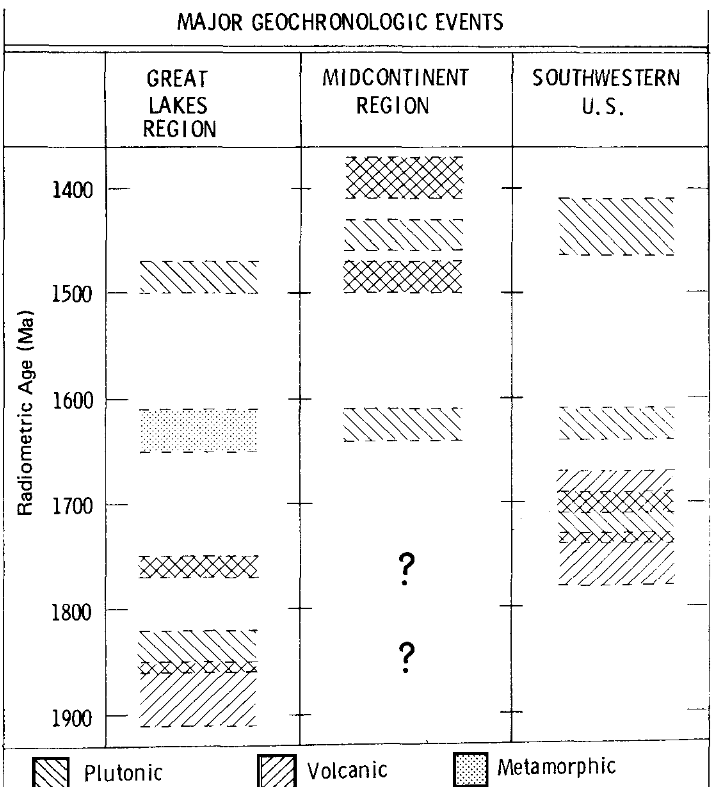

Fig. 11-4. Chart showing absolute ages of igneous and metamorphic events in the midcontinent region and surrounding terranes. Question marks indicate presence of rocks whose ages are not well known. Age assignments of events based on U-Pb zircon dating except for metamorphic event.

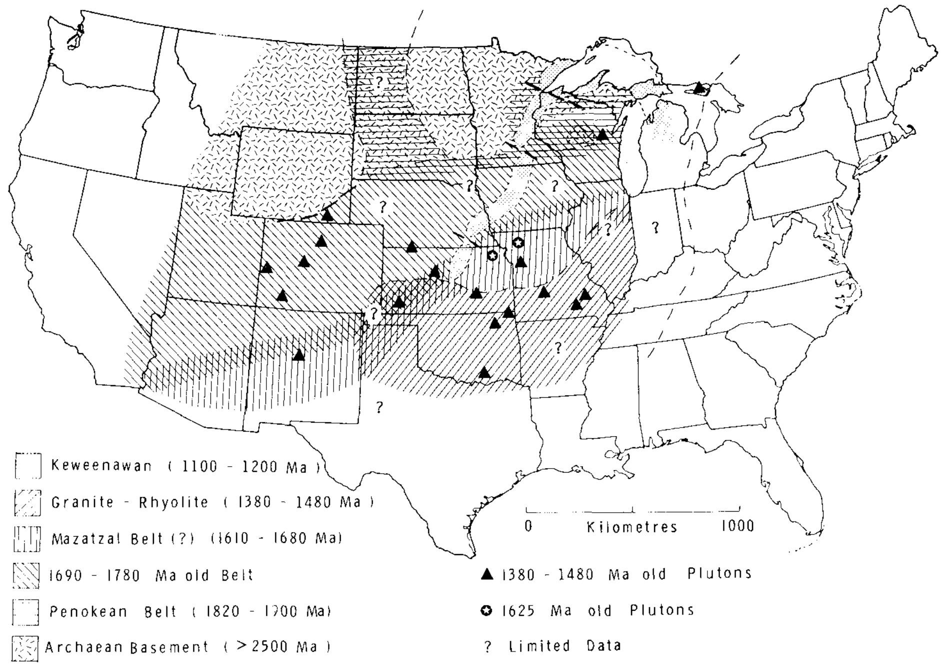

Fig. 11-5. Map of the United States showing the inferred basement chronologic terranes and distribution of plutons of the 1380—1480 Ma event. The two identified 1625 Ma old plutons represent the only igneous rocks in the midcontinent older than 1500 Ma for which precise and accurate ages are available. Question marks indicate regions of poor control.

least as far as Illinois. Thus, we anticipate that future results from the subsurface of Illinois (and possibly southeastern Iowa) will include identification of additional plutons about 1625 Ma old.

One important line of evidence that could be used to support a continental margin setting for the Penokean Orogeny would be extensive lateral development of the fold belt. The belt is terminated in the east by the Grenville Province (Fig. 11-1), and at present there are no published U-Pb (zircon) ages in the interval 1820—1900 Ma for orogenic igneous rocks anywhere west of Minnesota. The only possible correlative age reported is a Rb-Sr isochron age of 1850 ± 100 Ma for the Big Creek Gneiss in the Sierra Madre range in south-central Wyoming (Divis, 1977), but the result is based on limited sampling and subject to the uncertainties inherent in Rb-Sr dating. Consequently, *we consider it inappropriate at this time to correlate the Penokean Orogeny of the Great Lakes region with orogenic terranes in Colorado—New Mexico—Arizona.* Accordingly, we have shown the Penokean terrane dying out westward (Fig. 11-5).

Eastward extension of the Colorado orogenic terrane is also uncertain.

It appears that the Harney Granite in the Black Hills is coeval with the plutonic suite in Colorado, indicating extension of that terrane at least that far east. No plutons or metasedimentary rocks that are 1690—1780 Ma old and have orogenic characteristics have been found in Wisconsin yet, although the 1760 Ma old rhyolite and granite suite is approximately coeval with early volcanic rocks of the Colorado—Arizona terrane (1740—1780 Ma; Table 11-II). Whether there is any common tectonic link between these terranes remains a major unsolved problem and detailed zircon age studies on subsurface rocks from Iowa, Nebraska, southern Wisconsin and northern Illinois will be essential. However, minimum ages in the northern part of the buried basement fall in the range 1600—1800 Ma, with many over 1700 Ma, and it is reasonable to assume that the basement terrane of Nebraska and Iowa includes units correlative with either the Penokean terrane to the east or the orogenic terrane of Colorado and northern Arizona (or both). We see no reason at this time to postulate any additional events.

We would like to point out one additional and intriguing possibility regarding correlation of the early Proterozoic rocks of Colorado with the Penokean Fold Belt. In many respects the orogenic terrane in Colorado is similar to that in Wisconsin and vicinity even though it is about 75 Ma younger. If these terranes were formed by subduction-zone and island-arc activity along an active continental margin it is possible that they represent a time-transgressive orogenic belt formed during oblique collision. Island-arc—continent collision could have occurred first in Wisconsin and migrated westward. Such oblique geometry would also be consistent with the extensive development of northeast-trending shear zones which have been considered indicative of lateral displacement (Warner, 1978; Houston et al., 1979). This would require only two major orogenic events (Penokean, Mazatzal) rather than three as implied by the data taken at face value.

In gross aspect our presentation in Fig. 11-5 does not differ appreciably from earlier compilations: the midcontinent region consists of NE trending igneous and metamorphic terranes (belts) that young to the southeast. We believe the southward younging is real, but it may be much more complex, with more overlap, than indicated in Fig. 11-5.

Origin of the early Proterozoic orogenic belts

There are three models that could apply to the early Proterozoic tectonic belts of the midcontinent: ocean—continent convergent plate margin (including arc—continent convergence), continent—continent convergent plate boundary, and intracratonic rifting. Of these we see the least evidence of continent—continent convergence. As mentioned earlier an ocean—continent convergent plate boundary has been proposed for development of the Penokean Orogeny. There are, however, several aspects of the regional geology that do not fit readily into such a model. Among these are:

(A) The prevailing structural trend in the eastern part of the fold belt is E—W, and it is not possible to continue the fold belt westward on the same trend. Western Minnesota is apparently underlain by an unbroken Archaean basement, and in southwestern Minnesota, specifically the Minnesota River Valley (Goldich et al., 1970; Grant, 1972), Penokean igneous and metamorphic activity was minor to absent (Goldich, 1972). Although there was Penokean deformation and plutonism in east-central Minnesota (Keighin et al., 1972) the main part of the orogenic belt must either die out or swing sharply southwestward. The latter possibility would be consistent with the predominant northeast—southwest structural trends in Wisconsin (LaBerge, 1972; Sims, 1976; Maass et al., 1980). Thus, if the Penokean Fold Belt was developed along a continental margin, that margin must have been somewhat irregular.

(B) Subsequent to the suggestion of a convergent plate-margin setting for the Penokean Orogeny by Van Schmus (1976) it was discovered that many of the gneissic rocks of central Wisconsin are Archaean in age (Van Schmus and Anderson, 1977). These Archaean rocks were intruded by a number of plutons of Penokean age (1820—1850 Ma, Van Schmus, 1980), indicating that this terrane was extensively involved in that orogeny. Thus, the simple idea that the volcanic-plutonic belt of northern Wisconsin was originally an island arc with oceanic crust lying to the south became less tenable. At least four alternate models can be suggested. First, the basement under northern Wisconsin may consist of contiguous Archaean continental crust (Sims, 1976; Morey, 1978), indicating that the volcanic-plutonic terrane was a continental arc rather than an island arc (if it is an arc at all). Second, the Archaean block in central Wisconsin may be a detached piece of the Superior craton left behind during an early Proterozoic rifting event that created a continental margin. Third, the Archaean block may have been a microcontinent that was accreted onto the Superior craton during the Penokean Orogeny and formed part of the crust upon which an arc was developed. Fourth, much of the early Proterozoic terrane may be allochthonous, analogous to the southern Appalachians (cf. Cook et al., 1979; Harris and Bayer, 1979), the Archaean rocks being exposed in a window in the upper plate.

(C) Much of the tectonic activity in northern Michigan consisted of vertical movements (block faulting and uplifts, etc.; Cannon, 1973; Klasner, 1978). This, plus the possible continuity of Archaean basement through Wisconsin, led Sims (1976) and Morey (1978) to argue in favour of intracratonic tectonism. There is no evidence in Michigan or northern Minnesota of a well-developed foreland thrust belt such as is well developed in the Coronation Geosyncline region (Hoffman, 1973). This could be an artifact of depth of erosion but it could also indicate that compressive forces were relatively unimportant and that vertical movements were dominant. Unfortunately, exposures are very scattered in Wisconsin, and the volcanic stratigraphy of that area is poorly known, so that we do not know whether a more compressive structural regime might have prevailed there as required by some of the alternatives mentioned in the preceding paragraph.

(D) In so far as is currently known, the volcanic and metasedimentary rock suites do not bear a close resemblance to modern convergent plate-margin assemblages. There are no recognized ophiolites or paired metamorphic belts and the lithologies seem to be different. For example, there is no abundant andesite and sedimentary rocks with island-arc eugeosynclinal character are lacking (many of the "eugeosynclinal" rocks in the early Proterozoic section do not appear to be derived from adjacent volcanic sources). How can we account for abundant rhyolite in some areas (especially southern Wisconsin), particularly in light of the shortage of andesite? Why is no oceanic crust preserved?

As an alternative to a convergent continental margin tectonic setting Barker et al. (1976) and Condie and Budding (1979) have suggested that many of the rocks of the early Proterozoic terranes in the Colorado—New Mexico region have characteristics suggesting formation in a multiple rifting or ridge-and-basin environment. Implied in this is the presumed existence of older continental crust under these terranes.

Although the rifting model has merit with respect to rock types, both sedimentary and igneous, we are concerned about the apparent absence of Archaean or older Proterozoic basement under the tectonic belts, particularly for the 1690—1780 Ma old and 1610—1680 Ma old terranes in the Colorado—Arizona—New Mexico region. In addition, rifting implies older continental material on both sides of a rift, and as far as we can tell the general trend across the belts reflects a monotonic decrease in ages southward. If the rifting environment hypothesis is to be viable it must explain why there were recurring rifting events in the same general belt and where the complementary part of the crust (the rifted away portion) of the original continent went.

An alternative to trying to explain the features of the early Proterozoic orogenic belts with modern styles of subduction or rifting is either to invoke a more complex model that has features of both or to propose different types of plate tectonics in the early Proterozoic. Examples of models that involve both rifting and convergence or subduction are those of Lipman et al. (1972) and Christiansen and Lipman (1972) for the Cenozoic evolution of the western United States, of Kröner (1977, 1979) for intracratonic mobile belts in Africa and of Baragar and Scoates (this volume, Chapter 12) for the mobile belts of the Churchill Province that surround the Superior Province of the Canadian Shield. These are all basically intracratonic models and a major problem in applying any type of intracratonic model to the midcontinent region is: (a) the virtually complete absence of older basement in the early Proterozoic belts of the midcontinent (the block of Archaean basement in central Wisconsin can easily be explained in a continental margin model); and (b) the progressive decrease in ages outward from the

Archaean core in the north with no hint of an older piece of continental material to the south.

At this point we prefer to regard the early Proterozoic mobile belts of the North American midcontinent region as continental margin features, produced by plate-tectonic-like processes in a mobile lithosphere. Thus, many of the differences (compared to modern examples) are due to incomplete information, somewhat different geochemistry and petrology in the early Proterozoic, or both.

Origin of the middle Proterozoic granite-rhyolite terrane and undeformed granite plutons

Surely one of the most puzzling problems of the Precambrian geology of the midcontinent portion of North America is the origin of the vast terranes of middle Proterozoic granitic and rhyolitic rocks that have been described above. There is no easy way to establish a genetic correlation between the isolated undeformed plutons that occur sporadically within the foliated granitic and metamorphic rocks of the 1600—1800 Ma old terranes and the epizonal granites and rhyolitic volcanic rocks of the southern midcontinent. Most of the known undeformed plutons whose ages fall in the 1380—1480 Ma range are granitic. However, in at least one exposure of rocks of this suite, in the Needle Mountains of southwestern Colorado, Bickford et al. (1969) showed that the Electra Lake Gabbro was formed about 1425 Ma ago and is clearly part of a suite of undeformed plutons, mostly granitic, of this age that intruded foliated and deformed older rocks. However, such mafic rocks are relatively rare in exposed regions and they also appear to be rare in buried terranes. Thus, the plutons appear to be part of a mostly granitic suite and so are compositionally similar to the granite-rhyolite terrane to the south.

Age data strongly suggest that the epizonal granite-rhyolite terrane exposed in the St. Francois Mountains of Missouri and known over a large area of the southern part of that state in the subsurface (Bickford and Mose, 1975; Kisvarsanyi, 1979; Denison et al., in press) was formed at the same time as many of the undeformed plutons. For example, the age of 1470 ± 20 Ma for the St. Francois Mountains rocks is analytically the same as that reported for the Wolf River Batholith, a complex of mostly granitic rocks in Wisconsin (Van Schmus et al., 1975b). But to the south and west the available age data indicate that the rocks of the granite-rhyolite terrane are about 100 Ma younger. Undeformed plutons of this age, about 1380 ± 20 Ma, are not known to the north or west (Table 11-III). It is not clear whether the various parts of the granite-rhyolite terrane were formed as parts of a single time-transgressive event or whether they are related to two distinct events. In any case the resulting rocks are strongly similar. We believe that the preponderance of the available evidence indicates that the rocks of the granite-rhyolite terrane and the undeformed plutons to the north were formed during a major event between 1480 and 1380 Ma ago and that they are genetically related.

An important aspect of these rocks is that they appear to have formed anorogenically, at least in the sense that they are not associated with sedimentary rocks or their metamorphosed equivalents and that they show essentially no compressive deformation. As noted above, mafic rocks are rare in these terranes, as are also intermediate rock types. Andesitic rocks have been reported in minor amounts from the St. Francois Mountains (Tolman and Robertson, 1969; Berry and Bickford, 1972) and from the subsurface in southeastern Kansas and adjacent northeastern Oklahoma (Denison, 1966), but these and mafic rocks probably make up less than five percent of the total volume of the terrane.

The origin of the magmas that formed the rocks of the granite-rhyolite terrane and the suite of undeformed plutons is obscure and there are few isotopic or chemical data to shed any light on this problem. The extremely silicic compositions of most of these rocks would seem to preclude an origin directly from the mantle and so we have assumed that the magmas originated through fusion of older crustal materials. Unfortunately, the Rb-Sr system has been shown to be disturbed in the rocks of the St. Francois Mountains, the only group of these rocks whose isotopic system have been studied in detail (Bickford and Mose, 1975). Initial $^{87}Sr/^{86}Sr$ ratios for bodies yielding concordant Rb-Sr and U-Pb ages range from 0.7048 ± 0.0017 (Wolf River Batholith, Van Schmus et al., 1975b) and 0.7054 ± 0.0025 (Red Willow Batholith, Nebraska, Harrower, 1976) to 0.7078 ± 0.0010 (granite in the subsurface of Manitoulin Is., Ontario, Van Schmus et al., 1975a).

The question of the tectonic setting in which such a terrane of silicic rocks could form is central with regard to the possible operation of plate tectonic processes in the Precambrian. The position of the granite-rhyolite terrane as a more or less arcuate belt lying across the southeastern and south central parts of the continent (Fig. 11-5) and the presence of older rocks to the north and west and younger ones to the south and east strongly suggests that the terrane formed on or near the margin of an older continental mass. A model involving a convergent plate boundary on the eastern and southern edges of the continent, beginning about 1480 Ma ago and with a northwestward-dipping subduction system, could account for the arcuate pattern of the volcanic and shallow granitic rocks and the presence of the undeformed plutons intruding older rocks to the north and east of the volcanic belt; in this model the plutons would be formed from magmas generated above the north-dipping subduction zone and rising to invade older rocks to the north and east of the volcanic front. It seems evident, however, that whatever the process of formation of these rocks was, it was not similar to those in modern circum-Pacific tectonic settings, for there are almost no rocks of mafic or intermediate composition and compressional deformation and metamorphism are missing. If these rocks were formed by a continent-marginal process it was clearly one that was different from modern convergent plate boundaries.

We have also considered the possibility that the granite formed in an extensional environment (e.g. Lipman et al., 1972; Christiansen and Lipman, 1972; Snyder et al., 1976). As noted by several previous workers such rifted continental settings are commonly marked by abundant rhyolite. However, they also typically include large volumes of basalt and are characterized by a bimodal volcanic suite, whereas the granite-rhyolite terrane of the midcontinent is almost devoid of mafic rocks. Moreover, a rift model requires that the rift has two sides. In the case of the midcontinent granite-rhyolite terrane the rocks on the south and east side of the belt are younger than the granite-rhyolite assemblage. Only some complex history of rifting, removal of the southern and eastern side of the rift and, finally, emplacement of a new terrane of rocks about 1100 Ma ago could explain the present distribution of rock types and ages.

Dewey and Burke (1973) have suggested that the large volumes of potassic ignimbrites and shallow granite plutons of the Tibetan Plateau are the product of melting of the lower crust during collision between India and the Asian continent. This model could also be attractive for the origin of the granite-rhyolite terrane of the midcontinent were it not for the presence of younger, rather than older, rocks on the southern and eastern sides of the belt. As in the rift model, only a complex history could account for this fact if the granite-rhyolite assemblage were formed during a continent—continent collisional event.

CONCLUSIONS

In this paper we have presented our best assessment of the data available on the Proterozoic geology of the midcontinent region of North America. We have treated the northern midcontinent and the southern midcontinent separately, for the rock assemblages are somewhat different. For each region we have attempted some speculation concerning the origin of the rocks and in particular the possible role of plate-tectonic mechanisms in their evolution. Here we wish to examine the evidence presented by the whole region with regard to the larger question of continental evolution in the Proterozoic and the possible role of plate tectonics in that process.

There are two outstanding facts that come from an assessment of the Precambrian basement rocks of the midcontinent. The first is the age progression from the Archaean craton in the north through progressively younger rocks to the south. This sequence of younger and younger rocks toward the southern edge of the continent suggests a process of continental accretion and has been noted by others in the past. The second fact is the absence of island-arc rock assemblages in the northern midcontinent and the great abundance of granite and rhyolite in the southern part of this region.

We have clearly had to grope for adequate plate-tectonic mechanisms to account for these great terranes of dominantly granitic rocks. Nowhere are the characteristic rock types of modern arc-trench environments — the andesitic volcanics, greywackes or metagreywackes, ophiolite assemblages and calc-alkaline plutonic suites — really well represented except, perhaps, in the Penokean Fold Belt where the sedimentary sequences and calc-alkaline plutons are known.

A possible synthesis is the following: the older rocks of the midcontinent region, including the 1820—1860 Ma old igneous rocks associated with the Penokean Orogeny, possible 1700—1800 Ma old plutons related to those known in the Front Range of Colorado and the 1625 Ma old plutons in northern Kansas and Missouri may have been formed at successively younger convergent plate boundaries in a manner more or less similar to that in modern plate-tectonic regimes. The rock assemblages are reasonable in that tonalitic to granitic plutons and remnants of metavolcanic and metasedimentary assemblages are known. Following each of these orogenic episodes there appears to have been a period of widespread felsic volcanism. Thus, in southern Wisconsin there occur rhyolites and shallow granites that were formed 1760 Ma ago, some 80 Ma after the main calc-alkalic igneous activity of the orogenic phase; plutons of this age occur to the north of the volcanic rocks where they intrude older Proterozoic rocks. In the southern midcontinent the granite-rhyolite terrane began forming about 1480 Ma ago, some 145 Ma after the "orogenic" igneous activity represented by the 1625 Ma old foliated plutons. Plutons formed at this time (1480 Ma ago) occur as isolated undeformed bodies to the north of the volcanic terrane as intrusives in the older rocks.

We suggest that, following each of the major calc-alkalic igneous episodes associated with convergent plate tectonics, a period of crustal thickening and heating occurred; this period culminated with melting and production of potassic and silicic magmas in the lower crust. These magmas then rose buoyantly and produced widespread ignimbrite sheets that were invaded at shallow depth by granitic bodies of similar composition. The present succession of rock types, then, is the result of exposure of older and deeper crustal levels to the north, presumably as a result of continued uplift of the central portion of the continent.

We recognize that this model is highly speculative and is based upon imperfect observations in a region mostly covered with younger sedimentary rocks. Enough is known about this large region to show that the rock sequence outlined above is real and, so far as we know, unique among Proterozoic terranes. It seems clear, however, that processes operated that were different in detail from those operating at modern convergent plate boundaries. Perhaps our suggestions will spark the further work necessary to understand the tectonic setting of the Proterozoic evolution of this continent.

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Chapter 12

THE CIRCUM-SUPERIOR BELT: A PROTEROZOIC PLATE MARGIN?

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ABSTRACT

Nine discrete segments of late Aphebian rocks irregularly distributed around the Archaean Superior Province have sufficient characteristics in common to invite the idea of a Superior craton ringed by a continuous belt of mid-Proterozoic sedimentary and volcanic rocks. On the north side are the Labrador Trough, the Cape Smith Belt, the Belcher Basin, the Sutton Inlier, the Fox River Belt, and the Thompson Nickel Belt; on the south side are the Lake Superior Association (Animikie—Marquette Groups), the Mistassini—Otish Groups, and the Southern Labrador Trough. The bulk of the preserved rocks were probably deposited upon sialic basement as indicated by evidence from the Labrador Trough, Cape Smith Belt, and Lake Superior Association. The earliest deposits in some (Mistassini—Otish, Labrador Trough, and Belcher Basin) and possibly all segments are rift-related continental redbeds, with and without potash-rich basalts. In general the deposits change from mio- to eugeo-synclinal type in passing up in the sequence and outward from the Superior Province, and commonly a major iron formation marks approximately the level of change. Magmatism differs significantly in the two halves of the belt; in the north, ultramafic intrusives and extrusives are a substantial part of the magmatic assemblage, but calc-alkaline rocks are rare; in the south, ultramafic rocks are unkown and calc-alkaline rocks fairly common.

An origin for the Circum-Superior Belt may be postulated which is consistent with a plate-tectonic regime. Stretching of a sialic crust owing to plate motions produced fracturing around a pre-existing stable node, the ancestral Superior Province, followed by necking along the fractures to produce an annular trough in which sediments accumulated. With continued separation, volcanism was initiated at the loci of spreading and volcanic rocks overlapped onto the earlier sediments. Subsequently the southern part of the rift system expanded at the expense of the northern part into an opening ocean and subduction was initiated at the southern continental edge. The northern rift zone ceased activity and eventually closed with the deformation of its contents that we presently see.

INTRODUCTION

Proterozoic sequences of broadly similar age, stratigraphy, and lithologies are unevenly distributed around the margin of the Superior Province. Proof of their physical continuity, if it exists, is variously masked or obscured by the waters of Hudson Bay, the Palaeozoic cover, and the superimposed deformation of the Grenvillian Orogeny. Nevertheless, although each of the segments has its own distinctive make-up they have sufficient characteristics in common to compel the idea of a Superior craton encircled by a continuous belt of mid-Proterozoic sedimentary and volcanic rocks. We will call this the Circum-Superior Belt. Whether or not the Circum-Superior Belt is a plate boundary or intracratonic fill is a question that is bedevilled by conflicting evidence. In this paper we shall summarize the principal characteristics of each of the major segments of the Circum-Superior Belt and give our reasons for the projected correlation between them. Finally, we shall discuss the evidence relating to the thorny problem of its tectonic origin.

ELEMENTS OF THE CIRCUM-SUPERIOR BELT

General

Nine discrete segments distributed around the margin of the Superior Province comprise the visible elements of the Circum-Superior Belt. Some of these are tenuously linked by the configuration of gravity and magnetic anomalies over the covered regions that separate them, others are related by little more than their common age and fringing relationships to the Superior Province. When viewed collectively, all can be seen to have features which are common throughout or at least to two or more segments. Working anticlockwise around the margin of the Superior Province, the segments which compose the chain are as follows (Fig. 12-1): the Labrador Trough, the Cape Smith Belt, the Belcher Basin, the Sutton Inlier, the Fox River Belt, the Thompson Nickel Belt[1] (and Molson dykes), the Lake Superior Association (Animikie—Marquette Groups), the Mistassini—Otish Groups and the Southern Labrador Trough. The axes of positive and negative Bouguer gravity anomalies can be seen in Fig. 12-2 to link many of these segments beneath the intervening cover. Anomalies of the Lake Superior Association, although not shown to be linked with those of the rest of the chain on this diagram (interpreted from the Bouguer anomaly map of the United States, AGU, 1964), have recently been interpreted by Klasner and Bomke (1977, 1978) as being continuous with them. Thus, only in the Grenville Province is the continuity broken, and there the original structure can be expected to have been altered by the Grenvillian Orogeny.

The idea of correlating between different parts of this belt is not new: the Quebec—Ungava elements (Mistassini—Otish Groups, Labrador Trough, Cape Smith Belt and Belcher Basin) were variously correlated by Wahl (1953), Bergeron (1957a), Wilson (1968), and by Dimroth et al. (1970) who termed them the Circum-Ungava Geosyncline. The Thompson Nickel and Fox River Belts were related to the Cape Smith Belt by Wilson and Brisbin (1961) and by Gibb and Walcott (1971); and finally, a relationship

[1] The "Thompson Nickel Belt" here refers specifically to the Ospwagan group rocks (Scoates et al., 1977).

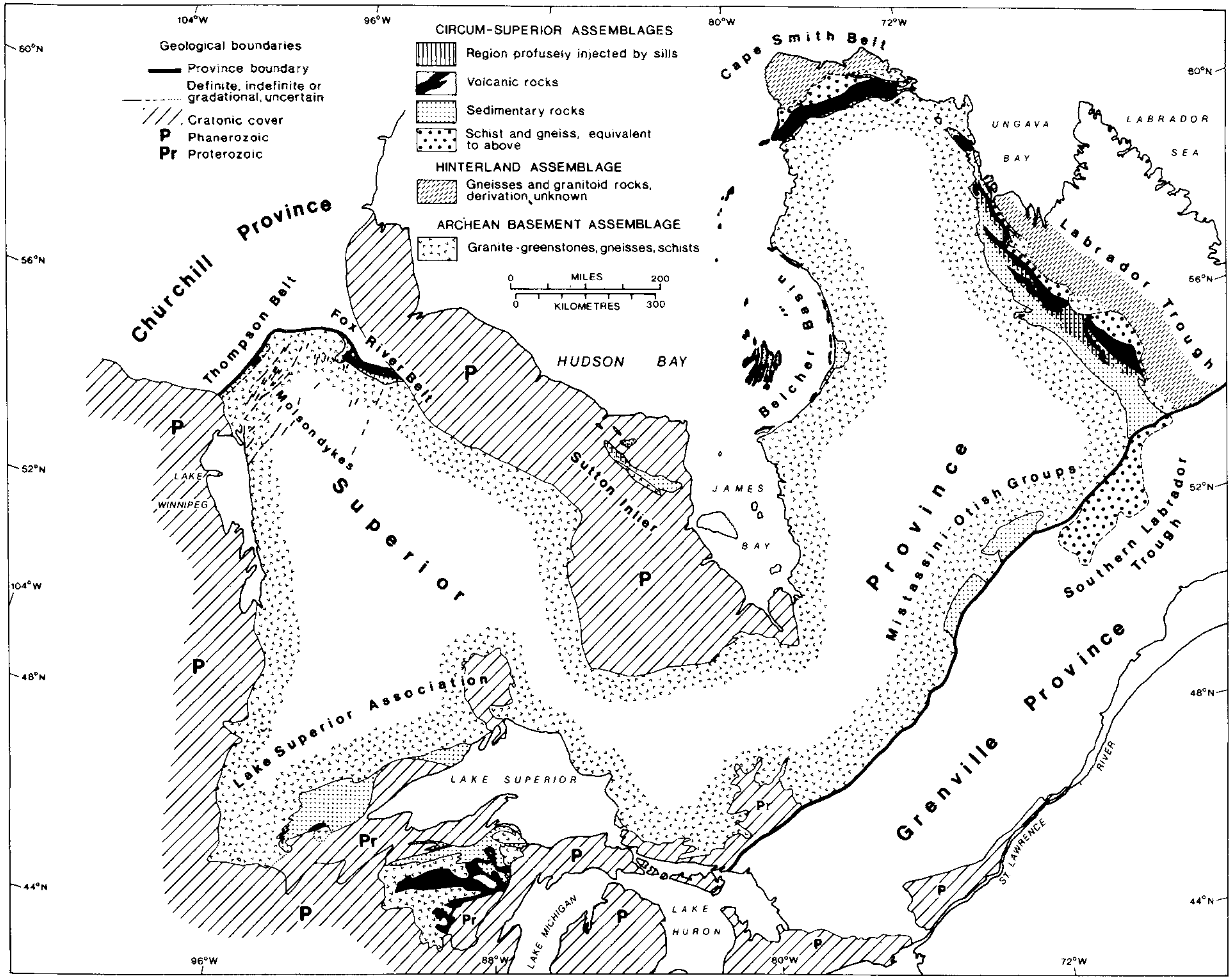

Fig. 12-1. Distribution of the exposed segments of the Circum-Superior Belt: Labrador Trough, Cape Smith Belt, Belcher Basin, Sutton Inlier, Fox River Belt, Thompson Nickel Belt, Lake Superior Association, Mistassini—Otish Groups and Southern Labrador Trough. In this and subsequent diagrams Thompson Nickel Belt is shortened to Thompson Belt.

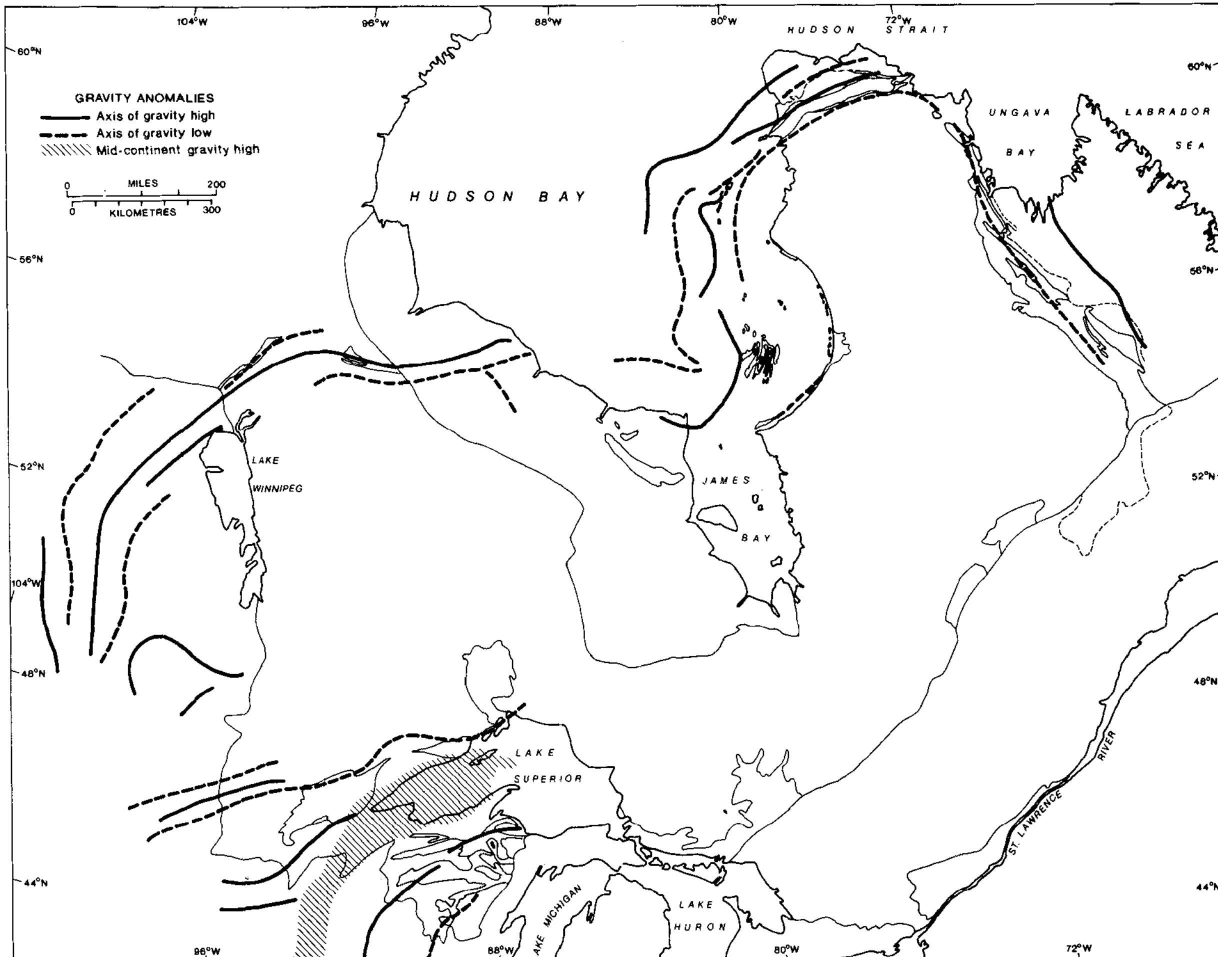

Fig. 12-2. Axes of gravity anomalies assumed to be related to the Circum-Superior Belt. The axes are interpreted from gravity maps as follows: American Geophysical Union (1964) and Department of Energy, Mines and Resources (1974, 1979).

between the Lake Superior Association and the Labrador Trough was envisaged by both Goldich (1973) and Sims (1976a). Thomas and Gibb (1977) introduced the term Circum-Superior suture to apply to the boundary between Churchill and Superior Provinces. Klasner and Bomke (1978), on the basis of the continuity of gravity anomalies, suggested the possible existence of an Aphebian fold belt around much of the Superior Province.

Age determination in rocks of the various segments of the Circum-Superior Belt are attended with some uncertainty. All strata are unconformable in their relationships to the Archaean Superior Province and all are affected by the Hudsonian or Penokean Orogeny. Hence, they are broadly Aphebian but radiometric age determinations on the rocks themselves range from about 2150 Ma B.P. (Rb-Sr, Schimann, 1978a) to 1590 Ma B.P.[1] (Rb-Sr, Brooks and Arndt, quoted in Schmidt, 1980). These are for volcanic rocks of the Cape Smith Belt and Belcher Basin, respectively. The Lake Superior Association appears to be the most reliably dated in that it is not only bracketed between the 2000 Ma B.P. age of underlying dykes (K-Ar, Hanson and Malhotra, 1971) and the 1840 Ma B.P. age of younger intrusives (Rb-Sr, Van Schmus, 1976) but is dated internally at about 1910 Ma B.P. (U-Pb, zircon from rhyolite; Banks and Van Schmus, 1972). The Circum-Ungava part of the belt was dated systematically by Fryer (1972) whose Rb-Sr ages for rocks of the Belcher Basin, Labrador Trough, and Mistassini Group ranged between 1705 and 1825 Ma B.P. All but one were done on sediments, the exception being the 1705 Ma age determination which was done on the upper volcanics of the Belcher Basin. The same volcanics yielded the 1590 Ma age reported by Schmidt. Older volcanics (Eskimo Formation) on the Belcher Islands, also dated by Brooks and Arndt and reported by Schmidt in the same paper, gave an age of 1700 Ma.

The positions of palaeomagnetic poles for each of these formations were found by Schmidt to be consistent with their respective ages as determined by Brooks and Arndt. Moreover, data for the younger formation gave a positive fold test, indicating that the polarity was fixed prior to folding, hence is likely to be primary. Therefore, unless one assumes that the Rb-Sr ages could be reset by a metamorphism that did not affect the magnetic polarity, the age of the Belcher Islands part of the belt should be interpreted as between 1600—1700 Ma B.P. The Fox River Belt recently yielded a preliminary Rb-Sr age of 1720 Ma for the Fox River Sill (Scoates and Clark, unpubl. data). This supersedes the age of 1610 Ma reported for this sill by Weber and Scoates (1978). The apparent spread in ages in different parts of the Circum-Superior Belt could indicate non-contemporaneity of emplacement; however, age determinations are so varied that it is difficult to have great confidence in any of them. Comparison of

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[1] All Rb-Sr ages are adjusted to be consistent with a decay constant of $1.42 \times 10^{-11} \times a^{-1}$.

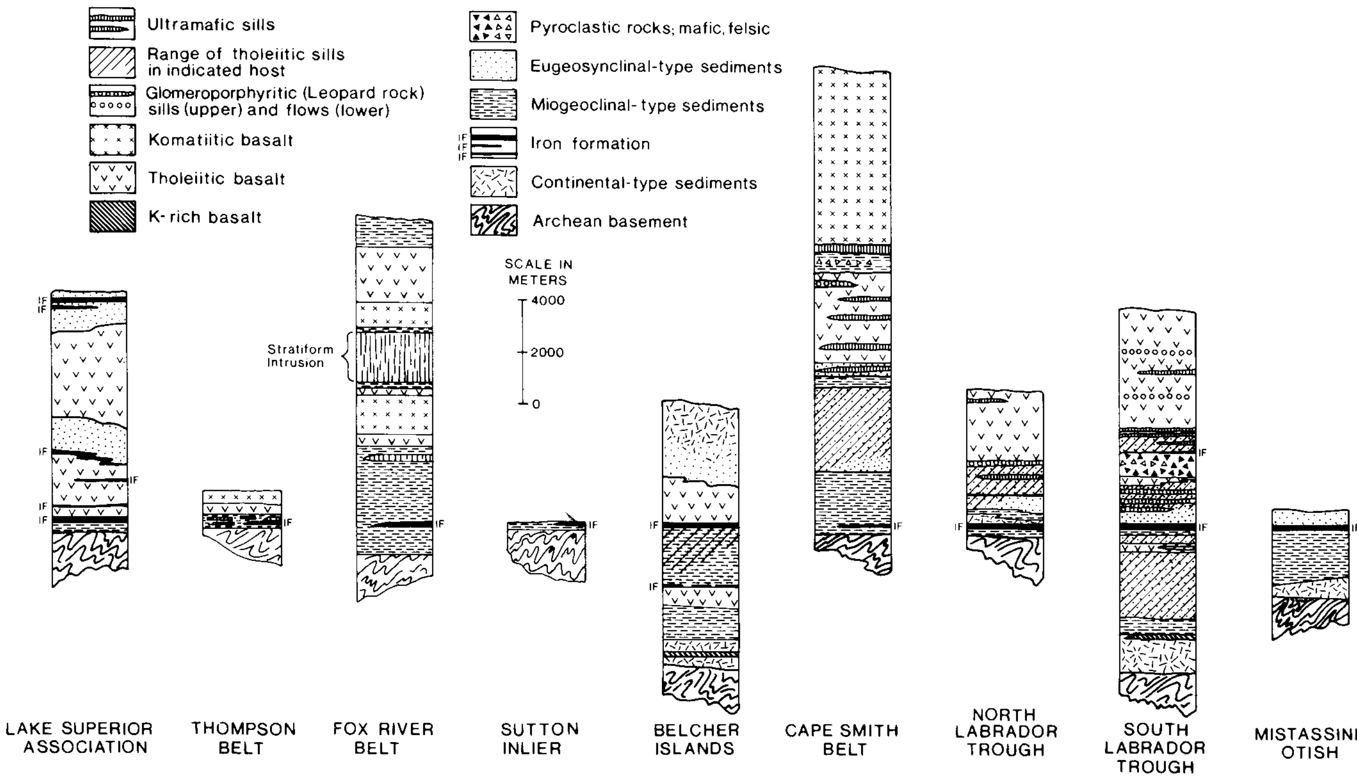

Fig. 12-3. Stratigraphic sections of each of the major segments of the Circum-Superior Belt. The thickness of individual units shown in the diagram represents an estimated average thickness or in some cases, where the unit is thin, a maximum thickness. Aggregate thicknesses, therefore, are not necessarily representative of any given section. Sources of information are given in the text. Note that South Labrador Trough refers to the southern part of the Labrador Trough, not the Southern Labrador Trough, as shown in Fig. 12-1.

lithologies and stratigraphies from segment to segment around the belt is probably a better basis for correlation at this time than are the radiometric ages.

Stratigraphic sections of the different parts of the Circum-Superior Belt are shown in Fig. 12-3. These are composite sections constructed to show the major stratigraphic elements in each segment. The thicknesses shown are generally the average in each segment, or even the maximum if the unit is thin, hence aggregate thicknesses may be somewhat artificial.

Labrador Trough

The Labrador Trough is well exposed and well studied; it would be fruitless to attempt here more than a very brief summary of the voluminous literature which it has inspired. Some of the more recent general references are as follows: Dimroth (1968, 1971, 1972), Dimroth et al. (1970); *see also Dimroth (this volume, Chapter 13) ed.*

On its west side sediments of the Labrador Trough rest unconformably on the Archaean basement. Its eastern boundary is less well defined. Generally it is placed along a string of faults that separate little-deformed Proterozoic rocks on the west from highly deformed rocks on the east. This is not, however, the eastern limit of the geosynclinal fill which is normally recognizable in schists that extend for an additional few tens of kilometres eastward. Farther east the identity of supracrustal rocks is generally obscured in the prevailing migmatites and gneisses. Jackson and Taylor (1972) believe that the Proterozoic rocks of the trough are preserved as relicts across the entire width of Hudsonian deformation (Churchill Province) to the Archaean strip (Nain Province of Taylor, 1971) along the coast. In this paper we will follow the tradition of restricting the term Labrador Trough to the little-deformed part where primary structures are readily recognized.

Metamorphism attendant upon the Hudsonian Orogeny increases in grade from west to east across the trough and into the schists and gneisses to the east. The grade along much of the western part of the trough is subgreenschist facies and rises eastward to amphibolite and locally granulite facies in the hinterland (Gélinas, 1965; Baragar, 1967; Dimroth and Dressler, 1978, Fraser et al., 1978). In its northern part the isograds divert westward across the strike of the trough and involve its entire width in the higher metamorphic grades.

The extent to which the Archaean basement underlies the Labrador Trough is a question of prime interest in regard to its tectonic setting. It unconformably underlies the western margin of the trough throughout its entire length and can be traced continuously around its north end west of Ungava Bay. Farther south its possible existence on the east side of the trough is obscured by severe deformation and metamorphism but has been identified in three places: near Leaf Bay in the northern trough where it has been identified by a relict Archaean age (Beall et al., 1963) and in the central and southern parts of the Labrador Trough where Dimroth (1964) and Wynne-Edwards (1960), respectively, observed granitic gneisses overlain by, and presumably contributing detritus to, basal arkoses and conglomerates of the trough sequence. In addition, several knobs of granitic basement emerge above the level of the Proterozoic rocks well within the boundaries of the trough.

The trough sequence is called the Kaniapiskau Supergroup and its deposition tends to be cyclic from stable to less stable regimes (Dimroth et al., 1970). Deposition began in the southern half of the trough with the accumulation of continental redbeds (sandstones, arkoses and conglomerates) and associated potash-rich basalts and andesites. These are related to faults and are reminiscent of a continental rift environment. They were succeeded by the first cyclic deposition which ranges upward from dolomites and sandstones to greywacke and shales accompanied in the medial and eastern parts of the trough by tholeiitic volcanism. The first cycle is confined to the southern half of the trough but in the stable period that followed, dolomite, quartzite, and iron formation spread over the rest of the trough and subsidence followed. The second cycle terminated in a major pyroclastic eruption throughout much of the trough and was succeeded by a second iron formation distributed unevenly along its length. Finally, the third cycle in all parts of the trough culminated in the eruption of a very thick assemblage of pillowed tholeiitic basalts. The most obvious characteristic of the stratigraphy of the Labrador Trough is the marked increase in the volcanic component upward in the sequence, both in individual cycles and from cycle to cycle. The volcanic rocks are predominantly tholeiitic basalts; felsic rocks are exceedingly minor and are mainly associated with the fragmental unit that caps the second cycle.

Doleritic sills of essentially identical composition to the basalts are a major component of the trough fill (Baragar, 1967; Dimroth et al., 1970). They can not be related to the volcanic rocks of a particular cycle except, of course, they must be stratigraphically lower than their surface equivalents. An exception are the distinctive glomeroporphyritic gabbros, known as leopard rock, which occupy a specific stratigraphic layer just below the dominantly volcanic part of the sequence and are represented in places as pillowed flows in the uppermost volcanic succession. Ultramafic sills present in the upper part of the sequence in both the southern and northern parts of the trough appear to represent the latest, or nearly the latest, magmatic activity.

The Kaniapiskau Supergroup was deformed and metamorphosed by the Hudsonian Orogeny during a period which, according to K-Ar ages, ranged from about 1600 to 1800 Ma ago. The deformation was directed towards the southwest and the Labrador Trough filling was folded, overturned, and thrust on to the stable mass of the Superior Province. Virtually no intrusive rocks were associated with the orogeny in the vicinity of the trough but Taylor (1977) has mapped major granitic plutons within the Churchill Province some 150 km east of the trough. These give an age of 1585 Ma (K-Ar).

Southern Labrador Trough

Southward the Labrador Trough crosses the Grenville Front and merges with the complexly deformed high-grade gneisses that are so characteristic of the Grenville Province. Insofar as it retains its identity, which it does with diminishing clarity as the distance from the crossing point increases, it is called the Southern Labrador Trough. Its characteristics have been summarized by Jackson in Dimroth et al. (1970). In the Kaniapiskau rocks the Grenville Front is marked by rapidly rising metamorphic grade and an increasingly complex structure due to the imposition of northeasterly trending Grenvillian folds on to a pre-existing northwesterly fold pattern (Gastil and Knowles, 1960). Immediately south of the front the Kaniapiskau strata swing southwesterly and can be traced in a direction roughly parallel to the front for about 300 km. Farther southwest they can not be recognized.

The Mistassini—Otish Groups

These are partly correlative groups which overlie Archaean basement and flank the Grenville Front some 300 km southwest of the Labrador Trough. They have been described by Bergeron (1957a) and Chown and Caty (1973). The succession is only gently folded and begins with current-bedded quartzitic and arkosic sandstones and conglomerates derived from Archaean terrane to the north and is succeeded by dolomite and finally iron formation, with associated quartzite and shale. The basin is truncated along the Grenville Front by a high-angle reverse fault and now abuts against gneisses of the Grenville Province. It is evidently part of a much larger basin that extended southward into the region of the Grenville Province. The obvious choice for its southern extension is the highly metamorphosed succession of the Southern Labrador Trough on the other side of the front, a few km to the southeast.

The Cape Smith Belt

The Cape Smith Belt is an obvious extension of the Labrador Trough with at least some of the differences readily attributed to differences in degree of preservation. For example, the paucity of miogeoclinal-type sediments, including iron formation (Figs. 12-1 and 12-4) can be assumed to be the result of its removal from the foreland area by erosion. There has been

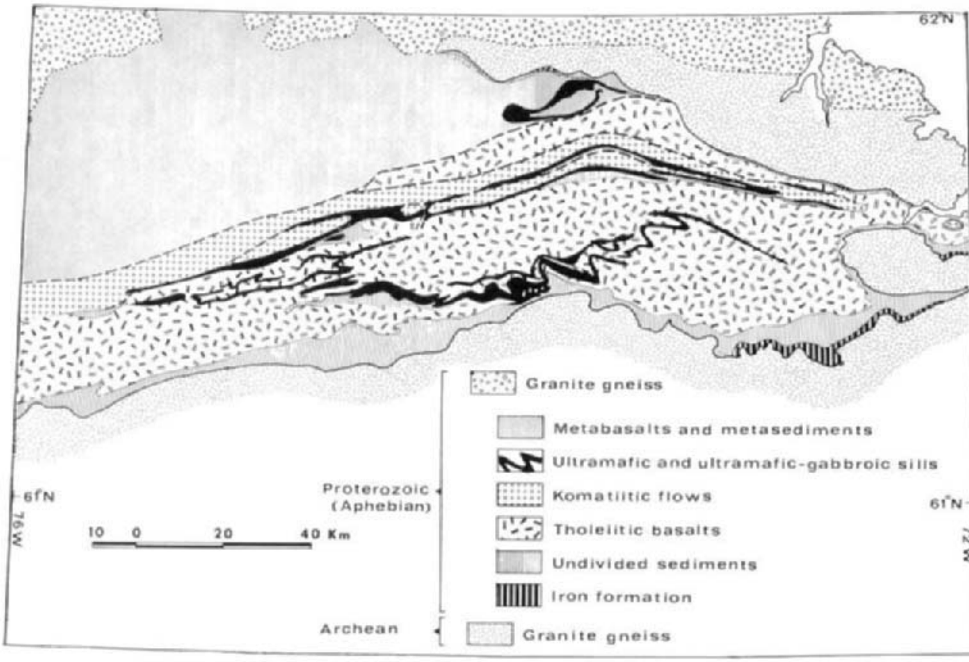

Fig. 12-4. Compilation geological map for the eastern half of the Cape Smith Belt. Sources of information are given in the text.

little quarrel over the years with Bergeron's (1957a) original correlation of the two belts made at a time when mapping in both was at a very rudimentary stage.

The geology of the Cape Smith Belt is largely attributable to the work of the Quebec Department of Natural Resources (DeMontigney, 1956; Bergeron, 1957b, 1959; Beall, 1959, 1960; Gold, 1962; Gélinas, 1962; and Schimann, 1978a) supplemented in more recent years by reconnaissance mapping (Taylor, 1974) and special studies (Baragar, 1974; Schwarz and Fujiwara, 1977; Miller, 1977; Francis and Hynes, 1979). The compilation map of the eastern part of the Cape Smith Belt, shown in Fig. 12-4, is based upon information drawn from all these sources.

The Cape Smith succession rests unconformably on the Archaean foreland along most of its southern margin. At its east end the Archaean basement passes around the belt and extends for a considerable distance along its northern side before losing its identity in the gneisses of the hinterland (Gélinas, 1962; Taylor, 1974). That the belt is a true intracratonic syncline is evidenced by the continuation around its eastern end of the lowermost stratigraphic units of the Cape Smith sequence for at least as far as the basement is recognized.

As in the Labrador Trough, metamorphism in the Cape Smith Belt generally increases in intensity outward (northward) from the Ungava craton (Westra, 1978) but is complicated by the westerly plunge of the belt, which results in deeper levels being exposed at its eastern than at its western end, and by localized shearing in the lowermost members of the sequence adjoining the contact. The major part of the belt is composed of relatively little deformed rocks which are separated from highly deformed equivalents to the north by a line of steeply-dipping faults that extends along most of the length of the belt. The metamorphic grade rises from subgreenschist facies at the western end of the belt south of this line of faults (Baragar, unpubl. data) to granulite facies in the hinterland near the northern end of the Ungava peninsula (Westra, 1978). Like that in the Labrador Trough, it is of intermediate pressure type (Westra, 1978; Schimann, 1978b). Gneisses in the hinterland are of uncertain derivation but could be reworked Archaean basement (Westra, 1978).

The stratigraphy of the Cape Smith Belt is similar to that of the Labrador Trough except that volcanic rocks predominate. The sequence, based on the least-deformed rocks south of the fault line, can be broadly divided into three parts: a lower sedimentary division, a middle division of tholeiitic basalts, and an upper division of komatiitic basalts. The sedimentary succession begins with a stable platform assemblage of dolomite, quartzite, shale, rare volcanics and irregularly occurring iron formation in the eastern part of the belt and is succeeded by variable thicknesses of mostly greywacke and shale profusely intruded by mafic and less commonly ultramafic sills. Massive mafic flows, difficult to distinguish from the sills, are also present but for obvious reasons their proportion is not easily estimated. Upward the sediments interfinger with tholeiitic basalts of the middle division. These occur as both pillowed and massive flows but are generally amygdaloidal and judged to have been extruded in fairly shallow water. The middle and upper divisions are separated throughout the belt by a layer of sediments, mostly quartzites, shales, and locally dolomites, and intermixed pyroclastic rocks of intermediate to felsic composition. The upper division of komatiitic basalts is predominantly pillowed and, in contrast to the tholeiitic basalts, is rarely amygdaloidal. Presumably it was extruded in deep water. Spinifex texture, so characteristic of many Archaean komatiitic flows, has not been observed in this belt. In the western part of the belt the komatiitic division is structurally overlain by a further division of tholeiitic basalts. These are almost identical in composition to the tholeiitic basalts underlying the komatiites and are tentatively interpreted as being the middle division repeated by faulting. Although this sequence is based mainly on observations made in the central and western parts of the belt, it appears to be generally applicable throughout, judging from the very similar succession described by Schimann (1978a) at its eastern end.

Ultramafic sills are present at several levels but are mainly confined to the eastern half of the belt. They are commonly composite in the sense that the borders range from pyroxenitic to basaltic whereas the interior is peridotitic. In the thicker sills a gabbroic layer may intervene between the peridotite and upper margin. Francis and Hynes (1979) suggest that such bodies formed by crystal fractionation of an originally komatiitic magma. The liquid line of descent in one sill that they studied appears to parallel that of the tholeiitic and komatiitic lavas which compose the Cape Smith succession. Hence, they postulate that the variation in lava composition is achieved by just such fractional crystallization.

Leopard rock, identical in appearance to that in the Labrador Trough, is recognized in one locality in the western part of the Cape Smith Belt where it is interlayered with the tholeiitic basalts that structurally overlie the komatiites. Assuming this is a faulted segment of the main tholeiitic division, the stratigraphic position of the leopard rock is interpreted as being within the lower tholeiitic sequence (Fig. 12-3), approximately equivalent to its position in the Labrador Trough.

Felsic volcanic rocks are minor and scattered in the Cape Smith Belt (Taylor, 1974); most occurrences seem to predate the komatiites. However, the sedimentary layer immediately underlying the komatiites appears to represent the most persistent level of felsic volcanism in the entire sequence (cf. Schimann, 1978a).

Schistose volcanic and sedimentary rocks north of the east—west line of faults dividing the belt longitudinally are of uncertain stratigraphic position because of the severe deformation which has obscured most of the primary features. The few analyses obtained from mafic schists in this part of the belt are indicative of tholeiitic rather than of komatiitic compositions and, moreover, felsic rocks are present in a number of places. For these reasons the schists are belived to be equivalent to the lower part of the sequence. This was also Bergeron's (1959) earlier interpretation.

Bergeron (1957b) and later other workers subdivided the Cape Smith succession into lower (Povungnituk) and upper (Chukotat) groups separated by an unconformity. In part the "unconformity" coincides with the division between tholeiitic and komatiitic basalts of this paper but recent work has failed to confirm the presence of an unconformity between these two divisions (Taylor, 1974; Baragar, 1974; Schimann, 1978a) and they are considered here as a conformable sequence.

Between Cape Smith and the Belcher Basin (Fig. 12-1) the Circum-Superior Belt is exposed in a string of islands, the Ottawa Islands, along the eastern side of Hudson Bay. There the komatiite member is well exposed (Baragar and Lamontagne, 1980) and includes spinifex-textured flows virtually identical in appearance to the classic flows of Munro Township, Ontario (cf. Pyke et al., 1973; Arndt et al., 1979).

The Belcher Basin

The Belcher Basin comprises the Aphebian deposits of the mainland, the immediate offshore islands and the Belcher Islands, all in the great circular bight of eastern Hudson Bay. Following the mapping of the Belcher Islands by Jackson (1960) and of Richmond Gulf area on the mainland by Woodcock (1960) the geology of the Belcher Basin was detailed by a host of recent theses (Leggett, 1974; Barrett, 1975; Stirbys, 1975; Hews, 1976; Ware, 1978; and Ricketts, 1979) and special studies (Hofmann and Jackson, 1969; Bell and Hofmann, 1974; Chandler, 1978, 1979; and Schmidt, 1980). Jackson, in Dimroth et al. (1970) provides the most comprehensive interpretation of the geology to date and an extensive bibliography of earlier work.

The base of the succession is exposed in the vicinity of Lac Guillaume Delisle (Richmond Gulf) on the mainland where it overlies the Archaean basement and dips gently westward. The first deposits are terrestrial arkoses and potash-rich basalts associated with east—west graben development (Woodcock, 1960; Hews, 1976; Chandler, 1978). In this respect the early development of the Belcher Basin parallels that of the Labrador Trough. These are overlain by shallow marine deposits of a stable platform type which persist through much of the subsequent sequence. Strata on the mainland and the immediate offshore islands are correlated by Jackson (Dimroth et al., 1970) with the Belcher Island succession on the basis of certain distinctive lithologies, particularly iron formation and volcanic units. The miogeoclinal part of the succession (the lower part) has much the same stratigraphic range in both places but is of greater thickness on the Belcher Islands. It generally comprises dolomites, quartzites and shales, but includes a sequence of massive tholeiitic basalts (Stirbys, 1975) and two iron formations, the upper of which is the major one and marks the termination of the platformal succession. Except for the early terrestrial deposits, which Chandler (1979) found were derived from the west, current direction studies indicate an eastern source for sediments of the platformal sequence (Barrett, 1975). Consistent with this is a generally westward thinning of the volcanic formation contained within the sequence.

Following deposition of the major iron formation the source region switched to the west of the Belcher Islands and the character of the deposits changed to eugeosynclinal type. A sequence of tholeiitic basalts thickening westward covers the iron formation and is succeeded by greywackes and shales. Upward in the sequence these latter deposits become increasingly arkosic and eventually are topped by red terrigenous arkoses and conglomerates. From the volcanic sequence upward, paleoslope studies (Dimroth et al., 1970; Ware, 1978) indicate a source region in the west. Bell and Hofmann (1974) have suggested that the beginning of the Hudsonian Orogeny was marked in this region by the upper volcanics which accompanied basin collapse and the subsequent rise of tectonic lands to the west. The appearance of arkosic sediments at the top of the succession most likely signifies the uncovering and degradation of either orogenic plutons or cratonic basement in the region of Hudson Bay west of the Belcher Islands.

The upper tholeiitic basalts of the Belcher Basin are very similar to those which underlie the komatiites in the Cape Smith Belt. Both are a mix of amygdaloidal, pillowed and massive basalts, indicative of eruption in shallow waters.

The "leopard rock" of the Labrador Trough may have its equivalent in a very similar coarsely feldsparphyric gabbro that occurs near the base of the upper tholeiitic sequence in the Sleeper Islands, just north of the Belcher Islands (Baragar and Lamontagne, 1980).

The metamorphic grade is subgreenschist facies throughout the Belcher Basin. On the Belcher Islands it is manifested by the presence of prehnite and pumpellyite (Leggett, 1974; Stirbys, 1975). Folding is of a type characterized by broad synclines and sharp narrow anticlines (Jackson, 1960); unlike folding in the Labrador Trough, axial planes are upright and there is little evidence that the geosynclinal filling was pressed on to the craton. Jackson (in Dimroth et al., 1970) suggested that the folding was related to a décollement between the supracrustal deposits and the basement.

Sutton Inlier

Inliers of Archaean and Proterozoic rocks exposed through the Palaeozoic rocks of the Hudson Bay Lowlands near Sutton Lake in northern Ontario (Bostock, 1971) are known as the Sutton Inlier. Correlation of the Proterozoic rocks with those in the Belcher Basin was suggested by Dowling (1905), Leith (1910), and Hawley (1926). Bostock (1971) re-affirmed the correlation and suggested that the Proterozoic of the Sutton Inlier forms part of the Circum-Ungava geosyncline.

Archaean rocks of the inlier occur in its southernmost part and are composed of granodiorite gneiss within a complex of granitic rocks. Proterozoic rocks unconformably overlie the Archaean gneiss and comprise a lower carbonate unit, an upper unit of chiefly greywacke, siltstone and iron formation, and a unit composed of basic sills intrusive into the sedimentary rocks. The Proterozoic rocks are undeformed, they are essentially unmetamorphosed and many are nearly flat-lying. As a result, their exposed thickness is only about 300 m.

The Fox River Belt

The Fox River Belt (Scoates, 1977, and in prep.) borders the northeast part of the Superior Province craton in Manitoba for approximately 300 km. The belt, which consists of sedimentary rocks, large differentiated sills, and ultramafic to mafic volcanic rocks, forms a 15 to 20 km wide north-facing homoclinal sequence that is interpreted to have been deposited upon Superior Province gneiss. The presence of prehnite and pumpellyite indicates subgreenschist facies metamorphic conditions in volcanic rocks of the upper part of the sequence, whereas those in its lower part are of lowermost greenschist facies grade.

Sedimentary rocks occur in three stratigraphic positions and are referred to informally as the Lower, Middle and Upper sedimentary formation. The Lower sedimentary formation is estimated to be 5 km thick in the western part of the belt where it consists of well-laminated siltstone, argillite, and shale interlayered with sandstone, quartzite, and dolomite. Much of the siltstone and argillite is calcareous. Specular hematite-, magnetite- and pyrite-bearing iron formation is locally developed, and significant concentrations of iron formation are interpreted as occurring in three widely separated areas of the formation. The uppermost part of the formation contains carbonaceous shale. Very little information is available from the eastern part of the belt.

The Middle sedimentary formation is the host to the Fox River Sill, a major differentiated stratiform intrusion, and the rocks of the formation occur as narrow layers along the north and south contacts of the sill. The formation is estimated to be 1 km thick. The rocks, which have been substantially modified by contact metamorphism, originally comprised a finely laminated sequence of quartz-rich siltstone, sandstone and feldspathic sandstone, with minor argillite.

The Upper sedimentary formation is estimated to be 1 km thick and consists of argillite, shale and carbonaceous shale.

There is an overall similarity among the rocks of the sedimentary formations. All rocks are fine grained and originally consisted chiefly of quartz, clay minerals and carbonate minerals. Variations in the abundance of hematite, magnetite and pyrite give rise to ferruginous rocks or iron formation. There is a distinct absence of lithic components.

Volcanic rocks intercalated with the sedimentary formations are referred to informally as the Lower and Upper formations. They are approximately 3 km and 2 km thick, respectively. Outcrop and drill hole data provide a source of information for the volcanic rocks over a strike length of approximately 50 km.

In the western part of the belt the Lower volcanic formation consists of a lower massive zone (750 m), a middle pillowed zone (1150 m), and an upper massive zone (400 m). The lower massive zone comprises basalt and komatiitic basalt characterized by pyroxene spinifex and differentiated layered sequences interpreted as layered flows. The latter range in composition from peridotite through pyroxenite and gabbro to fine-grained, highly recrystallized, brecciated and vesiculated rocks interpreted as flow tops. The middle pillowed zone consists of pillowed basalt and komatiitic basalt with some intercalated massive and composite flows, and the rocks range from plagioclase-bearing olivine clinopyroxenites with rare interpillow space near the base, to basalts with smaller pillows but greater interpillow space near the top. Composite flows, from 5 to 15 m thick, comprise a lower cumulus zone and an upper fine-grained columnar jointed zone capped by a vesiculated and brecciated flow top (Scoates, 1977). The upper massive zone consists mostly of homogeneous, massive basalt flows with vesiculated tops. Many display flow top breccias. Laminated, sulphide-bearing carbonaceous sedimentary rocks, as much as 10 m thick, are interlayered with the volcanic rocks.

The rocks of the Lower volcanic formation display a progressive change in character upward in the sequence. Olivine-rich cumulus rocks occur in the differentiated, layered flows of the lower massive zone. These are intercalated with pyroxene spinifex textured flows and they give way to basaltic flows in the upper part of the middle zone. The massive flows of the upper zone are plagioclasephyric basalts.

Information on the Upper volcanic formation is scarce but also indicates a regular change from pyroxenitic flows near the base to basaltic flows near the top.

Large differentiated sills intrude the upper part of the Lower sedimentary formation, and the Fox River Sill intrudes the Middle sedimentary formation. The former range from 1.5 to 20 km long, average 800 m thick, and each sill comprises five zones: a pyroxene-rich contact zone, a peridotite zone, a clinopyroxenite zone, a gabbro zone and a hybrid, granophyre-bearing roof zone. The Fox River Sill forms western and eastern segments, each about 70 km long, that are separated by a gap of 12 km. Distinctive aeromagnetic anomalies east of the eastern segment indicate that ultramafic rocks extend for another 100 km eastward beneath the cover rocks. The thickness of the sill in the western segment is estimated to average 2 km at the present erosional surface. The sill is subdivided into four zones: a marginal zone, a lower central layered zone, an upper central layered zone, and a hybrid roof zone. Each zone is characterized by distinctive lithologic units and, except for the hybrid roof zone, by a distinctive cyclic arrangement of units. Rock types range from dunite to mafic granophyre.

The progressive change in character of the volcanic rocks from the base to the top of the Lower volcanic formation suggests derivation of these rocks from a differentiating source (Scoates, in prep.). The obvious choice for such a source are the large differentiated sills that intrude the upper part of the underlying Lower sedimentary formation. The scanty evidence available on the Upper volcanic formation indicates a similar variation in that sequence which by the same reasoning could be attributed to differentiation in the Fox River Sill.

Thompson Nickel Belt

Supracrustal rocks of the Thompson Nickel Belt named the Ospwagan

Group (Scoates et al., 1977) comprise metasedimentary, metavolcanic and ultramafic rocks that occur as folded sequences in overprinted Archaean migmatites of the Superior Province. Pillowed and massive mafic flows are present from near Moak Lake southward to Mystery and Ospwagan Lakes. Similar volcanic rocks occur on Setting Lake at the south end of the belt. Pyroxene spinifex textures have been observed in massive flows on Mystery and Ospwagan Lakes. The flows range in composition from ultramafic varieties (metapicrite of Stephenson, 1974) to basalt. The metasedimentary rocks consist of laminated siltstone, sandstone, quartzite, shale, phyllite, dolomite and iron formation. Minor chert and greywacke are present in places. Ferruginous varieties of siltstone, shale and phyllite are common. Ultramafic rocks, which are dominantly serpentinites, occur as numerous pods and sill-like masses in Ospwagan group rocks and migmatitic gneiss. They are concentrated along the western margin of the Thompson Nickel Belt and their great abundance is one of its prime characteristics. Ospwagan group rocks have been deformed, metamorphosed and intruded by granitoid plutons during a late Hudsonian event. The severe deformation has to date precluded the establishment of stratigraphic relations among the units of the group.

There is a strong similarity between the Ospwagan group of the Thompson Nickel Belt and rocks of the Fox River Belt. Fine-grained, laminated, quartz-rich siltstones are common to both suites, as is an abundance of ferruginous rocks and the presence of dolomitic beds. The volcanic rocks display the same range in composition and pyroxene spinifex-textured rock has been observed in both belts. The most significant feature which they possess in common, however, is their great abundance of ultramafic rocks. In the Fox River Belt, the Fox River Sill and lower differentiated intrusions contain substantial volumes of serpentinized peridotite and dunite whereas the Thompson Nickel Belt is characterized by numerous large and small pods and sill-like masses of serpentinite, some of which are differentiated (Peredery, 1979).

The age of deposition of rocks of the Ospwagan group and Fox River Belt has not been unequivocally established. The Ospwagan group rocks are only known to post-date the last Kenoran event and to pre-date late Hudsonian deformation. The relatively unmetamorphosed and undeformed rocks of the Fox River Belt contrast with the deformed and metamorphosed (lower to upper amphibolite facies) Kisseynew-type gneiss of the adjacent Churchill Province (Scoates, 1977). This could be interpreted to indicate that Fox River Belt rocks were deposited after the peak of metamorphism and deformation in the adjacent Churchill Province had been accomplished. Alternatively, it could be argued that the Fox River Belt rocks were faulted against the higher grade rocks of the Churchill Province. However, the gneiss of the Superior Province is at least partly retrogressed granulite. If the retrogression of the granulite is related to the main metamorphic event in the adjacent Churchill Province then it is difficult to see how the Fox River Belt rocks could escape the effects of that event. The problem is still unresolved.

Molson Dykes

The Molson Dykes constitute an extraordinary swarm of northeasterly-striking dykes in northern Manitoba (Fig. 12-1) which, by reason of their association, composition and age, would seem to bear some relationship to the Circum-Superior Belt. They have been studied in some detail by Ermanovics and Fahrig (1975) and Scoates and Macek (1978). This dyke swarm parallels the Thompson Nickel Belt and the concentration of dykes increases towards it. Their composition ranges from mafic to ultramafic as is the case for igneous rocks of the Thompson Nickel and Fox River Belts and their age, although not yet precisely determined, is in the same range as is that of these belts. They post-date late Archaean metamorphism in the Superior Province and near the Thompson Nickel Belt are affected by the same metamorphism which affected the latter. On the basis of palaeomagnetic work Ermanovics and Fahrig (1975) suggested an age of 1800—2000 Ma for the magnetism of the dykes.

Lake Superior Association

The Lake Superior Association, as it is called in this paper, comprises the broadly equivalent Animikie and Marquette Range Supergroups. It is the most remote and tenuously connected element of the Circum-Superior Belt. Nevertheless, in its setting and stratigraphy it possesses the unmistakable mark of the Circum-Superior family. The literature of this region is extraordinarily voluminous and some recent general papers which provide excellent summaries of various aspects of the geology are as follows: Sims and Morey (1972), Bayley and James (1973), Cannon (1973), Morey (1973, 1978), Sims (1976a) and Van Schmus (1976).

The Aphebian deposits of the Lake Superior region overlie and tend to be distributed along a join between two fundamentally different terranes in the Archaean basement (Sims, 1976a; Morey and Sims, 1976). The join is a west-southwest-trending structural discontinuity which may have been a fault and possibly provided a depression in the Archaean surface in which the Aphebian deposits accumulated. Northwest of the join the terrane comprises 2700—2750 Ma old granite-greenstone complexes characteristic of much of the Superior Province; southeast of it the rocks are predominantly migmatitic gneisses of generally high metamorphic grade and complex history. Along the Minnesota River Valley the gneisses yield ages as old as 3800 Ma and appear to have been reworked at 3000, 2600, and 1850 Ma (Goldich et al., 1970; Goldich and Hedge, 1974; Hedge and Goldich, 1976)

In central Wisconsin gneisses which can be interpreted as an eastern extension of those in Minnesota were recently shown by Van Schmus and Anderson (1977) to have Rb-Sr isotopic compositions consistent with an age of 3500 Ma or greater. Thus the ancient gneisses could have formed a continuous mass around the southern margin of the Lake Superior Association. This is consistent with a widely held view that the Lake Superior deposits were emplaced upon a sialic crust (Cannon, 1973; Sims, 1976a; Morey and Sims, 1976; Morey, 1978).

Distribution of material within the Lake Superior Aphebian basin is not symmetrical with respect to its longitudinal axis; the north side is composed primarily of platformal or miogeoclinal types of deposits and the south side of eugeosynclinal types. In Minnesota the Animikie Group is predominantly of miogeoclinal type, whereas in Michigan a complete transition is present from miogeoclinal on the north to eugeosynclinal on the south (Sims, 1976a). The columnar section shown in Fig. 12-3 is compiled mostly from the Michigan part of the basin (particularly James, 1958). Stratigraphically the miogeoclinal deposits compose the lower and thinner parts of the succession and eugeosynclinal the upper and thicker parts. The transition, as in other parts of the Circum-Superior Belt, immediately follows deposition of the widespread iron formation correlative throughout the Lake Superior area, although not necessarily continuous (Bayley and James, 1973). Other iron formations are present in the eugeosynclinal part of the sequence but none appears to be so widespread as that which marks the termination of the platformal environment. The eugeosynclinal succession comprises thick assemblages of volcanic rocks as well as greywacke and shale. Basalts form the bulk of the volcanic rocks and are commonly pillowed but in places they are accompanied by felsic rocks some of which contain massive sulphide deposits (Sims, 1976b; Mudrey, 1979). Both tholeiitic and calc-alkaline characteristics are represented in these volcanic rocks according to recent studies by Cudzillo (1978) and Bowden (1978) (reported in LaBerge and Mudrey, 1979).

The Aphebian basin received contributions from basement sources on both sides of it according to the evidence accumulated by Sims (1976a). In Minnesota and Michigan the source area for at least part of the sediments on the north limb of the basin is identifiable as the granite-greenstone terrane to the north. Farther south in the basin scattered evidence variously suggests transport of sediments towards the north and derivation from local or more southerly basement sources (Sims, 1976a).

Aphebian deposition in the Lake Superior area terminated with the Penokean Orogeny (Goldich et al., 1961; Goldich, 1972; Cannon, 1973) manifested by deformation, metamorphism and intrusion. The deformation is not typically orogenic. According to Cannon (1973) it results not from horizontal compression but from gravitational sliding of the Aphebian cover on a tilted Archaean basement. This is followed by block faulting of the basement with resulting passive deformation of the overlying cover. The metamorphism defines a number of discrete high-temperature nodes (James, 1955) related to thermal domes associated with reactivated basement or uplifted blocks (Cannon, 1973; Sims, 1976a). Both syn- and post-tectonic intrusions accompanied the Penokean Orogeny (Van Schmus, 1976); the former range in composition from diorite to granite and are about 1820—1840 Ma old and the latter, mainly granites and contemporaneous rhyolites (partly ignimbritic), are about 70 Ma younger at 1760 Ma (Van Schmus, 1978). The age of the Penokean Orogeny is generally accepted as being roughly the interval 1820—1850 Ma which is a little out of phase with the Hudsonian Orogeny in the rest of the Circum-Superior Belt (assumed to terminate at c. 1800 Ma according to Stockwell (1973) but somewhat younger according to age determinations previously noted). However, the post-orogenic granites and rhyolites referred to above as well as a regional overprint at c. 1630 Ma ago which affects Rb-Sr ages in the Lake Superior Aphebian basin (Van Schmus, 1976, 1978), indicate that the thermal regime of the Penokean Orogeny overlapped the period of the Hudsonian Orogeny. Furthermore, given the uncertainty in the absolute dating of Hudsonian deformation as opposed to its thermal culmination, there is little reason to believe that they could not be penecontemporaneous.

CORRELATION WITHIN THE CIRCUM-SUPERIOR BELT

General

Correlation between various segments of the Circum-Superior Belt depends more upon a combination of factors than upon any single component. Nevertheless, Lake Superior-type iron formation (Gross, 1965), is one distinctive element that is common to all segments of the belt and is therefore used as the basis of the correlation shown in the stratigraphic sections of Fig. 12-3. The distribution of the iron formation and some of the other distinctive lithologies around the Circum-Superior Belt is shown in the map of Fig. 12-5. Factors important to both correlation and to interpretation of tectonic development in the Circum-Superior Belt can be discussed in terms of stratigraphic and magmatic relationships.

Stratigraphic relationships

Stratigraphic evolution in all segments of the Circum-Superior Belt is broadly similar in that it changes upward in the sequence without recognizable unconformity from mainly mio- to mainly eugeosynclinal type (Fig. 12-3). The major iron formation appears at or close to the level of change. Mafic and, in some segments, ultramafic igneous rocks dominate the upper parts of the sequence. In most places the miogeoclinal sequence underlying

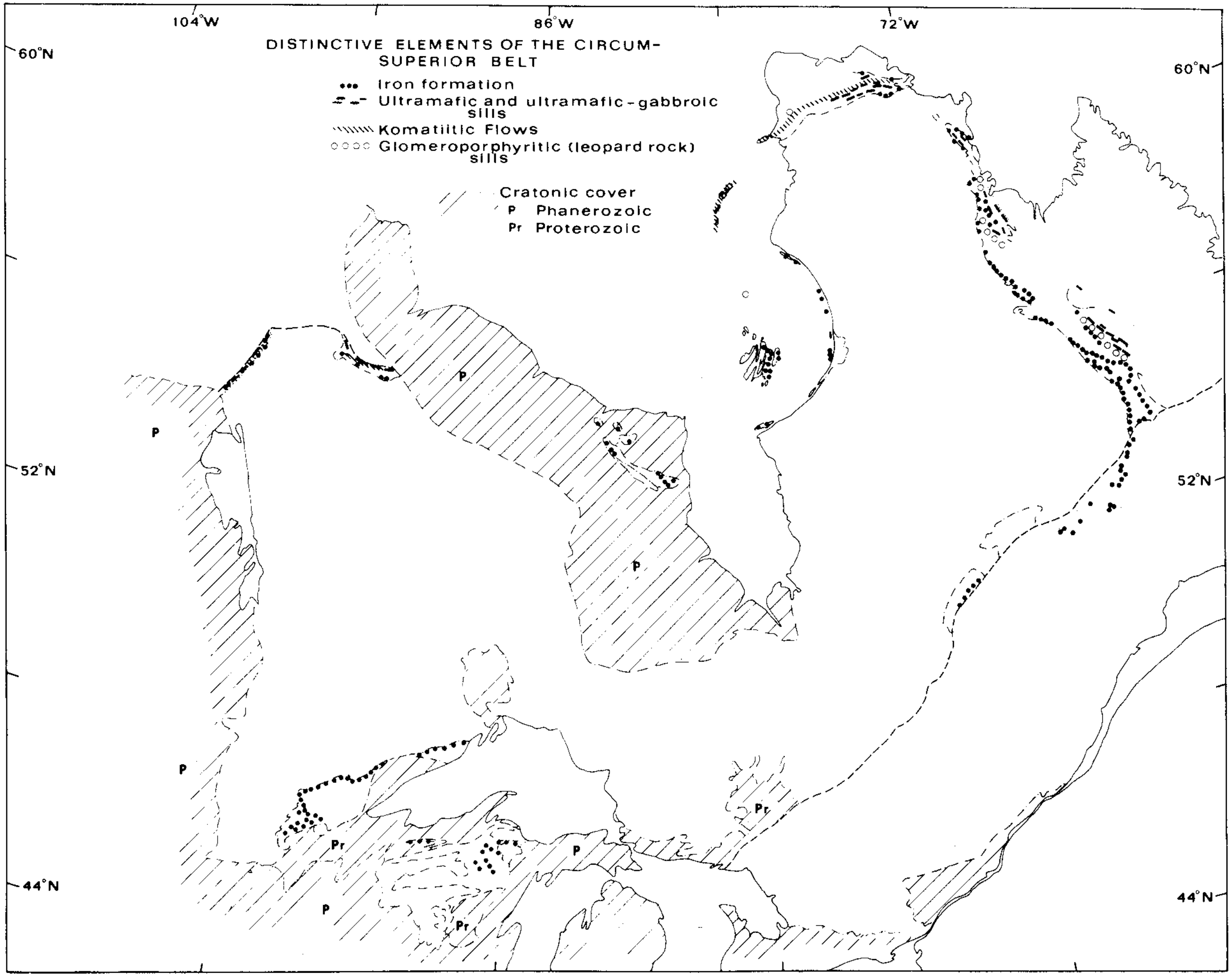

Fig. 12-5. Map showing the distribution of certain distinctive lithologies common to some or all segments of the Circum-Superior Belt.

the iron formation is thin, a few hundred metres thick, but in the Belcher Islands, southern part of the Labrador Trough, and Mistassini—Otish segments it is thicker and underlain by continental-type sediments accompanied in the first two instances by potash-rich basalts. This could also be the case elsewhere in the Circum-Superior Belt where early formed deposits may be hidden beneath the more extensive accumulations that follow. The continental beds appear to be related to rifting and it is possible that this is the essential clue to the origin of the whole belt.

The relationship of the supracrustal deposits of the Circum-Superior Belt to the Archaean basement is another mark of similarity among various segments of the belt. In the Labrador Trough, Cape Smith Belt, Belcher Basin and Lake Superior Association the basement clearly underlies most or all of the recognizable supracrustal rocks including the volcanic succession. In each of these segments the miogeoclinal deposits are on the inner side of the belt adjoining the Superior Province and the eugeosynclinal deposits are on its outer side.

Magmatic relationships

Magmatism in the northern part of the Circum-Superior Belt from the Labrador Trough to the Thompson Belt has much in common (Fig. 12-5). The Cape Smith Belt (Figs. 12-3 and 12-4) provides the best exposed and most complete example. Tholeiitic basalts are succeeded by a thick sequence of komatiitic basalts and peridotites. Sills of both types intrude sedimentary members of the stratigraphic column, particularly those interbedded with or immediately underlying the volcanic successions. The komatiities, because of their rarity in volcanic sequences, are a compelling element in the proposed correlation between segments of the northern part of the belt and, therefore, warrant some additional discussion.

In Fig. 12-6 the distributions of analyses of magmatic rocks in the various segments of the Circum-Superior Belt are displayed on a Jensen plot (Jensen, 1976), designed to show relationships in the whole domain of subalkalic volcanic rocks. Three of the segments as well as the Molson Dykes show trends that essentially coincide and range continuously from ultramafic komatiites to tholeiitic basalts. In the case of the Cape Smith Belt Francis and Hynes (1979) recognized in this continuity evidence supporting the view that the tholeiites and komatiites of that belt are co-genetic. The spread of analyses for the Molson Dykes of northern Manitoba is almost as extensive and is further evidence of the probable consanguinuity of the whole komatiite-tholeiite suite.

No komatiites have been identified among volcanic rocks of the Labrador Trough as shown in Fig. 12-6, but the ultramafic sills are remarkably similar to sills in the Cape Smith Belt that were recognized as komatiitic by Francis and Hynes (1979). They typically grade from a peridotitic zone in the lower

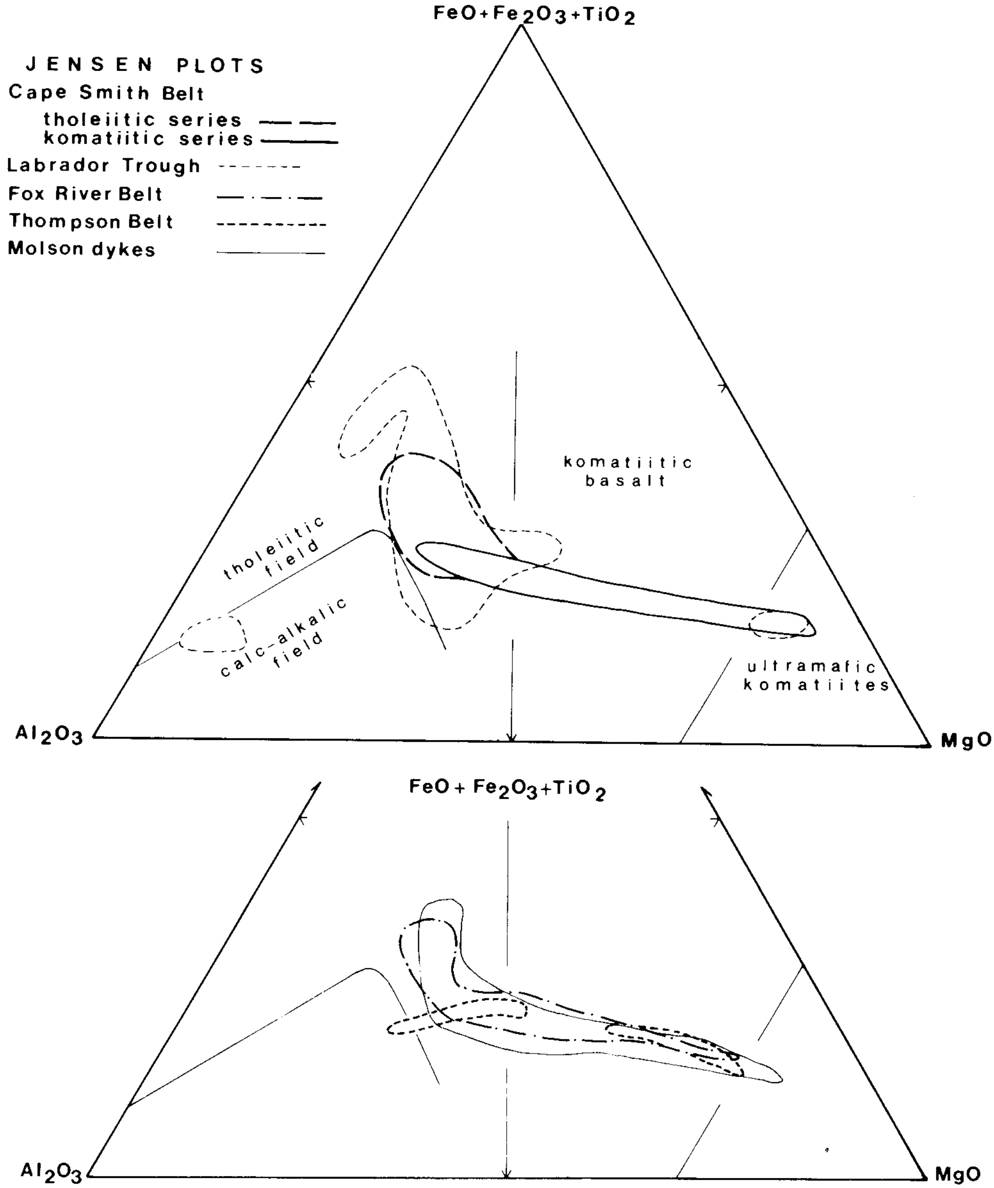

Fig. 12-6. Jensen cation plots of the fields of analyses from a number of the Circum-Superior segments as well as from the Molson Dykes. Sources of published data are noted in the text; unpublished data are from Scoates (Fox Lake) and Baragar (Cape Smith).

part of the sill to a gabbroic zone in the upper part (Fahrig, 1962; Baragar, 1967). A lower contact observed in one of the sills has the mineralogy of a very mafic basalt (Baragar, 1967) which, in the light of Francis and Hynes analyses of equivalent contact zones in a Cape Smith sill, is probably of komatiitic composition. Unfortunately no analyses are available of the chilled borders of the Labrador sills but analyses of the ultramafic parts of two such sills given by Fahrig (1962) and Dimroth (1971), respectively, are similar to the olivine cumulate phase of the Cape Smith sill studied by Francis and Hynes (1979). The late emplacement of ultramafic sills in the Labrador Trough also corresponds to the position of komatiitic magmatism in the upper part of the Cape Smith sequence. Komatiitic volcanics, therefore, may well have been erupted in the Labrador Trough, but at stratigraphic levels above those presently preserved.

Intermediate and felsic volcanic rocks and, with the exception of the Thompson Belt, their intrusive equivalents are rare in the northern part of the Circum-Superior Belt. Such felsic volcanics as have been identified in the Labrador Trough and Cape Smith Belt are dacites and rhyolites. Andesites are virtually unknown. The paucity and restricted composition of these felsitic rocks is more in keeping with an origin attributable to partial melting of an underlying sialic crust than to derivation from a subduction zone.

Magnetism in the Lake Superior region shows some differences to that in the northern part of the Circum-Superior Belt. Komatiites have not been identified but calc-alkaline rocks, both intrusive and extrusive, are evidently a significant part of the assemblage.

TECTONIC CONSIDERATIONS

Correlation of the various segments of the Circum-Superior Belt detailed in the earlier sections of this paper implies the existence of a continuous depositional belt around at least three quarters of the Superior Province. Closure of the belt through the remaining interval obscured by the Grenvillian Orogeny is a plausible assumption in view of its continuity outside the Grenville Province. The picture that emerges from such a construction is that of a craton surrounded by marginal deposits. Early in its history shelf sediments lapped on to the cratonic margin. Later they were succeeded without apparent break by predominantly mafic and ultramafic magmas. Plate-tectonic models have been proposed for various parts of the belt but no overall synthesis has yet been attempted.

Gibb and Walcott (1971) proposed that segments of the Circum-Superior Belt from the Labrador Trough to the Thompson Nickel Belt were parts of a continuous suture formed by the closure of an Aphebian ocean separating the Superior and Slave Province cratons. This proposal was later refined on the basis of gravity profiles taken across the Labrador Trough (Kearey,

1976) and Cape Smith Belt (Thomas and Gibb, 1977). The resulting crustal profile calculated from the gravity data shows thickened crust beneath the "suture" and an outward-dipping join between provinces of slightly different density; the younger (Churchill) being the greater. The similarity with Dewey's (1976) conceptual model of continental collision as noted by the authors (Thomas and Gibb, 1977) is striking. Burke and Dewey (1973) and Dewey and Burke (1973) elaborated on the Gibb and Walcott suture hypothesis by noting possible triple junctions in the Cape Smith and Labrador region and by interpreting the geology in terms of collision tectonics. For the southern side of the Superior Province Van Schmus (1976) suggested that the distribution and types of lithologies in the Lake Superior Association could be explained by a northward-dipping subduction zone. According to this interpretation the mafic and calc-alkaline volcanic rocks of the Lake Superior Association would have erupted onto the margin of the sialic crust and interfingered northward into mio- and eugeosynclinal rocks occupying an ensialic basin between the volcanic arc and craton proper. The later identification of early Archaean gneisses in the region of the volcanic "arc" (Van Schmus and Anderson, 1977) necessitated modifying the theory to relocate the subduction zone farther to the south or abandoning it.

In this paper we have attempted to summarize the major characteristics of the Circum-Superior Belt. The most important of these which must be satisfied by any tectonic model proposed for its origin are as follows:

(1) Its constituent deposits change in character from mio- to eugeosynclinal-type outward from the Superior Province and upward in the stratigraphic section. The earliest deposits in some and possibly all segments are rift-related sediments and volcanics.

(2) No major break is consistently present between the mio- and eugeosynclinal parts of the stratigraphic section. The change commonly occurs immediately, or almost immediately, after deposition of the major iron formation.

(3) The volcanic component comprises low-K tholeiites of oceanic type and, in the northern part of the belt, komatiites. These have not been emplaced by obduction as proposed in the models of Gibb and Walcott (1971) and Dewey and Burke (1973). Sills of related compositions intruding the mio- and eugeosynclinal sediments below the volcanics link them securely to the assemblage in which they are found.

(4) The calc-alkaline suite is rare or absent in the northern segments of the Circum-Superior Belt. It could be occult in the metamorphic hinterland but on the basis of work done to date it is unlikely to be abundant. However, calc-alkaline rocks are a significant component of the Lake Superior Association.

(5) Both mio- and eugeosynclinal assemblages are ensialic for as far out from the Superior Province as primary structures can be readily recognized.

Palaeomagnetic restraints must also be taken into account. According to

Irving and McGlynn (1976), Interior Laurentia, including the Superior, Slave and Nain Provinces, existed as a single entity during the period from 2700 to 1300 Ma B.P. This restricts the possible separation of continental masses on either side of the Circum-Superior Belt to about 1000 km, the uncertainty in the position of the Apparent Polar Wander path (APW) at that time. Christie et al. (1975) determined that the Kaminak dykes of the Churchill Province and the approximately contemporaneous and roughly parallel Matachewan dykes of the Superior Province have the same palaeomagnetic pole positions within their error of measurement. Again this precludes a major opening of the continent along the Circum-Superior Belt. Finally, Schmidt (1980) has recently shown that the pole position for the upper volcanic formation of the Belcher Islands is entirely consistent with the APW path of North America. This greatly reduces its likelihood of being transported and obducted oceanic crust.

The conditions imposed by a consideration of these points would seem to favour a model such as follows (Fig. 12-7).

Subsidence was initiated in a sialic crust by continental rifting, localized, perhaps, at the margins of a thickened and stable crustal block, the present Superior Province. Continuing separation along the loci of rifting resulted in attenuation of the crust, subsidence of its surface, and inundation by the sea which would then surround the Superior block[1]. Miogeoclinal deposits formed on the margins of the "continent". With further separation the crust eventually ruptured, the mantle became involved and an incipient oceanic rift was initiated. The resulting magmas invaded and overrode the shelf sediments previously deposited. Both the contribution of iron from volcanic sources and the restriction in circulation caused by the volcanic build-up promoted the conditions favourable for deposition of iron but later overwhelmed the deposition sites with volcanic materials. Subsequent spreading appears to have been slow and volcanic deposits accumulated to considerable thicknesses. Spreading along the northern ridge had not proceeded far when the total spreading movement was assumed by a rapidly opening southern rift. Activity in the north ceased and the trough closed with the resulting deformation of its contents that we presently see. Associated with the closure an incipient subduction zone, possibly initiated in the northern trough but transferring southward with increasing resistance there, developed on the southern flank of the Superior continent and gave rise to the calc-alkaline assemblage of the Lake Superior Association as envisaged by Van Schmus (1976).

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[1] Crustal thinning and subsidence accompanying rifting is a complex problem which we do not wish to minimize. See, for example, discussions by Bott (1979) and Keen and Hyndman (1979). However, a ductile spreading model analogous to that proposed for the Bay of Biscay continental margin by de Charpal et al. (1978) may be a suitable mechanism.

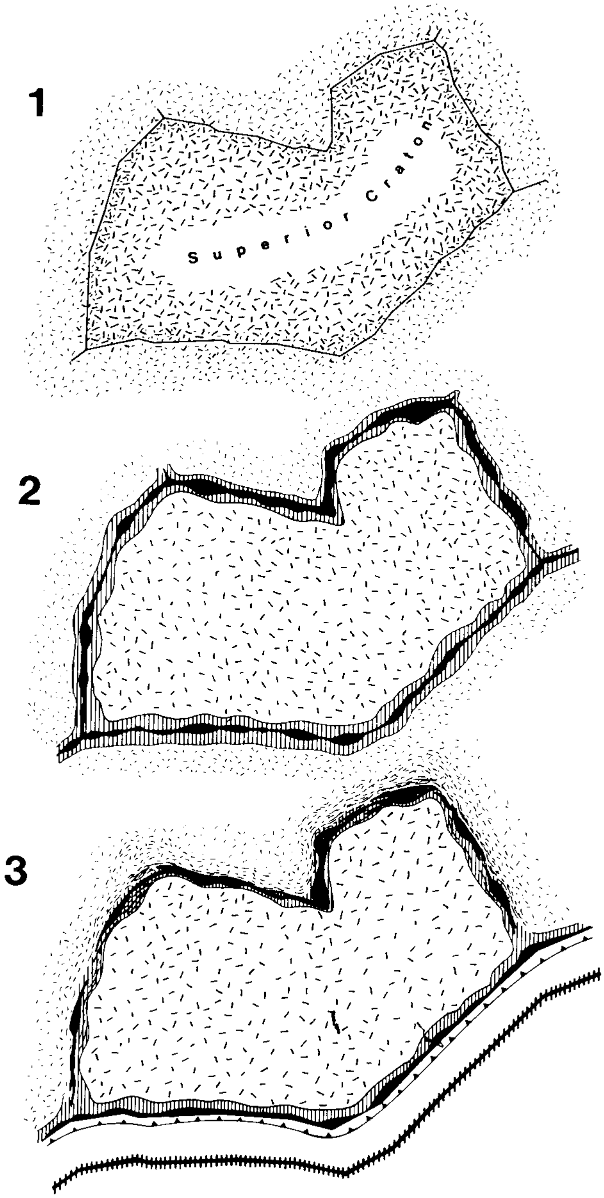

Fig. 12-7. Successive stages in the hypothetical development of the Circum-Superior Belt. 1. Stretching of the sialic crust develops fractures around a previously thickened node, the Superior craton. 2. Continued stretching results in necking of the crust along the fractures with attendant formation of an annular trough peripheral to the craton. Sediments (hatched) accumulate in the deepening trough and eventually, as the loci of spreading penetrate deeper into mantle, volcanism (black) is initiated. 3. Spreading is assumed by the southern rift which opens into an ocean (rift shown as hatched line); activity ceases in the northern rift zone and it closes. Subduction, possibly initiated in the northern rift zone, develops along the southern rim of the craton (toothed line shows direction of dip).

The closest modern analogy to the tectonic scheme outlined above may be the continent of greenland semi-surrounded as it is by active and inactive oceanic ridges. The Labrador Sea—Davis Strait—Baffin Bay channel along the inactive ridge would correspond to the northern arm of the Circum-Superior Belt, although considerably wider than we would propose for it. If its width were a little more than the combined width of continental basement underlying both sides of Baffin Bay and the Labrador Sea (Keen et al., 1974; Grant, 1975) it would be about 300 to 500 km. This would be a reasonable width to expect given the constraints of palaeomagnetism and the proposal that spreading little exceeded the limits of thinning of the continental crust. Interestingly volcanic rocks associated with the Davis Strait spreading centre and exposed on both sides of the Strait (Clarke, 1970; Clarke and Upton, 1971), were initially very magnesian, similar in composition to the komatiites, but evolved upward into feldsparphyric basalts. Clarke (1970) considered them to be primary melts derived directly by partial melting of garnet peridotite mantle.

In the Circum-Superior Belt the presence of komatiites is an exotic but uncertain factor in regard to its tectonic development. Gorgona Island off the coast of Colombia has the only known modern examples (Gansser, 1950; Gansser et al., 1979; Echeverria, 1979) from which to judge the tectonic environment. According to Gansser et al. (1979) they are an uplifted part of the ocean floor from either the Pacific crust or a marginal basin. In his earlier work Gansser (1950) linked the Gorgona Island rocks to the Coastal Cordillera of Colombia which includes algal limestones and siliceous shales. It is not clear, therefore, whether the komatiites were emplaced in an oceanic or a continental shelf environment (possibly related to the activity of a continental arc). Until more information on the tectonic environment of emplacement is available from the Gorgona Island occurrence the environmental significance of komatiites must remain somewhat uncertain.

CONCLUSIONS

Late Aphebian supracrustal sequences exposed in nine segments distributed unevenly around the Superior Province are parts of a more or less continuous belt of rocks which originally encircled a Superior craton. The loci of deposition was determined by rifting which spread around a previously thickened part of the sialic crust, the Superior "node", in response to crustal stretching. The first deposits are related to rifts but are followed by miogeoclinal-type deposits as the rift zone evolves into a shallow annular trough with further stretching and "necking" of the sialic crust. Volcanism was initiated when the loci of separation along the axis of the trough extended to sufficient depths in the mantle to tap a magmatic source. Volcanic flows and related sediments spread over the previously formed miogeoclinal deposits. Iron formation resulting from the restricted circulation and hydrothermal activity attendant on volcanism marks the beginning of a transition to eugeosynclinal conditions. Komatiitic flows followed an early succession of tholeiitic basalts in the northern half of the belt but their tectonic significance is as yet unknown. With continued spreading of oceanic rift developed along the southern side of the craton and assumed all of the divergent motion which was previously dispersed between the two halves of the annular belt. The northern part of the belt ceased spreading and eventually closed with severe deformation of its contents.

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Chapter 13

LABRADOR GEOSYNCLINE: TYPE EXAMPLE OF EARLY PROTEROZOIC CRATONIC REACTIVATION

ERICH DIMROTH

ABSTRACT

The Labrador Trough is the erosional remnant of an early Proterozoic geosyncline that formed by reactivation of an Archaean craton. Evolution of the geosyncline took place in seven phases: (1) development of a red-bed graben in the centre and south of the geosyncline; (2) slow subsidence and deposition of shelf sediments; (3) tectonic instability, faulting, erosional discordances, local basins of mass-flow conglomerate; (4) massive rapid subsidence, voluminous basalt volcanism, deposition of shale and greywacke; (5) restabilization, deposition of the shelf sequence of cycle II; (6) rapid subsidence, voluminous basalt volcanism and renewed deposition of shale and greywacke; (7) folding and metamorphism, subduction of the geosyncline below its hinterland. About 100—200 km of Archaean basement disappeared, probably by underplating, below the hinterland to the east of the trough.

The author speculates that plate-tectonic processes, or processes closely related to plate tectonics, probably took place since the Archaean. Plate—margin geosynclines and island arc—forearc basin systems formed, perhaps, where the underflow of descending mantle convection currents acted upon oceanic crust and upon the ocean—continent interface. Ensialic belts like the Labrador Trough formed, perhaps, where the underflow of such convection cells acted upon continental crust, within a large sialic block.

INTRODUCTION

The Labrador Trough deserves to be presented as a type example of an early Proterozoic fold belt that formed by reactivation of an Archaean craton: on the one hand, it shows some surprising similarities to Alpine-type mountain chains (which formed by plate-tectonic processes) in its stratigraphic sequence, in its magmatism, and in its structural and metamorphic patterns; on the other hand, it demonstrably formed on top of an Archaean craton, and demonstrably was underlain by sialic crust during its whole history. Furthermore, it can be demonstrated that lateral extension of the crust during its geosynclinal evolution was relatively minor. These striking similarities and dissimilarities of an early Proterozoic fold belt and Alpine mountain chains should provide food for speculation.

In this paper, the term Labrador Trough is used for a belt of folded early Proterozoic rocks in northeastern Quebec and Labrador, presently exposed.

The term Labrador geosyncline denotes the sedimentary basin within which the Labrador Trough rocks were deposited. The Labrador geosyncline is only one branch of a very large geosynclinal system, bounding the Archaean Superior Province in the west, north and east and called Circum-Ungava geosyncline (*see also Baragar and Scoates, this volume, Chapter 12, ed.*). Possibly, this geosyncline was even continuous with the Penokean fold belt in Wisconsin and Minnesota, across the Grenville Province (*see also Van Schmus and Bickford, this volume, Chapter 11, ed.*).

The geology of the Labrador Trough is quite well known. Aided by exceptionally good outcrop, and stimulated by the presence of huge ore deposits, most of the trough has been mapped in great detail. Most stratigraphic correlations along and across the trough are very well established, particularly in the zone south of 57°30′N (Dimroth, 1978; Dressler, 1979; Clark, 1977). More sophisticated laboratory research on the sedimentology, igneous and metamorphic petrology and structure of the geosyncline should be done, but the basic geological freatures have been well established.

The Labrador Trough (Fig. 13-1) is a linear belt of early Proterozoic sedimentary and volcanic rocks in northeastern Quebec and Labrador, about 800 km long and about 100 km wide. In the west, it is bounded by a foreland underlain by Archaean gneisses and granites which have not been affected by the Hudsonian Orogeny. In the east, it is bounded by a terrain of highly metamorphosed "granitic" gneisses containing some belts of metasedimentary rocks (paragneiss, metaquartzite, marble, etc.). Most of the "granitic" gneisses of this terrain are also of Archaean age but were subjected to a second high-grade metamorphism during the Hudsonian Orogeny. The paragneisses, essentially, are of Lower Proterozoic age and a few small stocks and batholiths of Hudsonian granite and granoriorite are also present.

In the south, the Labrador Trough is cut by the Grenville front, a zone of increasing metamorphism and deformation imposed by the Grenville Orogeny, and small remnants of the Labrador Trough sequence can be followed deep into the Grenville Province (see Baer, this volume, Chapter 14). In the north, the trough terminates as a southerly plunging synclinorium, and the Archaean granitoid gneisses are continuous around the nose of this synclinorium. Effects of the Hudsonian Orogeny can be seen in the Archaean gneisses around this synclinal nose; they are very weak in the west and increase rapidly eastward until the degree of Hudsonian metamorphism and deformation becomes extreme.

The filling of the Labrador geosyncline consists of arkose, conglomerate, orthoquartzite, dolomite, iron formation, shale, greywacke and of very large volumes of basalt and gabbro. It has been folded and metamorphosed during the Hudsonian Orogeny about 1.8—1.6 Ga ago. The intensity of the Hudsonian metamorphism and deformation increases from west to east, and the upper amphibolite facies is attained at the eastern limit of the trough.

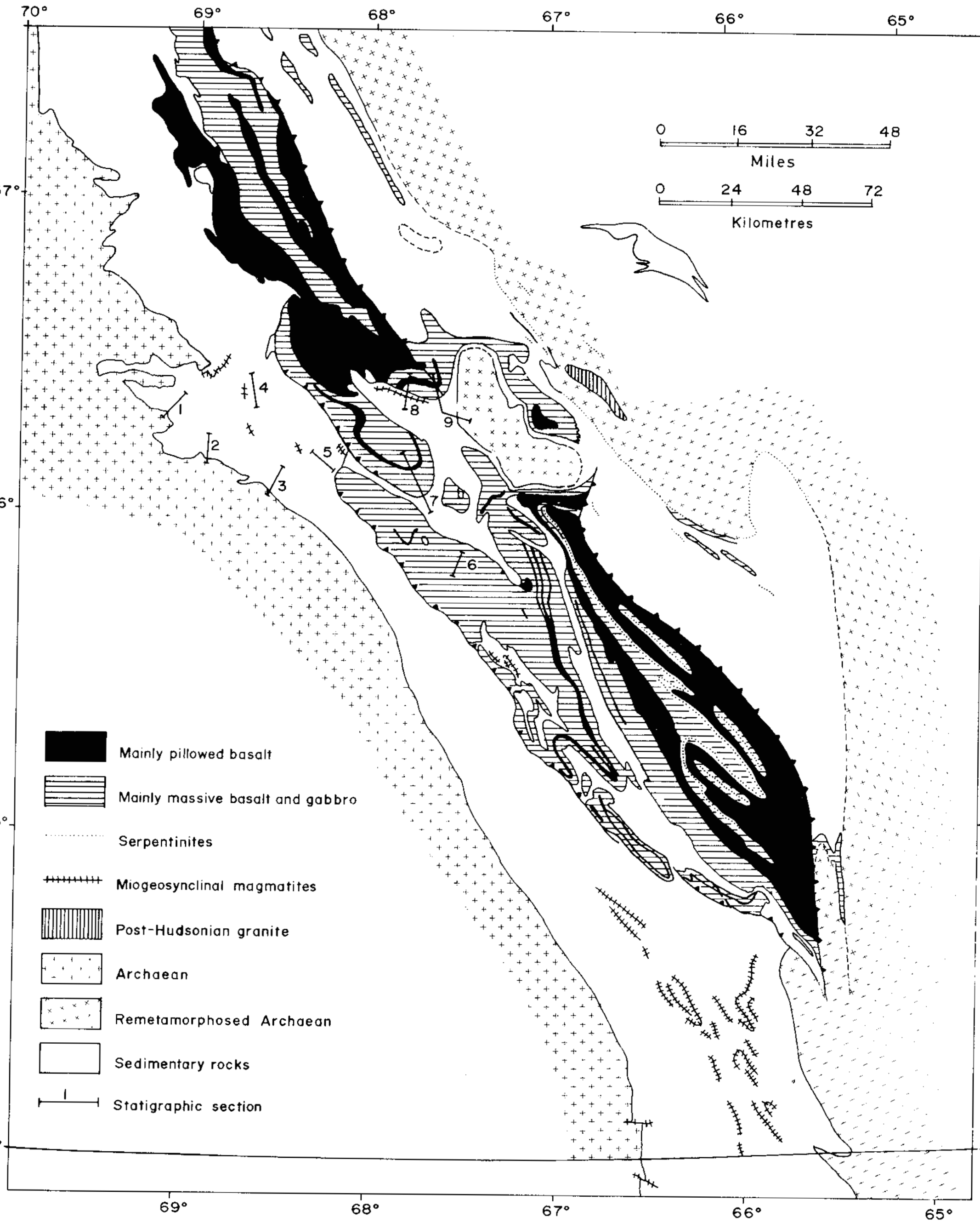

Fig. 13-1. Generalized geological map of the southern and central Labrador Trough.

The sense of overfolding and of thrusting is to the west. The metamorphic facies series indicates intermediate pressure.

The age of the Labrador geosyncline is not well known. The geosynclinal filling certainly is younger than about 2.7 Ga — the age of its Archaean basement — and older than 1.6—1.8 Ga — the K-Ar age of the Labrador Trough rocks. A Rb-Sr isochron of low-grade shales (Fryer, 1972) and a K-Ar determination on biotite of a volcanic rock (Dressler, 1975) gave ages of about 1.85 Ga. This has been interpreted as an age of deposition (Fryer, 1972; Dressler, 1975) but could in fact represent the age of the (or of an) orogeny or could represent a mixed age. Schiman (1978) obtained a Rb-Sr isochron of 2.3 Ga from volcanic rocks of the easternmost Cape-Smith belt and, consequently, suggested that this belt is older than the Labrador geosyncline. On the other hand, the stratigraphic sequence in the northernmost Labrador Trough and easternmost Cape-Smith belt is so similar that a formational correlation appears evident to me. In summary, the Labrador Trough is of early Proterozoic age but more detailed work is required to date it with greater precision.

This paper is an expanded abstract of previous more detailed descriptions (Dimroth, 1970, 1972; Dimroth et al., 1970; Dimroth and Dressler, 1978) and is based on a compilation of many sources of information. The reader is referred to my previous papers for detailed references and documentation.

THE GEOSYNCLINAL FILLING

Stratigraphic correlations across the Labrador Trough at about 56° N are shown in Fig. 13-2. Two sedimentary cycles are present, each beginning with deposition of stable shelf deposits (orthoquartzite, dolomite, iron formation) and culminating in deposition of shale, greywacke and in extrusion of very voluminous basalt. Few of the rocks of the upper cycle have been preserved at 56° N (Fig. 13-2) but the cycle is fully developed farther north and south.

Very thick red beds, the arkose and arkosic gneiss pebble conglomerates of the Chakonipau Formation, locally underlie the rocks of cycle I. Orthoquartzite, dolomite and pelite of the Portage, Lace, Alder and Uvé Formations and their stratigraphic equivalents follow. The upper two of these formations are absent over wide areas in the central Labrador Trough (Fig. 13-2, sections 5 and 6); however, their former continuity across the whole width of the geosyncline is well demonstrated by the detailed correlation of the stratigraphic sequence at Otelnuk and Romanet Lakes (sections 4 and 7, Fig. 13-2), documented in table 7 of Dimroth (1978).

Thus, parts of the shelf-type deposits were eroded in the central Labrador Trough and were redeposited in local basins as mass-flow conglomerates composed of orthoquartzite and dolomite pebbles (Romanet Formation).

Fig. 13-2. Stratigraphic correlations across the Labrador Trough at 56°15′N. For localities of sections see Fig. 13-1.

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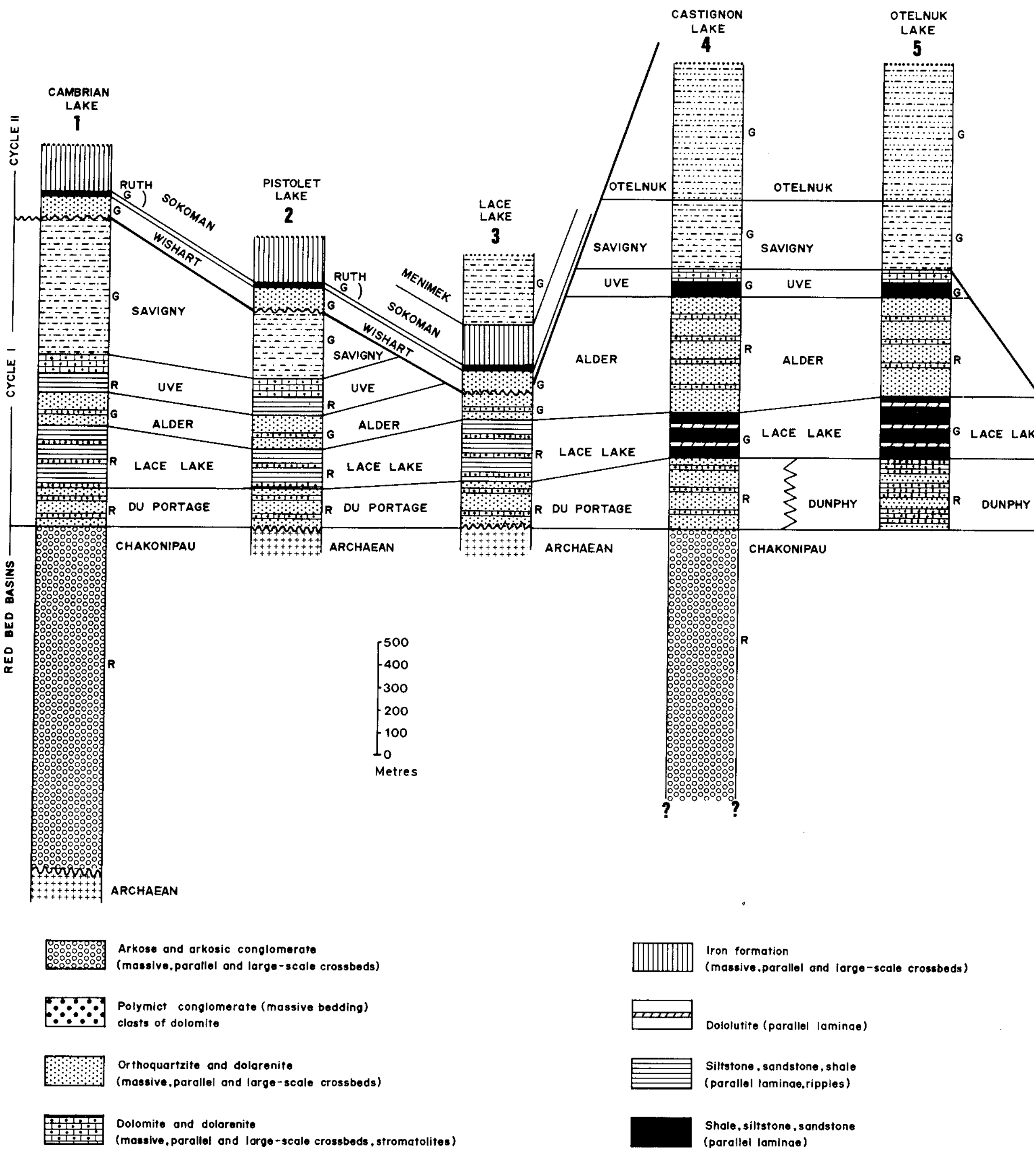
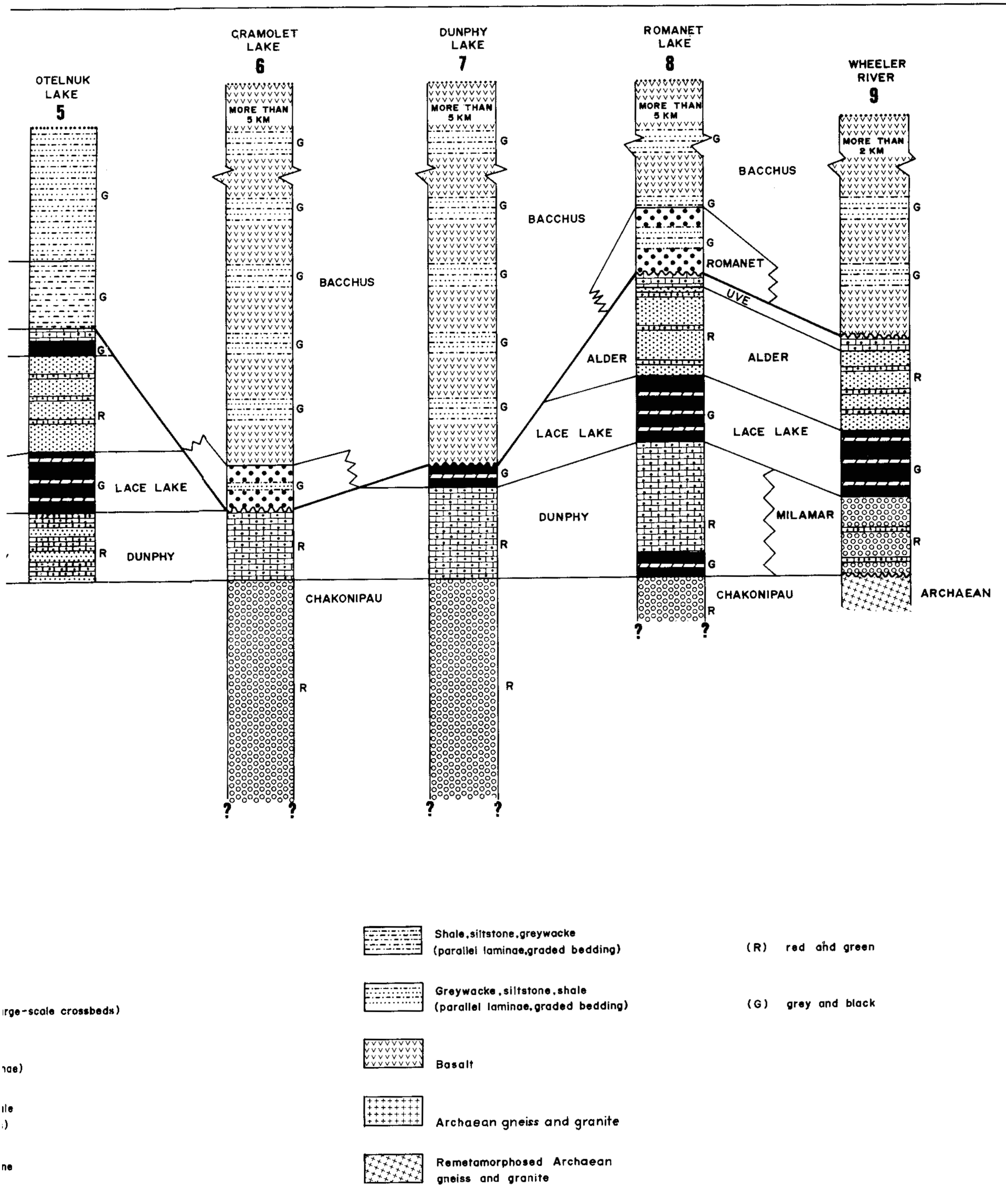


. 13-2. Stratigraphic correlations across the Labrador Trough at 56°15′ N. For localities of sections see Fig. 13-1.



:tions see Fig. 13-1.

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The very peculiar greywackes of the Savigny and Otelnuk Formations also are derived from shelf-type deposits. These local unconformities and local deposition of mass-flow conglomerates document a period of tectonic instability preceding the eruption of the basalts of the Bacchus Formation.

Pronounced facies changes permit to define three zones in the geosyncline. In the west, shelf deposits at the base of both cycles make up about 30—50% of the sequence and not more than a very small volume of volcanic rock is present. The total thickness of cycle I at 56°15′N is about 1—3 km. This zone can be considered a miogeosyncline.

The central and eastern zones can be considered to be the eugeosyncline because they contain voluminous basaltic volcanic rocks. Thickness and facies of the shelf-type deposits at the base of cycle I change little except in the extreme east of the Labrador Trough, where thick formations of dolomite and orthoquartzite grade eastward into semipelites (now biotite-plagioclase paragneiss) with interlayers of metaquartzite and marble. The total thickness of cycle I is about 10 km in the center of the geosyncline; thickness changes of the sedimentary sequence to the eastern part of the trough are unknown because of structural complications; however, the thickness of the basalts of cycle I again decreases from the center to the east of the geosyncline.

Sedimentary environments and provenance

We have sufficient information on the sedimentary structures and textures of the sediments in the west and center of the trough to draw general conclusions as to their environment of deposition. Sedimentary structures and parts of the textures have been destroyed by intense deformation and metamorphism in the east of the trough.

The red beds of the Chakonipau Formation are fluvial (piedmont fan, braided stream) fanglomerates. The orthoquartzites and dolomites at the base of cycle I are littoral and shallow marine deposits and intercalated pelites may in part be deltaic. The shales and greywackes of the Savigny, Otelnuk and Romanet Formations contain turbidite beds (see Dimroth, 1978, Fig. 11A and 11C for Bouma-cycled turbidites) and, to a degree, resemble flysch-type deposits. However, I do not believe that these are true deep-sea fan deposits. Rather, I feel that this sequence is comparable to the Jurassic "flysch" of Western Canada (Hamblin and Walker, 1979), deposited in a rather deep shelf sea below storm wave base. The conglomerates of the Romanet Formation mostly are massive-bedded debris flows.

The source of the sediments can be deduced from their composition: the red beds underlying cycle I are derived from Archaean granitoid gneisses and from minor synsedimentary andesitic volcanic rocks. The detrital sediments at the base of cycles I and II are derived from deeply weathered granitoid gneisses. Greywackes and conglomerates of the upper part of both cycles are derived from unmetamorphosed sedimentary rocks: they contain well-rounded grains of quartz and feldspar, together with splinters of siltstone, sandstone, dolomite, dolarenite etc. whereas components derived from volcanic rocks and from low-grade metamorphic rocks are absent. In particular, the greywackes are totally different compositionally from synorogenic (Pacific-type) flysch.

Provenance of sediments has been recognized by changes of grain size and facies. We find that the larger part of the sediments is derived from the foreland to the west of the geosyncline. However, local source areas within and to the east of the geosyncline were present at least during certain stages of geosynclinal evolution.

Volcanic rocks

Volcanic rocks are of three types:

(1) Low-K tholeiites are the predominant component and form much of the sequence in the centre of the geosyncline.

(2) Small volumes of alkali-basalt are locally present.

(3) Carbonatites and associated lamprophyric rocks (alnoites, mellilite-basalt, "meimechite") are also present.

The latter rocks were originally thought to be post-tectonic (Dimroth, 1971), but Dressler (1975) found mellilite-basalt tuffs intercalated with the Sokoman Iron Formation. Thus, they very likely are also part of the geosynclinal filling.

The feeders of the first two types of volcanic rocks are unknown. Thick gabbro sills intruded the lower part of the sequence in the center of the geosyncline. These are low-K tholeiites and clearly are equivalent to the basalts higher up in the sequence. Diabase dyke systems older and younger than the Labrador of geosyncline have also been mapped: a large dyke system extends in the Archaean terrain between the Cape Smith belt and the Labrador Trough but this system is overlain unconformably by the basal sediments of both fold belts; some younger diabase dykes are also present in the south-central part of the Labrador Trough, but these dykes cut across folds and are unmetamorphosed. It is noteworthy that not a single dyke which can be shown to be clearly co-genetic with the volcanic sequence, has been mapped in the whole of the Labrador trough, despite the excellent exposure.

Ultramafic rocks occur in three associations:

(1) Differentiated sills, ultramafic at the base, gabbroic at the top occur locally in the north and south of the Labrador Trough. These clearly are intrusive bodies, cogenetic with part of the volcanic sequence.

(2) Thin lenses of ultramafic rocks have been mapped by Baragar (1967) in the volcanic sequence of cycle II; they have been interpreted as intrusive rocks but might be komatiites (which were still unknown as a rock type at the time of mapping).

(3) Thin lenses of ultramafic rocks occur in the metamorphosed sediments in the easternmost Labrador Trough. The origin of these rocks has been obscured by strong metamorphism and deformation.

STRUCTURE AND METAMORPHISM

As in other mountain belts, the intensity of deformation, the style of the structures, and the degree of metamorphism change systematically across the Labrador Trough (Fig. 13-3). However, structural relations are complex because the orogenic structures have been molded by synsedimentary block faults. Thus, I will describe the normal change of structural style across the fold belt and will then discuss how the folds have been molded around synsedimentary fault blocks.

The orogenic front is defined, at high crustal level, by an array of imbricate thrust faults steeply dipping to the east. Downward these listric thrust faults flatten and merge into one continuous decollement. The remainder of the western (miogeosynclinal) zone has been folded into a closely spaced system of nearly isoclinal, doubly plunging, synclines and anticlines overturned to the west. One single fold direction, in general to the NNW, predominates and in general a single schistosity is present. The mostly volcanic sequence of the central part of the trough has locally been thrust upon the miogeosynclinal sequence. However, thrust faults are not continuous over the whole length of the geosyncline and normal contacts of the sequences in the western and central Labrador Trough exist, for example, north of 57° N and south of 55° N.

Folds in the central part of the trough generally show a pattern of domes and basins whose long axes trend NNW. These structures are produced by the interference of a predominating set of folds trending NNW and overturned to the west, with subordinate sets of folds trending E and NE. One, two or three cleavages are present but do not show clear age relationships. The various fold sets in this tectonic zone apparently formed during one phase of deformation and for this reason a clear pattern of superposition of folds and of schistosities was not recognized during detailed work.

Several (up to 5) generations of superposed folds, each with its own axial plane schistosity, are present in the easternmost Labrador Trough. One generation of recumbent folds is present and appears to become more prominent towards the eastern limit of the trough. Large-scale recumbent folds, refolded by several generations of later folds, have been mapped in the immediate hinterland of the Labrador Trough by Hynes (1978).

As noted above, the structural style is greatly influenced by synsedimentary block-faulting in the segment of the trough between 56° N and 57° N. Fig. 13-4 shows the pattern of normal faults that developed during cycle I of deposition and that has been reconstructed on account of abrupt changes of thickness and facies of formations. The normal faults brought into lateral contact blocks of very different mechanical properties; consequently, orogenic stresses were refracted in systems parallel to the block boundaries as documented clearly from Fig. 13-4. Folds are thus molded against the boundaries of the fault blocks.

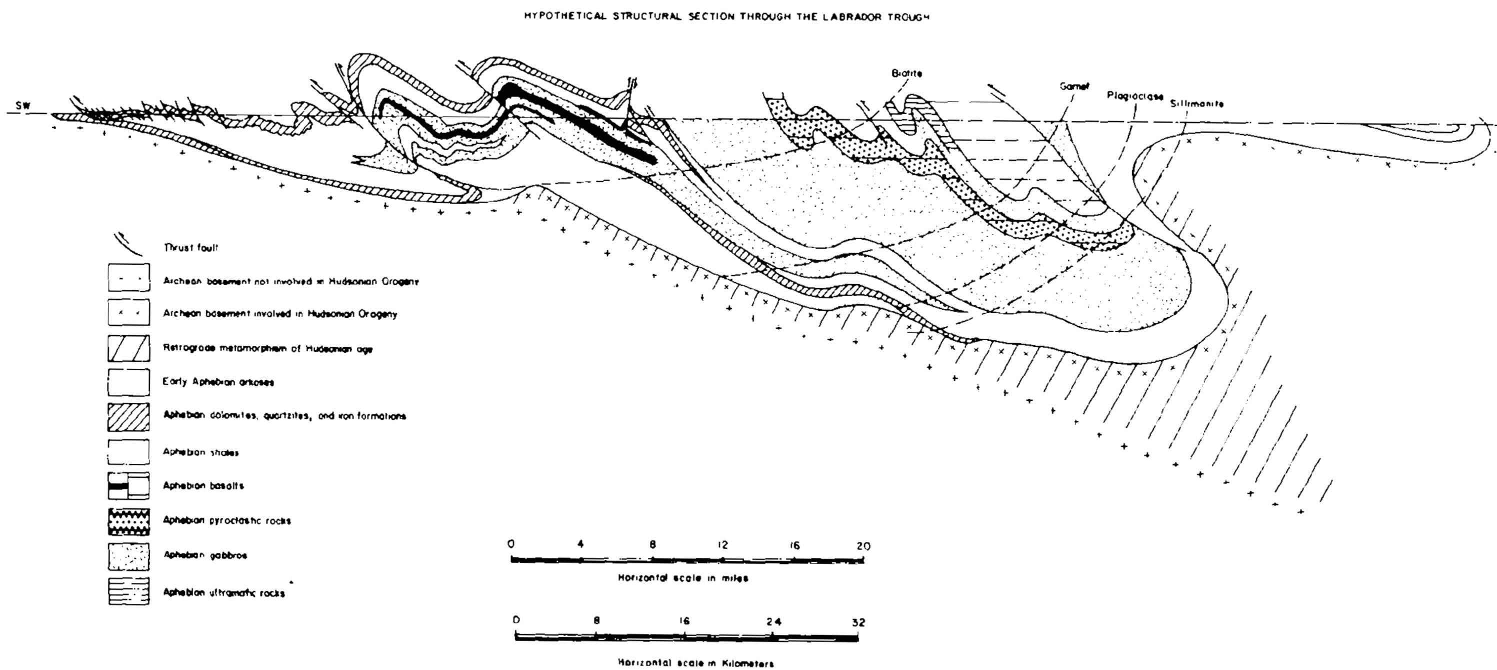


Fig. 13-3. Generalized geological section across the Labrador Trough. From Dimroth and Dressler, 1978.

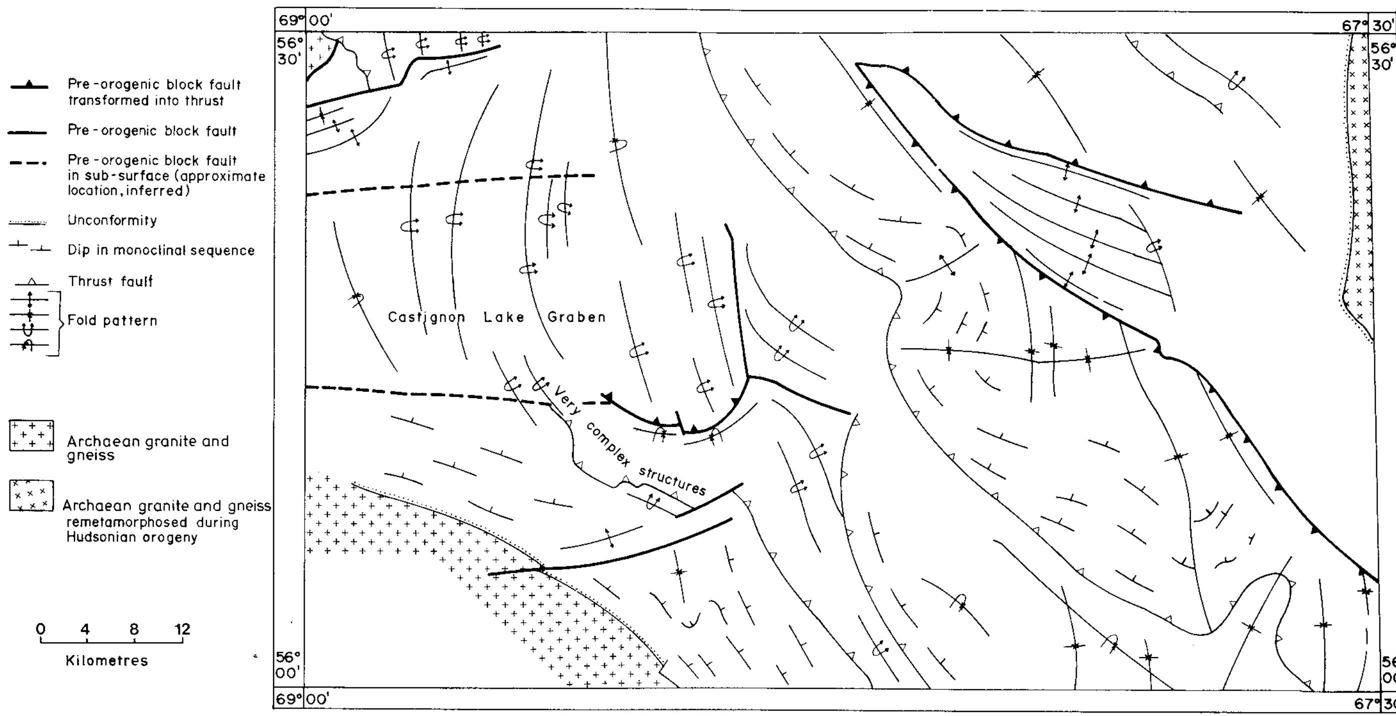


Fig. 13-4. Fold patterns in the central Labrador Trough and their relation to synsedimentary block faults.

The Archaean basement of the western zone of the trough in general has not been involved in the Hudsonian deformation, except locally in shear zones along the synsedimentary block faults. On the other hand, the basement has been involved in Hudsonian folding in the eastern and central zones and basement domes in the eastern zone as well as the basement gneisses of the geosynclinal hinterland were deformed together with their overlying cover.

Metamorphic grade increases from the pumpellyite-prehnite facies in the west to the amphibolite facies in the east. Mapping of isograds and the relation of metamorphic isograds to regional fold plunges suggest that the biotite isograd has rather shallow dips to the west, whereas the sillimanite isograd dips steeply (Fig. 13-3). Mineral associations suggest intermediate pressure metamorphism and a geothermal gradient of 20—40°C per km. Isograds cut across the Archaean gneiss domes and are unrelated to any type of heat domes. It is therefore believed that metamorphism resulted from the effect of deep burial of the easternmost trough and its hinterland by basement-cored nappes (Dimroth and Dressler, 1978).

Relations of mineral growth and tectonic structures suggest that the amphibolite-facies metamorphism at the eastern limit of the trough culminated at a late- to post-kinematic stage, whereas the greenschist facies metamorphism in the central Labrador Trough was synkinematic. These data and K-Ar age determinations (which presumably date the cooling of the rocks below a temperature of about 300°C) suggest migration of the orogeny across the geosyncline from east to west.

The geology of the hinterland east of the Labrador Trough is very poorly known. Taylor (1969, 1970) mapped vast areas at a scale of 1:500,000 and found most of the terrain to be underlain by a complex of "granitic" gneisses, migmatites and granulites. Narrow zones of paragneiss, marble and metaquartzite are present, and a few small stocks and batholiths of undeformed granite cut the whole sequence.

The metamorphosed sediments and the undeformed granites undoubtedly are of Proterozoic age. Taylor (1969, 1970) also suggested a Proterozoic age for the complex of "granitic" gneiss, migmatite and granulite. However, this interpretation is not consistent with the mapping by Wynne-Edwards (1960, 1961) and with the detailed cartography and petrology done by Gélinas (1965), the writer and Dressler (reported in Dimroth and Dressler, 1978). All these authors found that the "granitic" complex east of the trough forms the basement of the trough strata and has been metamorphosed for a second time during the Hudsonian Orogeny. In fact, the Hudsonian metamorphic isograds cut across gneiss domes. Thus, I consider the terrain east of the Labrador Trough to represent an Archaean gneiss complex that has been remobilized during the Hudsonian Orogeny.

Autochthonous position of basalts

The autochthonous position of the mafic extrusive sequence deserves to be stressed. To be sure, the Bacchus Formation and the units below it have been thrust over the miogeosynclinal sequence, perhaps for as much as 10—20 km. However, these thrust faults are not continuous over the whole length of the Labrador Trough. Furthermore, the Chakonipau, Dunphy and Lace Lake Formations are exposed in anticlinal zones within the central Labrador Trough. There, these formations have been intruded by an important system of gabbro sills which is cogenetic with, and indistinguishable from, the basalts of the overlying Bacchus Formation (Baragar, 1967).

The basalts of cycle II also are autochthonous. In Wakuach Lake area (Baragar, 1967), they form a monoclinal east-facing sequence, overthrust by the metasediments of the easternmost Labrador Trough. Southeast of Romanet Lake (Dimroth, 1978), the cycle II basalts have been mapped as a south-eastward plunging synclinorium, the eastern limb of which is faulted off. The metasediments of the easternmost trough have been followed, without interruption, into the less metamorphosed sediments to the west, around the nose of that synclinorium (Dimroth, 1978).

DISCUSSION

Stratigraphic relationships in the Labrador Trough permit unambiguous reconstruction of the processes that took place at the surface during the filling of the geosyncline. The evolution of the geosyncline was initiated by down-faulting of a graben in the south of the geosyncline and its filling with continental red beds several km thick (Fig. 13-5a). Deposition of the rocks of cycle I, also restricted to the southern half of the geosyncline, followed in three phases: First, shelf-type sediments were deposited in a slowly subsiding basin (Fig. 13-5b). Second, a period of tectonic instability and faulting followed with development of local discordances and local basins of mass-flow conglomerates (Fig. 13-5c). This period of tectonic instability initiated rapid subsidence of the central trough and intense basaltic volcanic activity (Fig. 13-5d).

The sequence of cycle II was deposited in the same way after restabilization of the geosyncline. The sediments and volcanic rocks of cycle II were deposited along the whole length of the geosyncline but it appears that the thickness of the volcanic sequence is greatest where the volcanic sequence of cycle I is absent (in the north of the Labrador Trough) or where it is fairly thin (in the south).

It appears to me important to point out the fundamental difference between the basalt sequence of the Labrador Trough and an ophiolite sequence. The basalts of the Labrador Trough form a horizontally stratified sequence overlying horizontally stratified sediments (Fig. 13-6a). Thus, the structure of the Labrador Trough basalts proves their origin by *vertical accretion*. True ophiolite sequences consist of laterally overlapping flow sequences overlying an intrusive complex (sheeted dykes, stocks) (Fig. 13-6b). The lateral overlap of flows (very well exposed on Iceland but not,

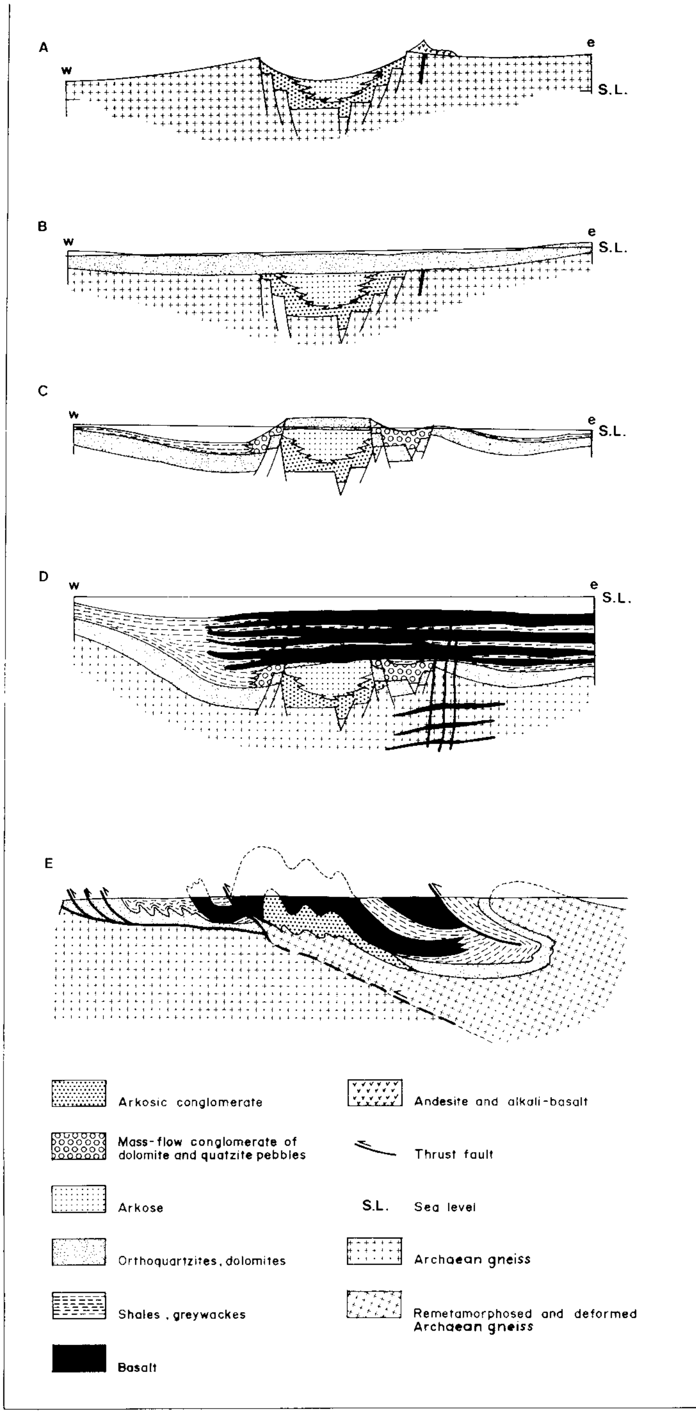


Fig. 13-5. Evolution of the Labrador geosyncline. A — Continental fault basins. B — Deposition of shallow marine shelf orthoquartzite and carbonate. C — Phase of instability, faulting, and deposition of conglomerate, shale and greywacke. D — Phase of basaltic volcanic activity. E — Final stage of orogeny.

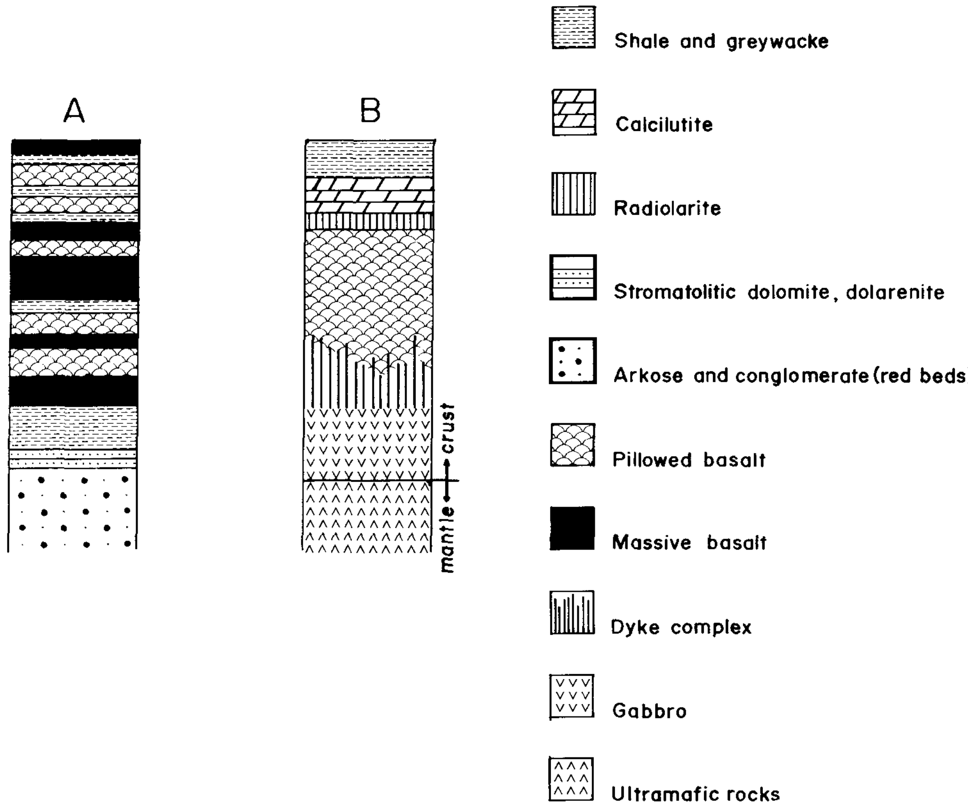


Fig. 13-6. Comparison of the stratigraphy of the volcanic sequence in the Labrador Trough (A) and ophiolite sequences (B).

in general, visible in small ophiolite complexes), and the relation of flows to the underlying intrusive rocks prove their origin by *lateral accretion.* The Labrador Trough basalts are similar in their structure to plateau basalts but formed below the sea; they are quite unlike any ophiolite sequence.

In summary, it is quite obvious from the stratigraphic record that the Labrador geosyncline formed essentially by vertical subsidence. Kröner (1979a, b) suggested that similar subsidence in the Damara orogen was due to intense crustal stretching. Normal faulting did take place during the evolution of the Labrador geosyncline and generally is accompanied by some crustal expansion. However, the continuity of formations demonstrated by mapping suggests the presence of only a few wide fault blocks rather than the presence of an intensely faulted terrain. In my view this excludes major crustal stretching during geosynclinal evolution. Below I will propose a mechanism of such geosynclinal subsidence, based on ideas of Martin and Porada (1977) and Bird (1979).

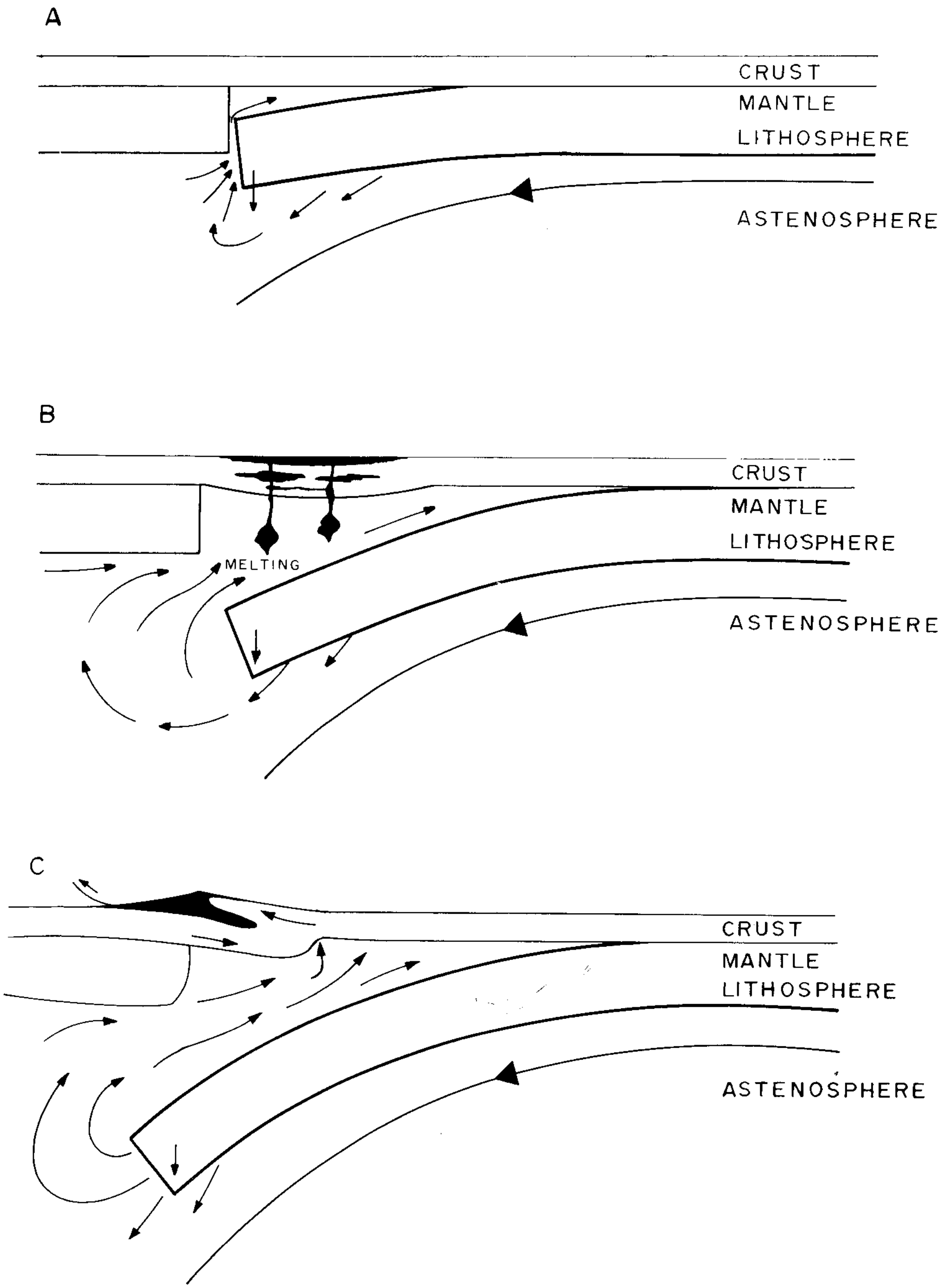
The pattern of overfolding and the direction of thrusting (Figs. 13-3, 13-5e) leave no doubt that the main process during the orogeny was the underflow of the western foreland of the Labrador Trough below its hinterland. The Labrador Trough, thus, is a clear example of A-subduction in the sense of Bally (1980). Structural cross-sections permit to estimate that the Labrador Trough has been shortened to roughly one half of its original width. Crustal subduction of an order of magnitude of 100 km can be deduced from the exposed structures. Furthermore, Dimroth and Dressler (1978) inferred that basement-cored nappes once covered the highly metamorphosed rocks of the immediate hinterland of the trough. The origin of these nappes would also add to crustal subduction, although perhaps not very much. It is noteworthy that the amount of crustal subduction inferred for the Labrador Trough (roughly 100—200 km) compares favorably with values of 50—270 km estimated by Bally (1980) for A-subduction in Phanerozoic fold belts.

The structural asymmetry of the Labrador fold belt (Figs. 13-3, 13-5e) suggests that the subducted basement has underplated the hinterland. Of course, I do not infer that the 100—200 km of "missing" basement are now to be found in a strip of equal width to the east of the fold belt. Subcrustal translation may have distributed this material over a much wider area. However, it should be noted that the intermediate-pressure metamorphism of the rocks of the hinterland of the trough suggests considerable uplift (by 10—20 km) which could be the effect of crustal thickening by compression, of crustal thickening by sialic underplating, or of both.

How could the origin of the Labrador Trough be related to plate tectonics? Of course, it is very likely that plate-tectonic processes or, at any rate, processes very similar to plate tectonics, were taking place since the Archaean. For example, the volcanic, sedimentary, tectonic and metamorphic evolution at the southern margin of the central part of the Archaean (2.75 Ga) Abitibi greenstone belt (Dimroth and Rocheleau, 1979) is most easily understood as a primitive island arc—forearc basin system. However, the "plates" involved were probably small, were moving rapidly, and were not rigid. Similarly, Hoffman (in press) presented good evidence that a plate-margin orogen formed in early Proterozoic time at the westernmost margin of the Canadian Shield.

Thus, it appears likely that plate-margin orogens and intra-plate orogens exist in the Canadian Shield, just as appears to be the case in Africa (Martin and Porada, 1977; Kröner, 1979a, b). The geosynclinal fillings and the orogenic movement patterns of both types of geosynclines, although not identical, are closely analogous. Both types of orogens should, perhaps, be considered the effects of the same basic process, namely the underflow of a descending convection current in the mantle. In one case this underflow would act upon a continental margin, in the other upon a zone within a sialic block.

I believe that the evidence presented in this paper is most readily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accommodated in the model of continental delamination of Bird (1978a, b, 1979). This model is based on the premise that subcontinental lithosphere is denser than the underlying astenosphere and, therefore, is inherently unstable. The astenospheric material, then, will rise in any fracture of the lithosphere and will spread laterally at the crust—mantle boundary. The lithosphere will thus be detached from the crust and will sink into the astenosphere. This process is called continental delamination.

Continental delamination, followed by massive injection of astenosphere-derived mafic and/or ultramafic magmas within the crust, can explain all features of the evolution of the Labrador geosyncline. A possible scenario is shown in Fig. 13-7. First, the lithosphere cracks and astenospheric material rises and spreads laterally (Fig. 13-7a). If this process takes place under the influence of astenospheric undercurrents, as I believe to occur, the astenospheric material will spread to one side only.

Bird's (1979) model calculations suggest that the immediate effect of such crustal delamination would be uplift of the crust by 1—2 km depending on the thickness of the crust. This uplift would again initiate the down-faulting of grabens and their filling by red beds (Fig. 13-5a).

According to Bird (1979) this uplift decays within about 100 Ma, a time sufficient to permit deposition of the shelf-type sediments (Fig. 13-5b). The further history of the Labrador Trough cannot be understood except by assuming massive intrusion of astenolith-derived mafic and/or ultramafic magmas into the crust (Fig. 13-7b). These intrusions, increasing the density of the crust, would first induce a short period of tectonic instability (Fig. 13-5c) and then a period of rapid subsidence during which we see the outpouring of several km of basalt on the sea floor (Fig. 13-5d).

Continuing sinking of the lithosphere, driven by a convection current, could finally account for the final effect of the orogeny (Fig. 13-7c). Such orogeny would be induced either by weakening of the crust, due to the mafic or ultramafic intrusions, or it could be induced by changes in the velocity of astenospheric currents.

Martin and Porada (1977) and Kröner (1979a, b) presented very similar models of the evolution of late Precambrian orogens in Africa. All these

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Fig. 13-7. The crustal delamination model for the evolution of the Labrador Trough.
A — The astenosphere rises in a fracture of continental lithosphere and spreads laterally at the crust—mantle boundary where astenosphere under-currents pull the lithospheric block down. Geological effects of this process are rapid uplift with faulting (Fig. 13-5a), followed by gradual subsidence (Fig. 13-5b) coincident with eruption of minor potassic basalt, carbonatite and alnoite.
B — Melting in the astenosphere produces voluminous mafic and/or ultramafic magmas. These are injected into the continental crust and, eventually, reach the surface. Geological effects are rapidly developing tectonic instability (Fig. 13-5c), followed by rapid subsidence and by voluminous basaltic volcanism (Fig. 13-5d). Lavas are low-K tholeiites.
C — The convection currents pulls the eastern block westward and the delaminated, thin lithosphere in the east overrides the stable crustal block to its west.

models still suffer from one major defect, namely that they have very limited predictive value. This defect is shared by plate tectonics since plate tectonic interpretations of Phanerozoic mountain belts also have had little predictive value. Rather than predicting which new geological observations could be made in specific regions, plate-tectonic interpretations have had to be readjusted in an ad-hoc fashion as new regional work has been done. This fact is a definite sign of the immaturity of our present understanding of orogenic processes and of their geological effects.

ACKNOWLEDGEMENTS

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Chapter 14

A GRENVILLIAN MODEL OF PROTEROZOIC PLATE TECTONICS

A. J. BAER

ABSTRACT

Uniformitarian and non-uniformitarian models proposed so far to explain the evolution of the Grenville Province have not proven entirely acceptable. Pre-Grenvillian evolution involved early Proterozoic reworking of a sialic crust, in part of Archaean age, and intrusion around 1500 Ma ago of a suite of anorthosites and associated felsic plutons. A major structural zone, the Chibougamau—Gatineau lineament, marks the western limit of large anorthosite plutons and may be a long-lived line of crustal weakness. The Grenville Group of marbles, pelitic rocks and metavolcanics dating from 1250—1300 Ma ago occupies part of the southern Grenville Province. It includes some of the few rocks that did not reach at least amphibolite facies of metamorphism. Geochronological data show two peaks of activity, at 1100 ± 50 Ma and 950 ± 50 Ma. In North America, the first corresponds to a tectonic disturbance that only affected parts of the central Grenville Province. The second was felt in all the province but was of lesser intensity. Similar conditions appear to have existed in east Greenland and in southern Scandinavia. In the foreland of the province, basaltic material filled dykes and grabens prior to its deformation, but this activity peaked during orogeny, around 1150 Ma ago.

Characteristics that are typical of Phanerozoic belts formed according to present-day plate tectonics have not been found in the Grenville Belt. It is argued that Proterozoic (and Grenvillian) tectonics involved intra-plate deformation but not subduction. Trends of dyke swarms, grabens, major shear-zones and folds are used to deduce the probable orientation of principal stress-directions at the time they formed. Stress-field orientations seem to have been similar inside and outside the Grenville Province. Palaeomagnetic hairpins faithfully match the structural evolution of the province. The drift that they indicate for eastern North America between 1200 Ma and 900 Ma ago may suggest motions of the continent inside a constant stress-field, or at least a stress-field independent of the drift of the continent. The stresses that were necessary for intracratonic deformation of the Grenville Belt could only build up because the lack of subduction caused plates to be locked together in "plate-jams". At present, the particular geometry of plates in regions surrounding Tibet keeps forcing the Indian craton into Asia, causing a localized plate-jam not entirely unlike those that accompanied the formation of the Grenville Province.

INTRODUCTION

Plate tectonics and subduction

It should not be necessary to define plate tectonics anymore, but if some things go without saying, they are even better said. Plate tectonics, therefore, is the theory and study of plate formation, movement, interaction and destruction; the attempt to explain seismicity, volcanism, mountain-building and palaeomagnetic evidence in terms of plate motions (Press and Siever, 1978).

Subduction is a process by which one plate (oceanic in most cases) sinks under another. In the present-day context of rigid plate tectonics, subduction is required to compensate the creation of new crust along oceanic ridges, or accretion zones. In fact, subduction zones are the most active and complex type of plate boundary on the present earth.

Except for the last 200 or 300 Ma of it, the Precambrian record shows none of the features associated with Phanerozoic subduction zones. The generally recognized lack of high pressure—low temperature metamorphic assemblages, the absence of ophiolites and of typical island-arc andesites have been interpreted by some to indicate a lack of subduction in the Precambrian, whereas others have argued that these rocks might have been removed by erosion. Uniformitarian and non-uniformitarian interpretations are thus possible. Considerations of thermal gradients have led Green (1975) and Baer (1977a) to suggest that Precambrian oceanic lithosphere was "unsinkable" because at high temperatures accompanying flatter gradients, oceanic lithosphere would be less dense than now and could not sink. This theoretical petrological argument strongly favours a non-uniformitarian approach to plate tectonics.

Plates are supposed to be rigid although the study of earthquakes in central Asia raises serious doubts in this regard (Tapponnier and Molnar, 1976). The aeromagnetic pattern of the western Churchill Province suggests an eminently ductile deformation on the scale of hundreds of kilometres (McGrath et al., 1977). If this pattern reflects the Hudsonian event, deformation was not limited to plate edges, but was intra-plate. Plate rigidity is therefore a concept that must be carefully redefined to avoid considerable confusion in applying plate tectonics to the Precambrian.

Palaeomagnetism has demonstrated the existence of Proterozoic continental drift, and although data are of uneven quality, the evidence in favour of more than one plate appears conclusive (McGlynn et al., 1975; *see also Irving and McGlynn, this volume, Chapter 23; McWilliams, this volume, Chapter 26, ed.*). If two or more plates move along the surface of the earth, it is correct to talk of plate tectonics, but because evidence for subduction is lacking and because plates were not all rigid at all times, I prefer a non-uniformitarian approach to Precambrian plate tectonics and I believe that plate motions without subduction were possible and were accommodated by intraplate tectonics and by deformation along plate boundaries.

Definitions

The Grenvillian event is conveniently defined, at least in first approximation, as the tectono-metamorphic event that affected the eastern Canadian

Shield around 1000 Ma ago. Geochronological evidence shows that deposition of sediments closely associated with this event did not start prior to about 1300 Ma ago, and that the last cooling of K-Ar systems was over by about 750 Ma ago.

The total area affected by the Grenvillian event is not known. The largest continuous block is the Grenville Province, that part of the Canadian Shield east of the Grenville Front, where K-Ar ages are commonly between 1100 Ma and 800 Ma. Farther south, geochronological data indicate the continuation of this belt into the western Appalachians, from Vermont to Tennessee, with age equivalents as far west as the Llano uplift of Texas. The width of this zone in Canada is commonly a few hundred kilometres, but it varies considerably.

Extension of the Grenville belt to other parts of the North Atlantic craton has provoked considerable discussion. It would now appear that undoubted Grenvillian supracrustals occur along the east coast of Greenland (Higgins, 1974). From geochronological evidence, it is possible that a major proportion of the Caledonides in East Greenland consists of rocks first deformed between 1200 Ma and 900 Ma ago (Steiger and Henriksen, 1972; Hansen et al., 1973a, b, 1974). In Europe, southern Sweden and southern Norway are the only areas of Grenvillian deformation not affected by a later event. However, as Grenvillian dates occur in the Caledonides of western Norway, from Bergen to the Lofoten Islands (Pasteels and Michot, 1974; Griffin et al., 1978), the conclusion is inescapable that a belt of Grenvillian deformation once extended along the North Atlantic. A possible extent through Europe and North Africa is far more problematic, and in spite of occasional speculations to that effect, well documented evidence of a Grenvillian "arm" into Europe and the Russian platform is still lacking.

In this paper, "Grenville Province" describes the eastern part of the Canadian Shield, whereas "Grenville Belt" refers to the entire strip of land affected by the Grenvillian event, from the southeastern United States to Greenland and Scandinavia (Fig. 14-1).

A choice of models

Geological studies in the Grenville Province began in the first half of the nineteenth century, and about 80% of the territory has now been mapped. The scale of maps varies from 1:30,000 or larger in parts of southern Ontario to 1 : 1 million reconnaissance. Tremendous variations in quality and age of adjacent map-sheets, heavily wooded topography, difficulty of access and complex geology make the Grenville Province a poorly understood part of the North Atlantic craton. Its enormous size (more than 1500 km long in Canada alone) means that extrapolation from small, better known segments to the whole province is fraught with difficulties. A model of evolution of the Grenville Province is by necessity a temporary figment of controlled

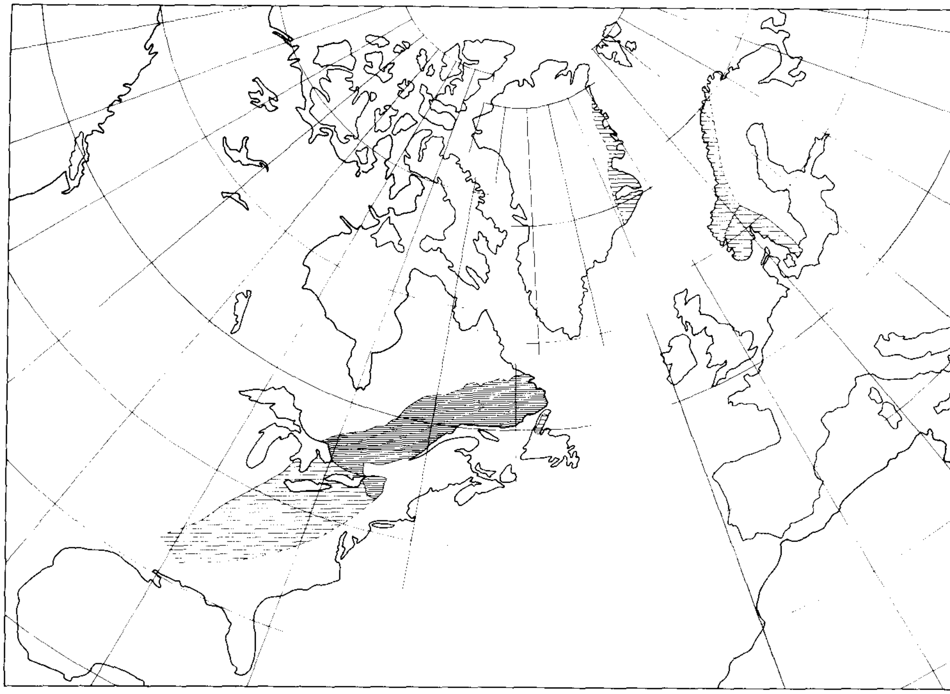

Fig. 14-1. Extent of the Grenville Belt and location of the Grenville Province (close lines).

scientific imagination. Such models have been developed along three main themes. In 1973 Dewey and Burke proposed an analogy between Tibet and the Grenville Province and suggested that the latter formed from a continental collision somewhere in or under the Appalachians. Variants of a collision model have been proposed by Irving and McGlynn (1976), Baer (1976), Brown et al. (1975) and extensive palaeomagnetic data have been used alternately to prove or disprove a possible collision until McWilliams and Dunlop (1978) showed that magnetisation was in all probability post-tectonic and not particularly relevant to the choice of a tectonic model.

By contrast with these essentially uniformitarian models, non-uniformitarian approaches have been followed by Wynne-Edwards (1976) and Baer (1977b). Wynne-Edwards considered the possible existence of deformable, ductile plates that would drift over sub-continental "hot-lines" (lines of hot-spots) and would deform in the process. Baer suggested that in the absence of subduction most of the drift of plates should be accommodated by steeply dipping shear zones, and proposed that the Grenville Province was essentially a "mega-shear".

None of these models accommodates all available data perfectly and the purpose of this paper is to present the known facts and yet another interpretation of the Grenvillian story. Since our factual knowledge of the Grenvillian event still allows for more than one interpretation, the one presented here is simply the scenario that I prefer at this time.

AN OVERVIEW OF THE GRENVILLIAN BELT

The geology of the Grenville Province in Canada has been most recently reviewed by Emslie (1970) and Wynne-Edwards (1972). A geological map of the Adirondacks (Isachsen and Fischer, 1970) incorporates most of the recent advances in that part of the Grenville Province. No recent comprehensive paper covers Grenvillian terranes of Scandinavia, but a few specific aspects have been treated, for instance by Patchett and Bylund (1977), Griffin et al. (1978) Klingspor (1976), Pedersen et al. (1978), Versteeve (1975) and others. The reader is referred to these papers for systematic descriptions. A general review of Grenvillian rocks and events in east Greenland can be found in Henriksen and Higgins (1976).

Pre-Grenvillian Proterozoic history

In the absence of geochronological data it is impossible to know what structures, what intrusions and what metamorphisms are truly Grenvillian, and not older, in many areas of the Grenville belt. With the progress of Rb-Sr whole-rock isochron and U-Pb zircon dating, the assumed effects of the Grenvillian event have progressively decreased in intensity or been confined to smaller parts of the whole belt.

Prior to about 1350 Ma ago, what is now the Grenvillian belt was a complex sialic crust continuous with adjacent parts of the North Atlantic craton. Its evolution can only be reconstructed in part, but it is important, because some of the structures and some of the intrusions that predate any Grenvillian event controlled Grenvillian deformation. The presence of structural discontinuities along parts of the future Grenville Front may well predate the creation of a Grenville Province, for instance.

In the northeastern Grenville Province pre-Grenvillian evolution can be unravelled in part because of the presence of sedimentary rocks, deposited in the Labrador Trough around or prior to 1800 Ma ago. They rest on Archaean basement and extend south of the Front into the Grenville Province, where their deformation must reflect the effects of multiple folding between 1800 Ma and 800 Ma ago. In the Grenville Province they were first deformed along ENE-trending axes, in a zone parallel with the Grenville Front. This earliest deformation was the strongest, caused the development of tight overturned folds and was accompanied by granulite facies metamorphism (Dalziel et al., 1969; Roach and Duffell, 1974). These early folds trend at a high angle to the Labrador Trough and appear to have been restricted to the Grenville Province, unless they are somehow correlative with folding of the Aillik Group in eastern Labrador.

North to north-northwest-trending folds represent a second phase of deformation inside the Grenville Province, where we would thus have at least two phases of folding of undetermined age but presumably younger than Hudsonian deformation. Metamorphism and deformation associated with this second event extended for an unknown distance into the Grenville Province, and if the general structural trend is any indication, this phase may have affected a 200 km wide zone extending southwest to the St. Lawrence River near the mouth of Saguenay River.

Emplacement of the anorthosite suite is well dated from Labrador at 1400 Ma to 1500 Ma (Emslie, 1978a; Krogh and Davis, 1973). K-Ar hornblende ages of 1533 ± 45 Ma and 1400 ± 57 Ma on the Mealy Mountains Complex, more than 100 km south of the Grenville Front (Emslie, 1978a), confirm that anorthositic bodies of the eastern Grenville Province are at least as old as those of central Labrador and predate any Grenvillian event. Various groups of clastic metasediments of unknown age, found along the St. Lawrence east of 64°W are possibly closely associated in time with the emplacement of anorthosites (Bourne et al., 1978).

The central Grenville Province (central granulite terrain and Baie Comeau segments of Wynne-Edwards, 1972) is poorly known, and most work has been concentrated in the western regions. There, geochronological studies (Davis et al., 1966; Doig 1977; Krogh and Hurley 1968; Krogh and Davis, 1969; Krogh et al., 1968, 1971) have shown that major metamorphic events did reset Rb-Sr whole-rock systems around 1600 to 1800 Ma ago. The Grenville Front near Sudbury was a structural element as early as 1500 Ma ago, as shown by ages of narrow granitic bodies that follow it southwest of Sudbury (Krogh et al., 1971). A Rb-Sr age of 1419[1] Ma on muscovite crystallized parallel to the down-dip lineation near the Front (Krogh and Davis, 1969) confirms the early age of this structure and shows that no major reheating accompanied the Grenvillian event in this area.

The time of emplacement of rocks of the anorthosite suite in the central and western Grenville Province has caused considerable discussion. On the one hand, the uniqueness of anorthosites and the continuity of a belt extending from Labrador into the central United States suggest that they should all be coeval, around 1400—1500 Ma ago (Emslie, 1978b). On the other hand, U-Pb dating on zircons and Rb-Sr whole rock isochrons give apparent ages of 1130 ± 10 Ma and 1124 ± 27 Ma (Silver, 1969; Barton and Doig, 1977). Furthermore, Martignole and Schrijver (1972) have reported blocks of Grenville Group metasediments as inclusions in the Morin anorthosite north of Montreal. As the Grenville Group is generally considered to be about 1300 Ma old, this observation tends to confirm a young age for the

[1] All Rb-Sr dates recalculated with $\lambda = 1.42$ (Steiger and Jäger, 1977).

anorthosites. I believe that final tectonic emplacement of the Morin and Adirondacks anorthosite plutons was synchronous with the last major metamorphism and deformation of the surrounding gneisses (Martignole and Schrijver, 1970) and is correctly dated by the closure of U-Pb and Rb-Sr systems, but primary segregation from the mantle and intrusion into the lower crust probably dates from around 1500 Ma ago.

The anorthosite belt seems interrupted by the Chibougamau-Gatineau lineament (Baer, 1976; Fig. 14-2). This lineament also appears to mark a boundary between lithologically monotonous, banded, light-grey, gently dipping quartzo-feldspathic gneisses to the west and structurally more complex, less well layered, locally garnet and biotite-rich gneisses to the east. Pyroxene-bearing gneisses in the granulite facies are more abundant to the

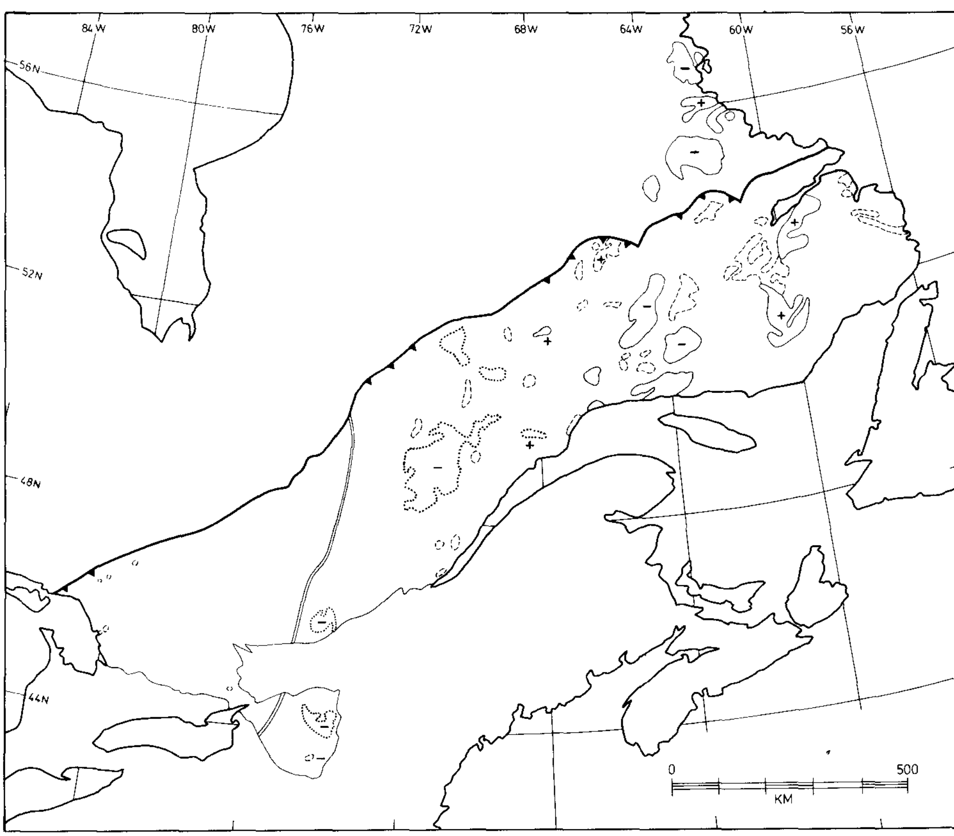

Fig. 14-2. Distribution of anorthosite plutons in the Grenville Province and Labrador. Dotted contours: deformed bodies; continuous contours: undeformed bodies; dashed contours: unknown. Plus signs refer to positive gravimetric anomalies, minus signs to negative anomalies. The Chibougamau—Gatineau lineament (double line) coincides with the western limit of extensive anorthosite outcrops.

east than to the west of the lineament (Bourne, 1978) but the Grenville Group appears to have been deposited on both sides of the lineament and to post-date its creation (Baer, 1976). A possible interpretation is that after emplacement of the anorthosites some vertical motion along the Chibougamau—Gatineau lineament accompanied a relative uplift of the central and eastern Grenville Province that brought the uppermost lower crust to surface. The area would have been eroded down to sea-level before deposition of the Grenville Group. The lineament would thus be a major structural element of the Province, pre-dating the Grenvillian event. In any case, and whatever its early evolution, the lineament was active during deformation and metamorphism about 1100 Ma ago (Wynne-Edwards et al. 1966; Dimroth, 1966). Its present nature reflects Grenvillian deformation.

Pre-Grenvillian events in other parts of the Grenville belt are known in less detail but confirm that Grenvillian events affected a continental crust comprising rocks with dominantly Proterozoic ages and some Archaean ages. The extent of an Archaean basement in the Province can be recognized from geochronology and from structural and lithological continuity with rocks of the Superior Province. Archaean rocks are found in the Front zone as far as 60 km inside the Grenville Province and possibly extend southeast for another 120 km in the Kempt Lake area north of Mont Laurier (Baer et al., 1977). Farther southeast they may be hidden under Proterozoic sequences or they may outcrop but have lost all isotopic evidence of their antiquity. No "Archaean" rocks have yet been reported from areas more than 200 km southeast of the Grenville Front.

Grenvillian history and isotopic dating

Grenvillian events are best understood in the western part of the Province, but evolution of this area appears to differ in many ways from that of northeastern regions. The Grenville belt can thus readily be divided into four segments, southwestern Grenville Province, northeastern Grenville Province, East Greenland, and Sveconorwegian Province.

Southwestern Grenville Province

Deposition of the Grenville Supergroup (carbonates, sands and pelites) occurred over an unknown area of the southern Grenville Province, where sedimentary rocks are associated with felsic and basic volcanics. Silver and Lumbers (1965) obtained a U-Pb zircon age of 1310 ± 5 Ma on volcanics near the base of the sequence, and this age has generally been taken for that of the Supergroup. Rhyolites from the Burnt Lake Formation have been dated at 1250 ± 25 Ma by the same method (Lumbers, 1967).

An alignment of nepheline-bearing syenites and nepheline-bearing gabbros dated at 1280 Ma (Krogh and Hurley, 1968) marks the approximate northwestern edge of the present extent of the Grenville Supergroup. These intrusives would thus be grossly coeval with deposition of rocks of the Supergroup and have been thought to underline a major rift-line (Baer, 1976; Fletcher and Farquhar, 1979). According to this hypothesis, the Grenville Supergroup would have accumulated in a subsiding cratonic graben because the great thickness of marbles (up to 15,000 m reported; Lumbers, 1967) would need to have been originally deposited close to sea-level. Around 1250 Ma ago, large trondhjemitic plutons were emplaced in the area occupied by volcanics of the Grenville Supergroup. The nature of this event remains elusive and no exact equivalent is known elsewhere in the belt. The younger Flinton Group (Moore and Thompson, 1972) of carbonate sandstones and shales rests unconformably on part of the older Grenville sequence and on one trondhjemitic pluton.

A major metamorphic and tectonic event affected the region around 1150 Ma ago (U-Pb on zircon, Silver and Lumbers, 1965). It was accompanied by metamorphism in the amphibolite facies, intrusion of granodioritic plutons, and is probably responsible for the northerly to north-northeasterly structural trends in the Grenville Group. Rb-Sr whole-rock isochron dates reflecting this event are found in a broad swath extending obliquely to the long axis of the Province from the Adirondacks to Lac St. Jean (Fig. 14-3). Final emplacement of anorthosites in the Adirondacks and north of Montreal probably coincided with this metamorphism. According to K-Ar dates on minerals, some reheating may have occurred at this time along the Grenville Front, but it is important to note that more than half the western Province (the Ontario and Québec gneiss segments of Wynne-Edwards, 1972) escaped the effects of this "orogeny".

Geochronological data are difficult to interpret because an unknown number of superposed phases of metamorphism has affected the Grenville belt. Dates ranging mostly from 2500 Ma to 700 Ma have been obtained by various methods. Pre-Grenvillian dates (mainly from Rb-Sr whole-rock isochrons and U-Pb ratios) have been recorded from the western Grenville belt, with the apparent exception of a north-northeast-trending zone from southeastern Ontario to the Labrador Trough (Fig. 14-3). This zone rarely yields isochrons older than 1200 Ma. I have attributed this apparent geochronological homogeneity to the effect of the early Grenvillian (or Morin) event, around 1150 Ma ago. Elsewhere, no clear peaks of pre-Grenvillian metamorphic activity have been recognized, in part because of the small number of available results, but also because as the number of dates increases in any particular region, the time-band that they define tends to widen.

Dates that spread between 1200 Ma and 800 Ma have generally been taken as evidence for one Grenvillian event. The considerable difference between extreme dates is caused mainly (but not entirely) by the fact that most K-Ar mineral ages and $^{39}Ar/^{40}Ar$ ages cluster around 900 ± 100 Ma, whereas Rb-Sr and U-Pb dates are older. This discrepancy has been attributed to late closure of K-Ar and ^{39}Ar-^{40}Ar systems, due to slow cooling of the belt

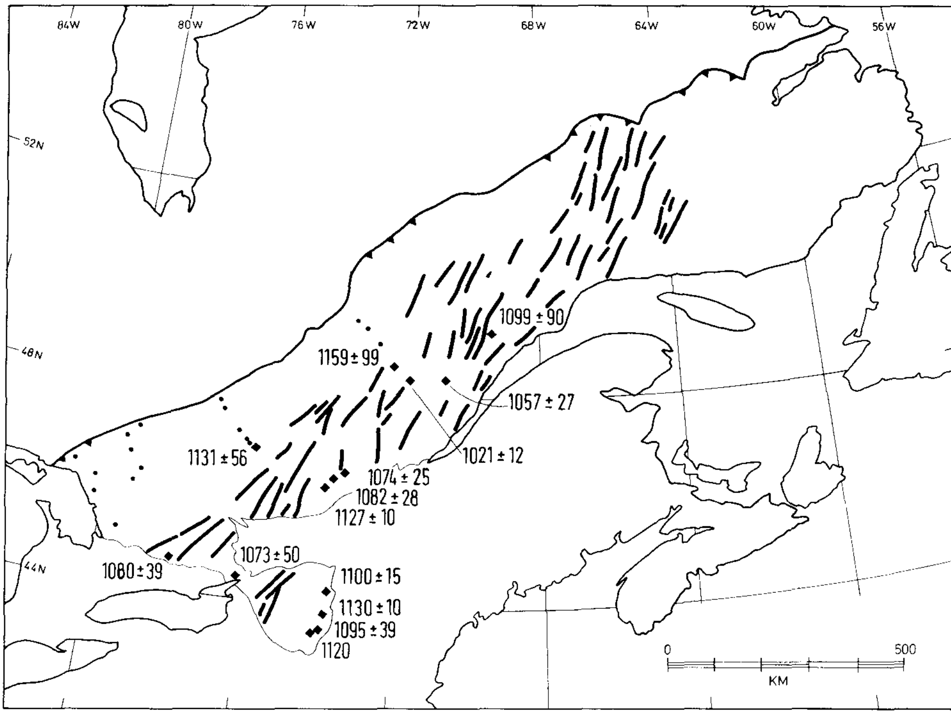

Fig. 14-3. Distribution of 1100 ± 50 Ma Rb-Sr whole-rock isochron dates in the Grenville Province. Circles correspond to isochrons giving older dates. Heavy lines represent structural trends. Note the coincidence of 1100 ± 50 Ma dates with the area of strong north-northeast structural trends.

(Harper, 1967). Although this phenomenon is well known from many other Precambrian shields, it would appear to be particularly well developed in the Grenville belt. Another possible interpretation, that this discrepancy actually reflects two distinct events, one around 1100 Ma ago and the other around 950 Ma ago, has been postulated by Baer (in prep.).

The great structural complexity of the Adirondacks in northern New York State can be resolved into four superposed phases of folding (McLelland, 1977). The earliest formed isoclinal and recumbent folds of easterly trend. The second is approximately coaxial with the first, but axial planes tend to be upright. F_3 folds are best developed in the eastern Adirondacks and trend north to northeast, whereas fourth phase folds are prominent in the northwest where they trend northwest with upright axial planes. Anorthosites were intruded prior to F_1 but no detailed geochronological framework is yet available to date the various phases of folding. An early proposal that anorthosites and associated rocks represent the basement of the Grenville Supergroup (Walton and de Waard, 1963) is hardly tenable any longer. The presumed unconformity at the base of the Lower Marble Formation may be entirely tectonic, and "basement gneisses" may be the next layer down in a continuous stratigraphy (Isachsen et al., 1975). Furthermore, field work in the southern Adirondacks has resulted in the recognition of a large number of anorthositic sills that intrude at stratigraphic horizons lying far up the supracrustal sequence (Isachsen et al., 1975).

The Adirondacks thus show evidence of three or more periods of deformation, intrusion and metamorphism of different areal extent. It is too early to date the various phases of folding in absolute terms.

Northeastern Grenville Province

The geology of the central part of the Grenville Province is poorly understood. The extreme northeastern part of the Province has been better studied, although much remains to be done. The last deformation consisted of steep reverse faulting and thrusting that accumulated a pile of south-dipping slices and pushed them on and over the foreland (Emslie et al., 1978). Abrupt changes in metamorphic grade (differences of 4 kb and 300°C near the Front at Ossokmanuan Lake; Bourne, 1978) and the presence, at surface, of sillimanite-orthopyroxene assemblages stable in the very deep crust suggest that some thrust-slices were brought up from the lower crust. About 100 km from the Front, in the Grenville Province, the Mealy Mountains anorthosite complex is in greenschist or low amphibolite facies, and its hornblende has preserved a K-Ar pre-Grenvillian age. Some blocks must thus have maintained their pre-Grenvillian elevation, whereas others were thrust over considerable distances. Late-Grenvillian tectonics are fault-controlled in this area. Thrusting along the Front must date from about 950 Ma ago because of K-Ar ages of micas in folded Seal Lake Group volcanics of the foreland. The age of regional metamorphism may be older, but must be younger than 1324 Ma since it affects the Seal Lake Group (Baragar, 1978b). Farther north, outside the Grenville Province, adamellites and granites have been dated at 1138, 1150, 1160, 1165, 1170 and 1175 Ma respectively (K-Ar on hornblende and biotite, Wanless et al., 1972, 1973) and ten syenitic bodies from southwest Greenland date between 1119 and 1159 Ma (Blaxland et al., 1978). Dates of 1150, 1165 and 1185 Ma (K-Ar on biotite) from inside the Grenville Province in the same region probably reflect the same thermal event (Geological Survey of Canada, 1970) and possibly date the metamorphism of anorthosite plutons. South and southwest of the Labrador Trough the latest tectonic event along the Front also involved shearing and thrusting along northeasterly trending zones accompanied by blastomylonitization and retrogressive metamorphism (Roach and Duffell, 1974). It presumably dates from around 950 Ma ago.

The Grenville Province thus shows evidence of two "Grenvillian" events; a late, often brittle deformation that started K-Ar clocks around 950 Ma ago and an earlier event of more limited extent but greater intensity (Fig. 14-3), dated from about 1100 to 1150 Ma ago. This early event imparted a consistent north-northeasterly structural grain to areas that it affected, by contrast with the later, less intense (?) deformation. To avoid confusion, I have called the 1150 Ma event the Morin event, from an anorthosite pluton deformed by it, and have kept "Grenvillian" to describe the 950 Ma event (Baer, in prep.).

Greenland

Part of the Caledonide belt of east and northeast Greenland contains reworked material of Grenvillian age, and detailed studies will undoubtedly reveal more of it. One such succession is the Krummedal supracrustal sequence of Scoresby Sund area (Higgins, 1974; Escher and Watt, 1976). The succession comprises more than 8000 m of rather monotonous, banded quartzitic and pelitic metasediments. The upper part of some sections is more quartzitic and, on Hinks Land, the lowest visible levels (that may be the true base) comprise minor marbles and amphibolites. Breccias and agglomerates have been recognized, suggesting that the amphibolites correspond to a volcanic episode. The rocks were brought up to amphibolite facies, but were retrograded later during Caledonian deformation. The major phase of metamorphism has been dated at 1162 ± 85 Ma (Hansen et al., 1974) and overprints the main north-trending folds that are moderately tight to isoclinal and overturned to the west. According to Higgins (1974) many dates in the 950—1150 Ma range are probably related to the same orogenic episode.

Elsewhere along the east coast of Greenland, metamorphic crystalline complexes, considered originally to be Caledonian, have given dates in the 1100 Ma range. Zircon ages on augen granites of the Gåsefjord—Stauning Alper region (70—72°N) give 950 Ma (Steiger and Henriksen, 1972). A preliminary whole rock isochron at 1060 Ma and Rb-Sr mineral ages as old as 1150 Ma all confirm that the region suffered metamorphism in Grenvillian times (Hansen et al., 1974). Farther east, K-Ar hornblende ages of 1189 ± 55 Ma and 1124 ± 50 Ma show that the intensity of Caledonian reheating was locally very low.

The poorly known northeast Greenland area, between 76°N and 81°N, was deformed by a "Carolinidian" orogeny (Haller, 1961) tentatively dated from about 1000 Ma ago. On Dronning Louise Land, the sedimentary succession comprises 200—300 m of basal quartzites and limestones, 2000—3000 m of semi-pelites and 3000 m of psammites. K-Ar age determinations on dykes that cut these sedimentary rocks in the foreland have yielded ages of 982 ± 19 Ma and 799 ± 69 Ma (Henriksen and Higgins, 1976). The date of 982 Ma was confirmed by a $^{39}Ar/^{40}Ar$ spectrum analysis that gave 988 ± 20 Ma.

Scandinavia

The Sveconorwegian Province of southern Sweden and Norway has been recognized for some time as a probable geochronological equivalent of the Grenville Province of Canada (Kratz et al., 1968). It is made up of polymetamorphic gneisses last deformed around 900 Ma ago. The Dal Group rests unconformably on these gneisses west of Lake Vännern and its deposition and deformation have been bracketed between 1240 ± 30 Ma, the age of the basement (Welin and Gorbatschev, 1976) and 910 ± 35 Ma, the age of the late-tectonic Bohus granite (Skiöld, 1976). The group consists of basal sandstones and conglomerates grading up into "lower slates" with calcareous intercalations. These rocks are overlain by spilitic horizons, "upper slates", quartzites and more slates (Skiöld, 1976). Dal supracrustals were deformed first by north-trending folds and, in what was probably the continuation of the same phase, into west-dipping thrust-slices separated from each other by shear-zones and mylonites (Laarson, and Sandgren, 1956). Rb-Sr isochrons on rocks from southern Norway indicate the existence of two tectonic and geochronological events. The latest one is dated from 900—1000 Ma (Pedersen et al., 1978). It presumably corresponds to the last deformation of the Dal Group. An earlier event dates from about 1100 Ma in Rogaland but its age is less clearly defined in the Ostfold area. The two phases of folding are coaxial in some regions and correspond respectively to the Dalslandian (900—1000 Ma) and the Ostfoldian (1100 Ma) (A. Berthelsen, pers. commun., 1979).

Geophysics can help

Geophysical data confirm and amplify geological information. Thickness of the crust varies from a maximum of about 50 km along the Grenville Front west of the Labrador Trough (Berry and Fuchs, 1973) to 30 and 33 km near Lac St. Jean (Mereu and Jobidon, 1971). Elsewhere, it appears to vary between 35 km and 45 km. Comparison with the gravity map (Earth Physics Branch, 1974) suggests that anomalously thick crust may underlie the Grenville Front zone between Lake Mistassini and the Labrador Sea, but not the western extension of the Front. The same map shows that some anorthositic plutons in the northeastern part of the Province correspond to positive gravity anomalies, whereas others correspond to negative ones. As anorthosite has a low density (~ 2.65—2.70) positive gravity anomalies can only be explained by the presence of dense rocks at depth under most anorthositic plutons east of a line running from the mouth of Saguenay River to Seal Lake (Kearey and Thomas, 1979). Anorthosites east of this zone are gravimetrically light, and they have been deformed and metamorphosed. Since those located to the east are not deformed, a relationship must exist between deformation and absence of mafic root. As the Lac St. Jean and Adirondacks massifs are partly sheet-like (Simmons, 1964; Kehlenbeck, 1972) they may have been thrust and cut off from their roots during deformation.

Tectonics of Grenvillian surroundings

If the Grenville belt represents a major mobile zone of the earth's crust, the study of immediately adjacent areas cannot be neglected. The Himalayan collision, for instance, had a considerable effect upon the tectonics of Central Asia (Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1976) and one may look for similar relationships along the Grenville belt. Accessible parts of the hinterland are the eastern Canadian Shield, including southern Greenland, and the southeastern Baltic Shield. Other regions are either covered by younger formations or have been deformed in later orogenies.

In the eastern Canadian Shield west of the Grenville Province tectonics of the 1300—800 Ma interval are those of a brittle platform. Major events are the formation of grabens (Keweenewan, Seal Lake, Gardar, Figs. 14-4, 14-5) and the intrusion of basic dyke swarms. The earliest activity was apparently in the Seal Lake graben (1324 ± 92 Ma; Baragar, 1978b) and in its approximate continuation in Greenland, the Gardar region (Berthelsen and Henriksen, 1975; Fig. 14-4).

Activity in the Keweenawan graben was dated at 1140 Ma to 1120 Ma by Silver and Green (1972) but more recent measurements on Michipicoten

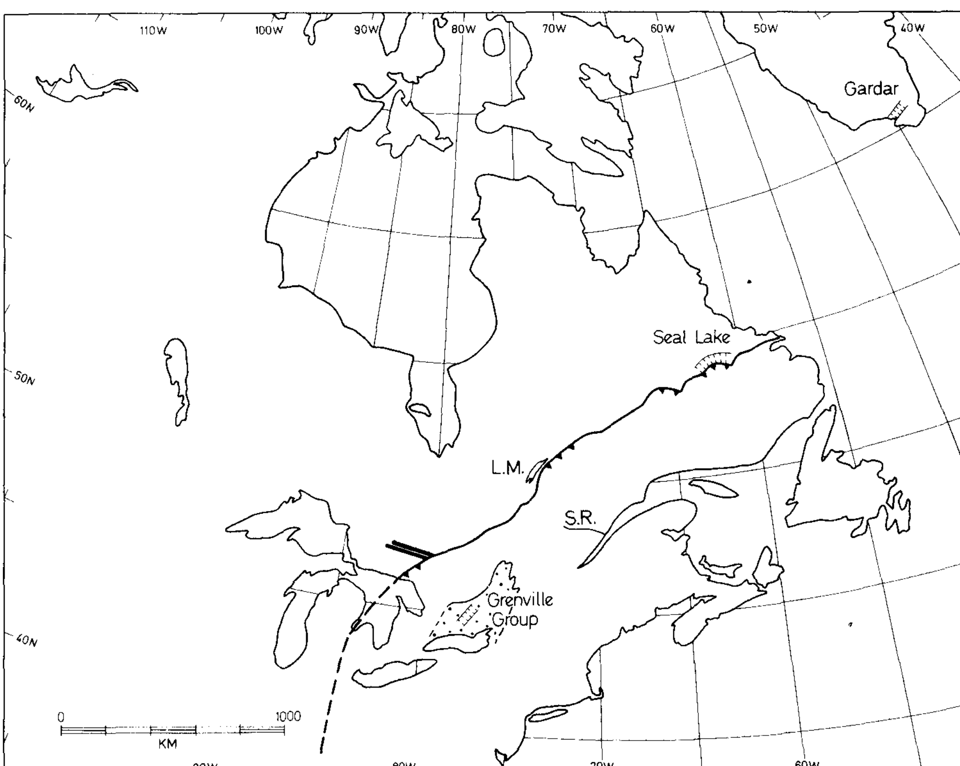

Fig. 14-4. Dykes and graben of the eastern Canadian Shield around 1300 ± 50 Ma ago. The Grenville Group is dotted in. *L.M.* = Lake Mistassini; *S.R.* = Saguenay River.

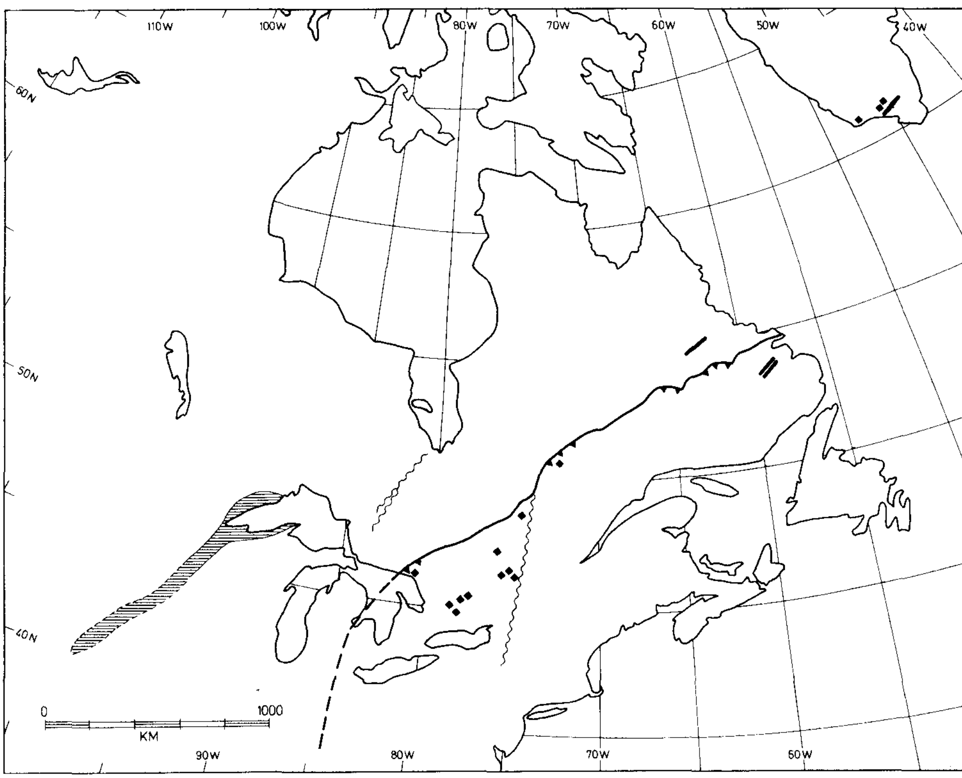

Fig. 14-5. Dykes, graben (close lines) and alkaline intrusives (diamonds) of the eastern Canadian Shield around 1100 ± 50 Ma ago. Alkaline intrusives after Baragar (in Baer et al., 1974). Grenville front indicated for reference.

volcanics (1004 ± 64 Ma; Baragar, 1978a) suggests protracted or sporadic activity over at least 100 Ma. The time of early Keweenawan activity coincides with a second, major phase in the Gardar (1160 Ma to 1150 Ma; Fig. 14-5). The direction of main dyke swarms and graben is approximately parallel to the Grenville Front.

In southern Sweden various dolerite dyke swarms are parallel to, but outside of, the Sveconorwegian Schistosity Zone (the eastern boundary of the Grenville belt)(Fig. 14-6). Their ages, measured from Rb-Sr mineral isochrons, run from about 1000 Ma to 900 Ma (Patchett and Bylund, 1977). These are thought to be true ages of intrusion because the rocks cut by the dykes have not been modified by any Grenvillian thermal effect. If their position relative to the Front is thus comparable to some Canadian examples, their ages are sufficiently different to correspond to different events.

A GRENVILLIAN MODEL

No simple, general model explains as yet the tectonic evolution of the

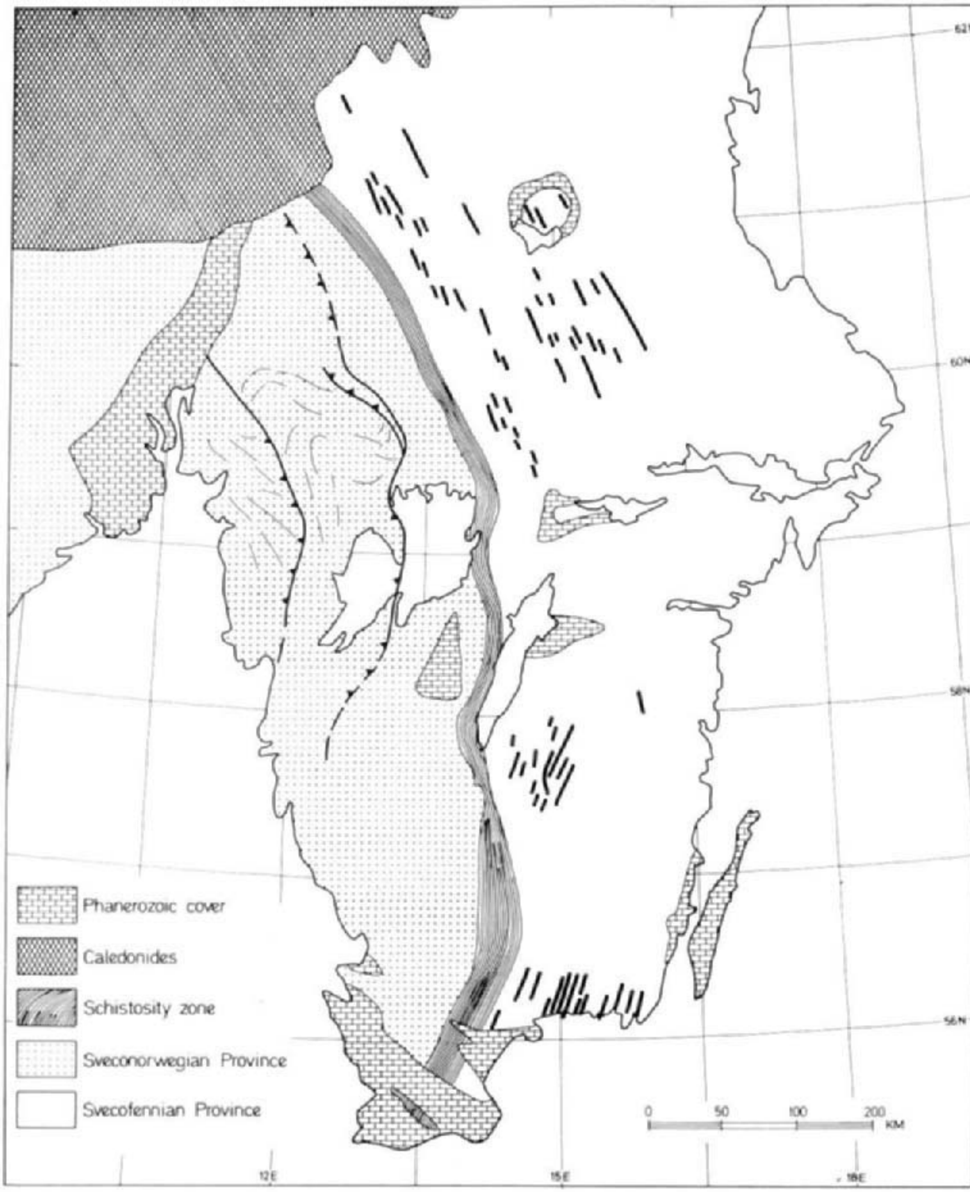

Fig. 14-6. Distribution of 950 ± 50 Ma old dolerite dykes (thick lines) in the foreland of the Sveconorwegian Province. After Patchett and Bylund (1977); structures in the Sveconorwegian Province after A. Berthelsen (pers. comm., 1979).

Grenvillian belt. The plate tectonics hypothesis has been particularly successful in coordinating a wealth of data about the present-day earth and in explaining with simplicity and elegance the tectonics of the Cenozoic. As it gained momentum and made more converts, this hypothesis was first extended to the Mesozoic, then to the Palaeozoic and finally to the Precambrian. However, whereas for the present earth geophysics and ocean-floor geology bring their essential support to the theory, these methods are not available to students of the Precambrian. Acceptance or refusal of a uniformitarian plate-tectonic model must rest on geological arguments alone.

The failure of Phanerozoic models

Examination of present-day tectonics and of Cenozoic geology allows one to define criteria for plate tectonics, or at least to recognize some characteristic signatures. The nature of these signatures may be stratigraphic, lithological, chemical, petrological, structural or palaeomagnetic. Some are not necessarily unique to plate tectonics and are therefore more difficult to interpret. How well does the Grenville belt satisfy these criteria?

A stratigraphic prism (geosyncline) characteristic of passive continental margins is common to many Phanerozoic belts, but is completely missing from the Grenville belt. Supracrustal rocks of Grenvillian or immediately pre-Grenvillian age are of limited extent. They were deposited upon continental granitic basement (Dal Group) or most probably so (Grenville Supergroup). The Krummedal succession is the only one that may have accumulated along the edge of a basin. "Passive-margin" sedimentation cannot be recognized along the belt.

Some phases of the Phanerozoic Wilson cycle are accompanied by the formation of characteristic rock-types. These are, for instance, glaucophane schists, ophiolites and island-arc andesites. These "type-fossils" of plate tectonics have rarely, if ever, been documented from the Grenville belt. The closest to ophiolites is a narrow zone of ultramafic rocks in southeastern Ontario, interpreted as possible ocean-floor (Brown et al., 1975). Elsewhere, ultramafic rocks are only represented in rare mafic plutons of undetermined age. High grades of metamorphism and intense deformation might, however, have destroyed possible lithological evidence. Attempts at recognizing a typical rock chemistry suffer from similar difficulties. For instance, although the chemistry of some metavolcanics in the Grenville Supergroup may be andesitic, that of others is basaltic with alkaline affinities (Lumbers, 1967; Sethuraman and Moore, 1973; Condie, 1975).

The fact that chemical analyses of Grenvillian metavolcanics may not be compatible with island-arc characteristics indicates only that chemical methods cannot by themselves discriminate in favour of one particular model.

Numerous Phanerozoic mountain belts owe their shape and the overall parallelism of their structural elements and of their metamorphic zones to the collision of lithospheric plates. This characteristic structural grain is hard to erase in later orogenics. In the Grenville Province the dominant structural grain is north-northeast, oblique to the elongation of the belt

(Fig. 14-3). It may coincide with a central subprovince, intensely deformed around 1150 Ma ago. Another structural zone of similar or older age runs northerly through southern Scandinavia. Still others may be found in east Greenland. Whatever their origin may be, they appear to be discontinuous and to have left large blocks of undisturbed continental crust between them (e.g. western Grenville Province and northeastern Grenville Province). With some exceptions, like for instance the granulite facies metamorphism of Rogaland in southern Norway, deformation by the later Grenvillian event (1000 Ma to 900 Ma ago) seems to have been more brittle, and metamorphism weaker. Although the entire Grenville belt was apparently affected, it did not acquire from it much of a common structural grain, except for a well-defined Grenville Front. Thus, although deformation might have been caused by plate collision, what is known of the structural history is unlike that of Phanerozoic mountain belts. Modalities of the orogeny were not those associated with plate collision initiated by subduction.

In favourable circumstances continental collisions may be recognized from their characteristic palaeomagnetic signatures (Irving and McGlynn, 1976). Results from metamorphic terranes are difficult to interpret, however, because the time of magnetization of the rocks cannot always be determined satisfactorily. Numerous studies undertaken in the Grenville belt have revealed complex cooling histories, but no simple collision path can be deduced from apparent polar wander curves.

In summary, criteria that individually or in combination are most obviously linked to plate collisions and orogenies do not seem applicable to the Grenvillian belt. This "failure" of usual Phanerozoic criteria must mean that the evolution of the Grenville belt followed some other pattern.

Dewey and Burke (1973) described the uplift and evolution of the Tibetan Plateau and showed that it formed as a consequence of the collision of India with Asia. Reasoning by analogy, they proposed that the Grenville Province was a deeply eroded equivalent of Tibet. The corresponding collision zone and the former suture would now be hidden under or in the Appalachians. The analogy is only partial, however. The Sveconorwegian Province lies east of the Appalachians and has a strong easterly vergence along its front. As this frontal "Schistosity Zone" is structurally similar to the Grenville Front, the Sveconorwegian Province should also be analog to the Tibetan Plateau. However, this bilateral symmetry of the Grenville belt differs fundamentally from the characteristic asymmetry of the Himalayan—Tibet orogen and it cannot be explained in the same terms. It is particularly difficult to reason by analogy when one compares a poorly known region with a lesser known one and when the effects of an assumed difference in levels of erosion of 10 km or more (between Tibet and the Grenville belt) cannot be assessed with any rigour.

Since popular models of orogenic belts do not explain the evolution of the Grenville belt, it is intriguing to find out how unique it really is, and to discover what controlled its evolution. Towards this goal, it becomes necessary to regroup available facts into "reasoning units" or partial models and to coordinate those into increasingly complex patterns (Harrington, 1973). The model proposed here can only reflect the state of the art in one person's mind at one point in time. It is presented to provoke discussion and to contribute to our understanding of mid- to late Proterozoic tectogenes.

Stress-fields of Proterozoic plates

Rigidity of the lithosphere must have been controlled by temperature and thickness, which is temperature-dependent. Much has been written about high Precambrian geothermal gradients but cooling curves of the earth, curves of radioactive decay and calculated "fossil" geotherms only show minor changes from 1300 Ma to the present. There is no evidence for a greatly increased lithospheric ductility in mid-Proterozoic times. By contrast, thickness of the continental lithosphere may have doubled since that time (Baer, 1977a). It probably reached only 60—70 km prior to Grenvillian deformation and would have been deformable under smaller differential stresses than it is now. If stresses in the lithosphere are caused (in part) by asthenospheric convection, and if the latter tends to slow down with time, the cumulative effect of greater stresses applied to thinner lithosphere may have made intracontinental orogenies commonplace in the Proterozoic.

Neotectonics of Asia shows that, under special circumstances, even thick and so-called rigid continents are deformed for thousands of kilometres away from collision zones (Molnar and Tapponnier, 1975). Analogous considerations have been applied successfully to deformed Proterozoic regions of Greenland and Canada (Watterson, 1978). The complex shape and the structure of the Churchill Province and particularly of its eastern part show that between 1800 Ma and 1700 Ma ago, this irregular shaped block of crust, at least 1000 km wide, was not acting like a rigid plate. Although one may wish to look for possible collisions along plate boundaries, it remains that deformation and metamorphism must have affected large volumes of the plates themselves. Linear orogenic belts along plate boundaries seem to have had little relevance in a Proterozoic context, at least as late as 800 Ma ago.

The episodic nature of tectono-metamorphic events came out clearly from early programmes of K-Ar dating (Gastil, 1960; Stockwell, 1961). This means that except for relatively short periods of time (the "orogenies") stresses acting upon Proterozoic continents were too weak to cause their deformation. It also means that because mechanical properties of the lithosphere do not change spontaneously, most probable causes for such changes are temperature increases, or increases in differential stresses, or both. Since the mantle is convecting, its upper surface is thermally heterogeneous, and plates will ride over warmer and colder zones. Because drifting plates do not all move around the same pole of rotation or at a constant speed, they will interact with each other and be pushed or pulled in various directions by a combination of asthenospheric drag from below and "jostling" from the sides. Stress fields applied to them will therefore vary in intensity and in direction.

Is it reasonable, or indeed possible, to postulate stress-fields of Precambrian orogens? Heterogeneity and anisotropy of continental crust may not be such a serious obstacle after all. Indeed, at the scale of a continent, anisotropies will tend to cancel each other. Also, basic dykes are known to cut across all types of rocks and structures for hundreds of kilometres without changing strike, thus demonstrating that the stress-field responsible for their formation was constant over large areas. In the Canadian Shield for instance, the Mackenzie dyke swarm is over 2000 km long, from Coronation Gulf to Lake Winnipeg. At that scale, possible stress indicators can only be major tectonic elements such as dyke swarms, rift-graben, thrust planes, shear-zones or fold trends of regional extent. A simple picture may thus emerge with the help of such simplifying assumptions. Recent attempts on Asia (Tapponnier and Molnar, 1976) and on Greenland (Escher et al., 1976) have given encouraging results.

Palaeomagnetic wanderings

Palaeomagnetic apparent polar wander paths for North America show dominantly equatorial tracks separated by hairpins that correspond closely to periods of deformation (Irving and Park, 1972; Baer, 1979; *see also Irving and McGlynn, this volume, Chapter 23, ed.*). Palaeomagnetic studies are unable to measure longitudes, but assuming the simplest possible polar wander path, possible corresponding motions of a continent may be accounted for. In the case of the Grenville belt, only North American data can be taken into consideration because of the uncertainty in palaeoposition of Scandinavia relative to the Canadian Shield. As the weight of the evidence, both palaeomagnetic and structural, suggests that the Grenville Province was a part of North America at that time (the so-called one-plate-model), the wander path of the Province will match closely that for the rest of North America.

Tectonic events in and around the Grenville Province are bracketed between ~ 1400 Ma and ~ 800 Ma ago. During this time, the corresponding apparent polar wander path went through a north-closing loop (the Logan Loop) then through a south-closing loop (Grenville Loop) to join the Cambrian track, possibly by going over the present pole (Morris and Roy, 1977) (Fig. 14-7).

Approximate motions of the Grenville Province during this period were a southerly drift of more than 80° between ~ 1400 Ma and ~ 1150 Ma, an apparently rapid northerly drift of about 90°, another southerly move by the same amount until around 950 Ma ago and a renewed northerly drift

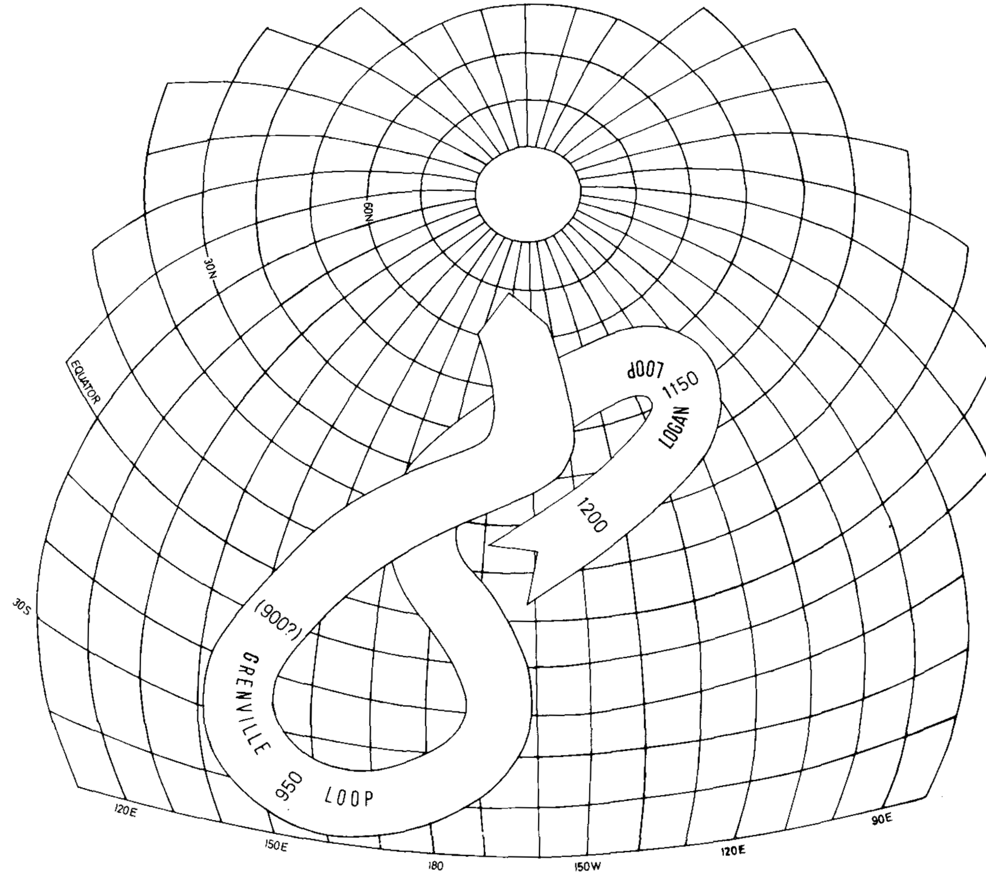

Fig. 14-7. Palaeomagnetic apparent polar wander path for North America between 1200 Ma and ?800 Ma ago. After Irving and McGlynn (1976) and Morris and Roy (1977).

thereafter. In most cases, uncertainty about the true age of magnetisation is the largest single source of possible error in latitude, but the two palaeomagnetic hairpins correspond to the two tectonic events recorded from the Grenville Province. It is easy to see that, independently of any thermal considerations, each change in drift pattern must mean a change in stresses applied to the continent. The same is true of earlier Proterozoic tectonic events, each one of which corresponds to a palaeomagnetic hairpin. This typically Precambrian polar wander path is not recorded from rocks younger than ~ 800 Ma, a time that seems to coincide with the beginning of subduction (Baer, 1977a). Changes of differential stresses accompanying Proterozoic "orogenies" are caused by abrupt changes in direction (and speed ?) of continental drift. Phanerozoic orogens are fundamentally different in that they are associated with collisions that would be impossible without subduction.

Three Grenvillian stress-fields

Geochronological data on the tectono-metamorphic evolution of the eastern Canadian Shield suggest a division of the ~ 1400 Ma to ~ 700 Ma interval into four periods. These correspond to a pre-tectonic, a syn-tectonic, a late-tectonic and a post-tectonic stage. Activity would peak around ~ 1300 Ma for the first, at ~ 1150 Ma for the second and at ~ 950 Ma for the third. Data are not always comparable and often are not numerous enough or of sufficiently high quality to date major events better than within ± 50 Ma.

The first stage corresponds to the early fracturing of the craton and is characterized by intrusion of dyke swarms and alkaline plutons, by the formation of rifts and grabens and by extrusion of associated lavas. Major activity in the Gardar Province and in Seal Lake area date from this time. In the Grenville Province, nepheline syenites and volcanism in the Bancroft area may represent similar environments (Baer, 1976). Between these two regions a series of alkaline intrusive plugs of uncertain but pre-tectonic age may belong to the same period (Baragar, in Baer et al., 1974), and the same is possibly true of the Harp dykes, cutting the Harp Lake anorthosite pluton (Meyers and Emslie, 1977).

Opening of the Gardar and Seal grabens must have meant that the least principal stress, σ_3, was oriented northwest (all such directions refer to present geographical coordinates). Faults and dykes of the Ivigtut area indicate that σ_1 was horizontal in a northeast direction (Berthelsen and Henriksen, 1975) so that the stress-field is reasonably well determined. The primary trend of lavas, dykes and plutons that are now metamorphosed and deformed inside the Grenville Province cannot be determined, but the overall map-pattern of the Bancroft area suggests a gross primary distribution of these rocks in a northeasterly direction. They would thus be compatible with the same stress-field as the Seal and Gardar rifts. In fact, the presence of nepheline syenite and the radiometric composition of lead in marbles (Fletcher and Farquhar, 1979) suggest a rift environment. It is possible, therefore, that the overall stress-field of the eastern Canadian Shield was everywhere the same with σ_1 vertical, and σ_2 northeast and σ_3 northwest in the horizontal plane (Fig. 14-8). In the northwestern Shield the Mackenzie dyke swarm runs northwest and has been dated from around 1200 Ma ago. It clearly does not belong in the same stress field as the Gardar or Seal swarms. Most maps show an extension of the Mackenzie swarm into the eastern Shield, but the dating of the swarm in this region is most uncertain. It is impossible at this stage to draw accurate boundaries to the regions where σ_3 was trending northwest.

This first period lasted at least until about 1100 Ma ago, because the first activity in Seal and Gardar grabens was succeeded, after some tectonic quiescence, by more dyking in Gardar, emplacement of Harp dykes and opening of the Keweenawan rift-graben. This renewed activity corresponds

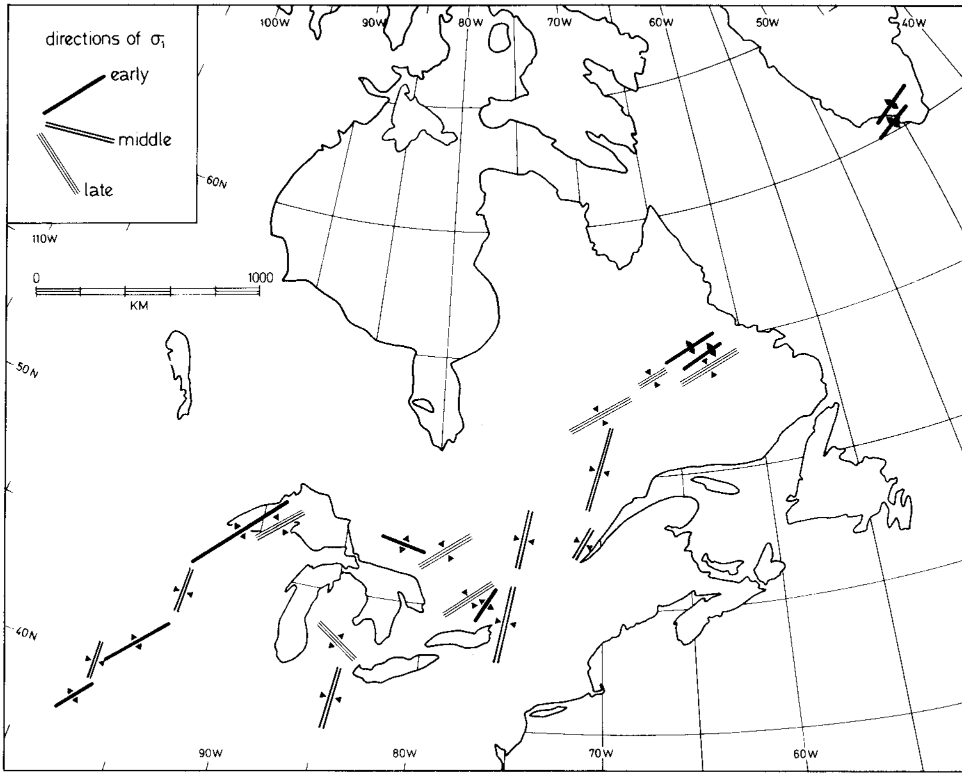

Fig. 14-8. Changes in the assumed stress field of the eastern Canadian Shield between 1300 ± 50 Ma ("early") and 950 ± 50 Ma ago ("late"). Arrows pointing towards the lines indicate the direction of σ_1; arrows pointing away from the lines indicate the direction of σ_3. Other explanations in the text.

to the same stress-field as the first, with σ_3 in a northwesterly direction.

The next phase is harder to date accurately, but it corresponds to the peak of high-grade metamorphism in the Grenville Province and to a deformation fabric at high angle to the Grenville Front (that of the Morin event). Structures used as stress indicators are the trends of major fold axes and those of thrust-planes that may be associated with the peak of metamorphism. They are difficult to recognize in the numerous regions where reconnaissance mapping has been little more than a lithological cartography. Fig. 14-8 was drawn assuming that σ_1 was sub-horizontal and perpendicular to major fold axes. As shortening appears to have generally occurred in a northwesterly direction, and because vertical shear-zones have rarely been recognized, σ_3 was probably vertical and σ_2 horizontal. The direction of σ_1 appears to run in an easterly direction, but cannot be determined accurately.

The third phase is the Grenvillian event sensu stricto. It was marked by thrusting and block-wise uplifts in the Grenville Front zone. Beyond the Front, folding and minor metamorphism of the Sims Group (Wynne-Edwards, 1961) probably belong here. Systematic strike-slip faulting of the Keweenawan is compatible with the same event. A date of 1004 ± 64 Ma on volcanic rocks on Michipicoten Island (Baragar, 1978a) and another age of 1030 Ma on a northwest trending dyke on the north shore of Lake Superior (Geological Survey of Canada, 1970) confirm renewal of activity in the rift at that time. Except for the Front zone, this stage is often difficult to recognize and to document structurally in the Province. It may be represented near Bancroft (Fowler and Doig, 1979) and may also be responsible for some left-lateral displacement along the Chibougamau-Gatineau lineament (see Dimroth, 1966). This phase coincides with the closure of all K-Ar systems in the Province and must therefore have meant a general uplift accompanied by a thickening of the crust. Principal stress directions again appear to be constant over most of the eastern Shield (Fig. 14-8), σ_1 is horizontal and perpendicular to the Grenville Front, σ_2 is parallel with the Front, and σ_3 is vertical.

No useful stress indicators accompany the post-orogenic and last phase of igneous activity. Locally, 750 Ma old diabase dykes trend northwest in southwestern Ontario and north on the Long Range peninsula of Newfoundland (Wanless et al., 1973). In the last 800 Ma differential stresses have no longer been able to deform the Canadian Shield on a regional scale.

Mobile plates or changing fields?

In spite of great structural complexity in detail, the overall stress-field of the Grenville Province and its surroundings may have been constant for periods of at least a few tens of millions of years. The same is true, of course, of Phanerozoic collision belts, where the dominant compressive stress is perpendicular to the long axis of the belt. Stress-fields of Phanerozoic deformations are caused by the displacements of plates and are thus ultimately controlled by mantle convection. This must mean that causes for the presence of the stress-field are external to the region being deformed (in this case eastern North America) and possibly also external to the plate that carries this region. When the relative orientation of the field changes, as for instance between ~ 1150 Ma and ~ 950 Ma ago, three explanations are possible: the plate is moving in a fixed stress-field, or the field moves around the plate, or both are moving at the same time. According to measurements of palaeomagnetic inclination and declination one has to choose the third of these possibilities because changes in direction of the assumed stress field do not match those of the drifting continent. Changes in stress directions have also been accompanied by changes of relative stress intensity, if only because the area was not being deformed on a regional scale prior to 1400 Ma ago, and reached tectonic quiescence again after about 800 Ma. As to the relative importance of changes in drifting speed, and that of mantle heat sources in "softening" the rocks, they can only be speculated upon.

Comparison of apparent polar wander paths and changes of latitude of the Grenville Province with variations in relative direction and intensity of the stress field suggest the following model.

Around 1350 Ma ago the eastern Canadian Shield that, according to palaeolatitudes, had been coasting north of the equator changes route and heads into the southern hemisphere. The long axis of the Grenville Province is then slightly south of east and will remain in the southeast quadrant during all the period that concerns us here. The simplest probable cause is that some "plate-jam" forces the continent to modify its route. New tectonic stresses involved in this change may be responsible for dyking and rifting during this first stage. In first approximation, the region seems to drift in a direction perpendicular to the Grenville Front. As this is also the direction of σ_3, everything is as if the region was being pulled from the southwest.

Around 1150 Ma ago the apparent drift of the pole passes the apex of the Logan loop and the region changes drift direction to move north again. It will remain close to the equator but north of it, until 1000 Ma or possibly 950 Ma ago. During its northerly drift (the descending limb of the Logan loop) the long axis of the Grenville Province remains about the same, and deformation patterns suggest that the province was pushed from the south (present east) back towards the equator. From about 1000 Ma onwards the entire region again heads into the southern hemisphere, only to drift back north once more. This is the Grenville loop of the palaeomagnetic path. During this interval σ_1 appears to be perpendicular to the long axis of the Grenville Province and to the Grenville Front. It is as if the region was now being pushed from the northeast (present northwest) and as if it encountered so much resistance that it had to move back north again, almost in the same orientation as before.

During these 500 Ma of evolution the Grenville Province rotated a number of times by a few degrees, clockwise as well as anticlockwise. The net result is a clockwise rotation by about 45°, but the episodes of deformation appear to have little relation to such pivoting motions. In fact, accurate timing of polar wander is impossible because of uncertainties in age determinations and in palaeomagnetic measurements.

A parallel evolution can be postulated for the Sveconorwegian Province in spite of some doubts about its exact position relative to the Grenville Province (Morris and Roy, 1977; Patchett et al., 1978). Its tectonic evolution is similar to that of the Grenville Province and if one accepts the reconstitution of Morris and Roy (1977) the orientation of σ_1 is the same as for the Grenville Province.

The last effect of the Sveconorwegian event was the intrusion of a dyke swarm parallel with the Schistosity Zone, but in its foreland (Fig. 14-6). These dykes have been dated at between 900 Ma and 1000 Ma ago and are thought to have formed in response to post-tectonic uplift of the Sveconorwegian block (Patchett and Bylund, 1977). It may be more realistic to assume that the necessary crustal extension was a reaction of the foreland to the accumulation of thrust-sheets coming from the west onto the craton (Fig. 14-9).

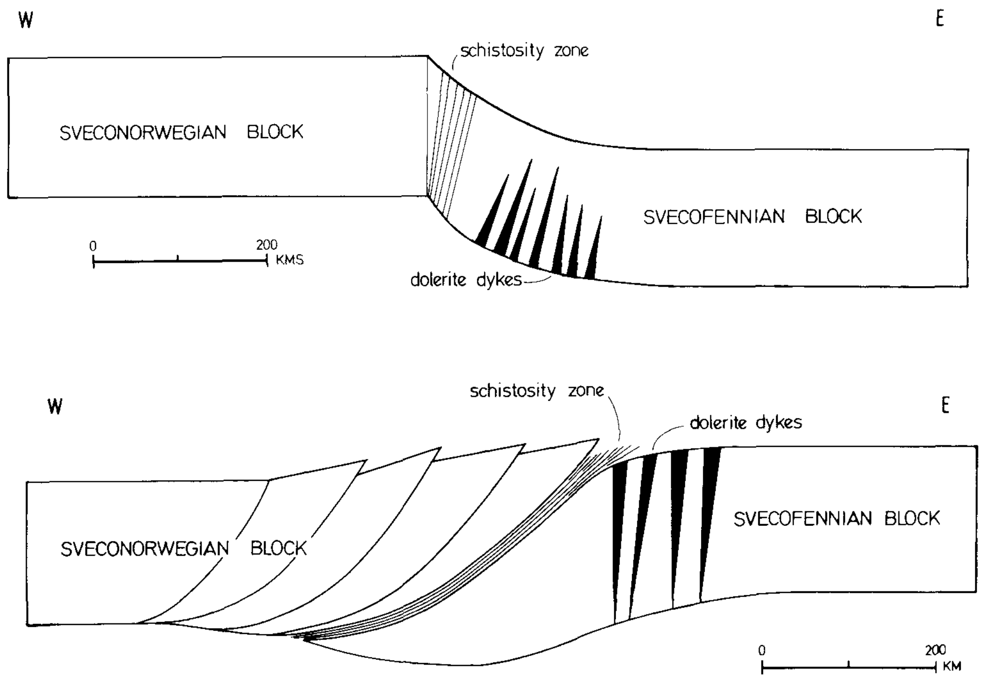

Fig. 14-9. Cartoon of tectonics postulated to explain the emplacement of dolerite dykes. Top: from Patchett and Bylund (1977); bottom: preferred interpretation. For explanation see text.

Proterozoic tectonics and a Grenvillian Tibet

The model proposes a mechanism of intra-continental (intra-plate) deformation by plate jostling or "plate-jams" that differs from the plate-edge type of deformation of Phanerozoic plate tectonics. In fact, this mechanism is hardly compatible with the presence of active subduction. Subduction prevents plate jostling because the denser oceanic plates slide under the continents and in this way they escape being jammed against each other. When continental collision occurs the two continents are locked together, but on the scale of the globe plates keep moving because other subduction zones replace those that are destroyed.

If subduction did not operate prior to about 800 Ma ago, plates would have moved in two dimensions along the surface of the earth. Unless plate motions were perfectly synchronized (in the extreme case there would only be one continuous shell) deformation would occur either along steep shears, equivalent to plate edges or through intracontinental shearing.

The recent deformation of the Tibetan Plateau and adjacent parts of Asia is particularly interesting in this regard. The Himalayas developed along the collision zone of two continents and are a typical plate-edge type of mountain belt formed after subduction of oceanic crust. The usual Phanerozoic evolution is that stresses are dissipated some time after collision, as for instance after the Ural or Appalachian collisions. In the case of Tibet something went wrong, and stresses have persisted for about 45 Ma. The reason is very probably to be found in the local plate geometry. The large oceanic plate that carries India is being subducted along the Java—Sumatra trench. As the plate is rigid, Tibet represents an obstacle that prevents its western end from moving north. North-trending stresses are therefore applied continuously to the Himalayas, Tibet and central Asia. The situation is evolving towards shearing of the ocean-floor along the Ninety-east ridge (Stein and Okal, 1978). When India becomes decoupled from the subducting lip of the plate, Tibet and central Asia should enter a period of tectonic quiescence. The situation of Tibet is anomalous in the present plate tectonics context but the consequences of this particular situation are a good analog of Proterozoic events, in spite of totally different mechanisms. Intra-continental deformation will occur whenever continents are held long enough under high directed stresses. During plate tectonics with subduction this rarely happens, and then only where a peculiar local plate geometry exists such as in the case of Tibet. In plate tectonics without subduction, however, this is common wherever drift motions lock plates together. This is the Grenvillian case and, more generally, that of most Proterozoic orogens.

The fundamental change of orogenic processes from intra-continental to peri-continental occurred around 800 Ma ago and was accompanied, if not caused, by two events. One was the beginning of subduction, due to overall cooling of the earth and associated changes in upper mantle density. The other was an acceleration of the thickening of continental lithosphere, also controlled by the cooling of the earth (Baer, 1977).

CONCLUSION

The geology of the Grenville belt is sufficiently well known to gain a general impression of its structure and its evolution, but much detailed work remains to be done. As an "orogen" or "tectogene" the belt differs fundamentally from Phanerozoic mountain-belts, but no more so than say the Churchill Province of Canada, the Kibaran or the Eburnean orogens of Africa, the West Kimberleys of Australia, or the Svecofennides of northern

Europe. These "belts" represent intra-continental (intra-plate) deformation and are often not belt-shape at all. They contrast with peri-continental (peri-plate) Phanerozoic belts. This essential difference of tectonic regime is attributed to the absence of subduction of oceanic plates prior to about 800 Ma ago.

Deformation of the Grenville belt was caused by "plate-jams" that modified the stresses applied to the lithosphere of the continent. Hairpins of palaeomagnetic apparent polar wander paths faithfully and accurately record corresponding changes in the drift pattern.

Because of this non-uniformitarian evolution, the pre-tectonic evolution of a region is not related to its tectonic history in any immediate and constant way. In a Wilson cycle, the rifted edge of a continent is all important. It controls sedimentation and dominates structural trends of the Andean- or Himalayan-type belts. By contrast, the tectonic development of the eastern Canadian Shield between ~ 1500 Ma and ~ 1200 Ma was in no way a necessary precursor to the Grenvillian events. The only continuity between pre-tectonic and tectonic evolution is of structural nature, because deformation is guided and controlled to some extent by the existing geometry of the continent.

Intra-continental and peri-continental deformation models are not entirely incompatible. Tibet and central Asia are a special case of contemporaneous plate tectonics where typical Proterozoic plate-jams are replaced by a very peculiar local plate geometry. In this sense Tibet and the Grenville Province are comparable, but ultimate causes of deformation need not be similar simply because final products are similar. Comparative anatomy of orogens must take convergence into account.

Integrated structural, geochronological and palaeomagnetic studies and more accurate determinations of the ages of rocks and of their magnetisation will be essential to improve our understanding of the Grenville belt.

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Chapter 15

PAN-AFRICAN (UPPER PROTEROZOIC) PLATE TECTONICS OF THE ARABIAN—NUBIAN SHIELD

I. G. GASS

ABSTRACT

Field, geochemical and structural evidence collectively indicate that the Upper Proterozoic to Lower Palaeozoic (Pan-African) continental crust of the Arabian—Nubian Shield evolved through a period of about 700 Ma (1200—500 Ma) by progressive cratonization (continentalization) of numerous intra-oceanic island arcs. An episodic continuum of magmatic, metamorphic and sedimentary processes, similar to those operating above present-day oceanic and then continental subduction zones, is envisaged. Structural deformation and ophiolite obduction occurred primarily during phases of arc collision at c. 1000, 800 and 600 Ma. Although evidence presented relates to the Arabian—Nubian Shield, the cratonized island-arc model seems to be applicable in Africa as far south as Ethiopia in the east and across northern Africa as far as the West African Archaean craton. If this is the case then c. 5×10^8 km^3 of new continental crust was produced in the Upper Proterozoic by plate-tectonic processes directly analagous to those of present-day destructive plate margins.

INTRODUCTION

That part of the earth's continental crust described and discussed here is of Upper Proterozoic to Lower Palaeozoic (c. 1200—500 Ma) age and forms the crystalline basement to Phanerozoic sedimentary sequences in north-eastern Africa and Arabia. The regional extent of the area is shown in Fig. 15-1 and the extensive tracts of basement in western Saudi Arabia, the Egyptian Eastern Desert and the northeastern Sudan are collectively termed the Arabian—Nubian Shield.

African Precambrian consists of three major and several minor Archaean cratons surrounded and separated by non-Archaean Precambrian terrain (Clifford, 1970). In 1964 Kennedy identified the dominance of 450—650 Ma K-Ar ages for the non-cratonic Upper Precambrian and proposed the term "Pan-African tectono-thermal event" to identify this major episode in African geological evolution. Since then, the term "Pan-African" has been used to describe much of the non-Archaean African basement although the time span has been extended by Rb-Sr and zircon dating to 1200—450 Ma (e.g. Fleck et al., 1976; Hashad, 1980). There are therefore two views on the age range of the Pan-African. Some restrict the term to rocks of 450—650 Ma.

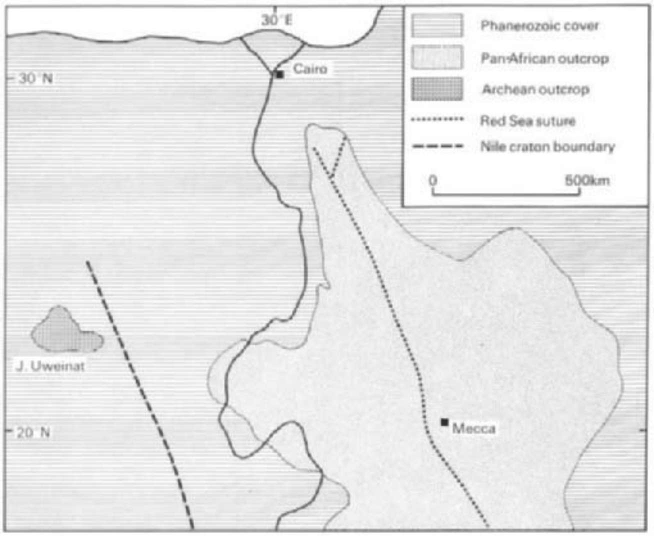

Fig. 15-1. Sketch map indicating the area of the Arabian—Nubian Shield. The existence of an Archaean to Lower Proterozoic craton to the west of the Nile seems likely on the basis of radiometric dates. The eastern boundary of the craton is very approximate and the Red Sea has been closed to its pre-drift position.

Others, of whom I am one, feel that as the longer 1200—450 Ma period is based on isotopic data not available in 1964, that its usage for the Arabian—Nubian Shield is more in keeping with Kennedy's original concept. So, the extended 1200—450 Ma Pan-African age range is used here.

Although the Pan-African basement has all the geological and geophysical characteristics of continental crust, Kennedy (1964) could see few indications of classic orogenesis and hence used the term "tectono-thermal event" to keep the interpretive options open. Since then, and largely in the last five years, two schools of thought concerning the origin of the Pan-African of the Arabian—Nubian Shield have developed. The first maintains that this part of the Pan-African is essentially Archaean crust that was thermally and tectonically reworked during Late Proterozoic (1200—500 Ma) Pan-African events.

The alternative hypothesis is a plate-tectonic model suggesting that the continental crust of the Arabian—Nubian Shield evolved entirely within the Upper Proterozoic. The envisaged processes started some 1200 Ma ago with numerous immature intra-oceanic island arcs forming as subduction occurred between converging plates of oceanic lithosphere. The island arcs, the proto-continents, evolved by repeated magmatic, metamorphic and tectonic events associated with concomitant erosion and sedimentation. Fragments of back-arc oceanic lithosphere were obducted and ophiolite zones marked the approximate site of plate margins as arcs collided to form larger "continental" masses. Finally, when subduction ceased, about 500 Ma ago, the whole region had developed a continental character. This model, based on field, geochemical and isotopic data, envisages and episodic continuum of magmatic, metamorphic and sedimentary processes while subduction continued; tectonism and ophiolite emplacement occurred primarily when arcs collided.

The absence of older radiometric dates ($>$ 1200 Ma), the generally low $(^{87}Sr/^{86}Sr)_i$ ratios (0.702—0.706), the suprabundance of volcanoclastic sequences and related cannibalistic sediments, the ubiquitous presence of calc-alkaline magmatic products and the identification of several ophiolites in linear zones of mafic—ultramafic complexes leaves little doubt that the "cratonized island arcs" hypothesis is substantially correct. In my opinion (Gass, 1979) no significant reason remains to preclude acceptance of the arc model, originally proposed by Greenwood et al. (1976), for the Pan-African of the Arabian—Nubian Shield. What does remain, however, is to understand more fully the multitude of complex processes that accompanied and/or caused the conversion of oceanic lithosphere into continental crust. In this context it is relevant to note that when Pan-African heat production is compared to that of the present day, that of the Pan-African (1200—500 Ma) would be 1.4—1.7 times present-day values (Brown, 1980). These figures suggest that although the slightly steeper thermal gradients and lower upper-mantle viscosities of the Upper Proterozoic may have produced thinner lithospheric plates, steeper subduction zones and narrower arc systems, but arc dimensions would, theoretically, be within 10% of present-day values. What follows is then a markedly uniformitarian approach of looking at present-day arc systems and comparing their features to those of the Arabian—Nubian Shield.

Present-day arc systems are narrow, rarely exceeding 100—150 km from trench to back-arc basin, with active volcanism (and presumably plutonism) usually confined to axial zones less than 50 km wide. It would therefore be quite possible to fit ten or more arc systems into the 1500 km N—S or E—W extent of the Arabian—Nubian Shield. And, during the 700 Ma span of the extended Pan-African, there could well have been numerous subduction zones widely distributed in space and time. Repeatedly, the products of an earlier phase must have been invaded and/or blanketed by those of later episodes. In this way, similar rock types and associations would be produced at about the same time in arc systems that were then widely separated. Conversely, similar rock types and associations would be produced in the same arc system at different times. With this temporal and spatial community of character, long-range stratigraphic correlations are hazardous and radiometric data on magmatic events cannot be given regional connotation. No detailed evolutionary picture of an individual arc has emerged although work, primarily in Saudi Arabia, has allowed a timetable of stratigraphic, tectonic and magmatic events to be erected (Fitch, 1978, tables 1 and 2) that can also be applied in Egypt and the northeastern Sudan. In Table 15-I a grossly simplified version of this stratigraphic table is presented to provide a time

TABLE 15-I

Rock types of the Arabian—Nubian Shield (Largely after Fitch, 1978, tables 1 and 2 and Brown and Jackson, 1979)

Age (Ma)	Rock types			Inferred tectonic setting	Comments and other data
	Plutonic	Volcanic	Sedimentary		
Post Pan-African	Alkaline and peralkaline granites characterized by high Ti, Zr, Nb, U, Th	Alkaline and peralkaline trachytes and rhyolites	Terriginous arkoses and shallow water shales	Continental	Continental character of region established; all magmatism of "within-plate" variety
500—600	*Diachronous end of destructive margin process*				
Upper Pan-African	Calc-alkaline granites and granodiorites with low Ti, Zr, U, Th and very low Nb	Rhyolites, dacites, trachytes and andesites	Conglomeratic and arenaceous units with granitic and rhyolitic clasts. Stromatolitic limestones	Continental with margins of Andean type	Regionally extensive, unmetamorphosed and structurally undeformed silicic volcanic and plutonic rocks.
600—670	*Major break, regional unconformity, change in composition of magmatism*				
Middle Pan-African	Calc-alkaline diorites and granodiorites	Calc-alkaline andesites and basaltic andesites with subordinate rhyolitic and dacitic units	Greywackes and minor arkoses. Stromatolitic limestones and shallow water shales	Numerous mature intra-oceanic island arcs. Major stratigraphic and regional breaks suggest complex evolution of several arcs	c. 600 Ma emplacement of ophiolitic complexes c. 800 Ma emplacement of ophiolitic complexes Complexely deformed and metamorphosed to green-schist facies c. 1000 Ma emplacement of ophiolitic complexes
c.1000	*Distinct structural, compositional and metamorphic break: arc collision (orogenesis at c. 960 Ma)*				
Lower Pan-African	Gabbros diorites, granodiorites	Low-K basalts and basaltic andesites	Immature greywackes, cherts, shales, occasional limestones	Numerous immature intra-oceanic island arcs	Sparce and highly deformed outcrops. Metamorphosed mainly to amphibolite facies
1200?					

framework for this assessment of Upper Proterozoic plate-tectonic processes.

As shown on Table 15-I the Pan-African of the Arabian—Nubian Shield can be divided into three major divisions; these are here termed the Lower, Middle and Upper Pan-African, respectively. The oldest rocks in the Lower Pan-African 1000—1200 Ma age range are highly deformed and metamorphosed but are comparable in composition to those from present-day immature intra-oceanic island arcs. Then, following a distinct stratigraphic, tectonic and compositional break at about 1000 Ma, the 400 million years between 1000 and 600 Ma saw the development during the Middle Pan-African of numerous maturing intra-oceanic island arcs. That these arcs collided at various times is indicated by the emplacement of ophiolite complexes at 1000, 800 and 600 Ma and also by several phases of compressional deformation. By 600 Ma it seems that most of the Middle Pan-African arcs had coalesced into a unified continental mass. But the continuation of calc-alkaline magmatism during the 600—500 Ma Upper Pan-African indicates the presence of a subduction zone beneath at least part of the region. Finally, at or about 500 Ma ago, alkaline and peralkaline magmatism with clear within-plate geochemical characteristics had replaced the previous calc-alkaline associations throughout the entire region. This is taken as indicating the attainment of true continental character, the end of subduction zone—destructive margin processes and the end of the Pan-African phase of continental evolution. The field, petrographic and geochemical characteristics and the plate-tectonic significance of the three divisions of the Pan-African and the post-Pan-African magmatic products will be briefly reviewed before a regional tectonic and petrogenetic appraisal is attempted. Because a comparison will be made between Pan-African magmatic products and those produced at or above ocean—ocean and ocean—continent subduction zones in the present plate-tectonic cycle, it is appropriate to identify the primary geochemical characteristics of "orogenic" (destructive margin—subduction zone) magmatic products.

Orogenic associations are characterized by volcanic rocks varying in chemical composition from basic to acid. The *island-arc tholeiite association* of immature island arcs is dominated by basaltic rocks with tholeiitic characteristics, (e.g. Fe-enrichment in an AFM diagram) whereas in the *calc-alkaline association* of more mature island arcs and continental margins, andesites and dacites and rhyolites are the major rock types. Calc-alkaline andesites are oversaturated intermediate rocks with 55—63% SiO_2, relatively high Al_2O_3 (c. 15—19%) moderate alkalis ($Na_2O + K_2O$ = 4—7%) and calcium contents (CaO = c. 5—6%) and showing no Fe-enrichment trend in an AFM diagram. In addition to these major element characteristics, orogenic associations have distinctive trace element abundances such as low Nb, Y, Zr, U and Th compared with other volcanic associations (Pearce and Gale, 1977; Pearce and Norry, 1979). The plutonic calc-alkaline association is of gabbro-tonalite-granodiorite-granite in which tonalites and granodiorites are dominant. Such intermediate calc-alkaline associations can be readily distinguished geochemically from anorogenic (within-plate) alkaline and peralkaline granitic associations. All magmatic rock types, except those produced by post Pan-African within-plate activity, are "orogenic". Also, it is assumed, because no detailed studies have proved otherwise, that the Pan-African volcanic rocks are the surface expression of the same processes that produced the plutonic masses.

FIELD, PETROGRAPHIC AND GEOCHEMICAL DATA

In working upwards through progressively younger Pan-African lithostratigraphic units, the wide variety of local formation and group names will be avoided as far as possible and attention concentrated on describing the main rock types and associations, their composition, age, tectonic setting and plate-tectonic implications. But, just what is the base of the Pan-African succession? In southwestern Saudi Arabia, Egypt and the Sudan, thick sequences of siliceous gneisses derived from sedimentary quartzites occur beneath metavolcanics with "arc" affinities. These quartzites are thought to represent a passive margin sedimentary wedge flanking the Archaean to Lower Proterozoic Nile craton (see Fig. 15-1). So far, no older (i.e. $>$ 1200 Ma old) basement has been positively identified northeast of this zone and all rock types formed in the following 600—700 Ma have calc-alkaline affinities and are believed to have formed over Upper Proterozoic ocean crust.

Lower Pan-African

There is little direct geochronological evidence to support the proposal (see Table 15-I) that the Lower Pan-African had a time span of 200 Ma between 1000—1200 Ma ago. Indeed, opinions differ as to whether these sequences represent the oldest formations or are facies equivalents of younger ones that have been more intensely deformed and metamorphosed than elsewhere (Schmidt et al., 1973; Greenwood et al., 1976; Fitch, 1978; Brown and Jackson, 1979). There are few radiometric dates that fall unequivocally in this time span and there is no precise isotopic control over the proposed age range. That the Lower Pan-African exists as a separate entity is best deduced from field and compositional evidence. Brown and Jackson (1979), summarizing field evidence from the southern part of the Arabian Shield, state categorically that there is a distinct structural, compositional and metamorphic break between overlying more siliceous, less metamorphosed and relatively undeformed formations and the oldest sequences here allotted to the Lower Pan-African.

Thick sequences ($>$ 12,000 m) of basalts and basaltic andesites with intraformational basic greywackes and subordinate carbonate and cherts form the main surface rock types in this division. These were invaded by c. 1000 Ma old composite gabbroic-dioritic batholiths with $(^{87}Sr/^{86}Sr)_i$ ratios of 0.7029 (*according to Fleck et al., 1979, the recalculated ages are 818—938 Ma with initial ratios of 0.7023—0.7030, ed.*). In the sediments there are no clasts of K-feldspar and no sign of terrigenous input, and Greenwood et al. (1976) envisage an immature, oceanic island arc with adjacent depositional basins as the most likely tectonic setting. Higher up the sequence sediments replace volcanics as the dominant rock type, and as there are few compositional differences between them it is likely that the sediments were derived by the rapid erosion of an adjacent, now underlying, volcanic arc. In support of the immature arc setting Greenwood et al. (1976) and Greenwood and Brown (1973) quoted the limited amount of analytical data then available (e.g. Jackaman, 1972) which indicated that the basic volcanics of this lowest group are compositionally similar to the present-day, immature island-arc tholeiites of Jakês and Gill (1970). Since that time samples of Lower Pan-African diorites-granodiorites have been analyzed for major, trace and rare earth elements (Nasseef and Gass, 1977; Gass, 1977). In Fig. 15-2, a $TiO_2:Zr$ plot regarded by Pearce and Norry (1979) and Pearce (1980) as the most realistic geochemical discriminator between arc and within-plate magmatic products, all Lower Pan-African specimens so far analyzed plot in the arc field. Although the fields on Fig. 15-2 have been defined on present-day extrusive rocks and many of the Pan-African specimens are plutonic and have been metamorphosed to the amphibolite facies, the compositional variation along the fractionation trend due to crystal accumulation and related processes would still keep them within the arc field. Metamorphic processes at these facies should not affect the abundance of either TiO_2 or Zr.

So, although the evidence is circumstantial, it is sufficient to suggest that the oldest Pan-African rocks are products of immature island arcs. There is no evidence yet on the number of arcs or where they (it) were on the surface of the Upper Proterozoic earth.

Middle Pan-African

Between the Lower Pan-African just described and the Upper Pan-African which began at or about 600—670 Ma, fall the bulk of rocks forming the Arabian—Nubian Shield. These are here allotted to the Middle Pan-African and despite the great number of identifiable individual formations, plutonic complexes and tectonic events, the entire division can be simply described. Plutonic rocks of dioritic-granodioritic-granitic composition form between 50 and 60% of the outcrop and have been emplaced into a host of andesitic-dacitic-rhyolitic eruptives and cannibalistic sediments derived therefrom. Stromatolitic limestones and chert horizons are widespread but quantitatively

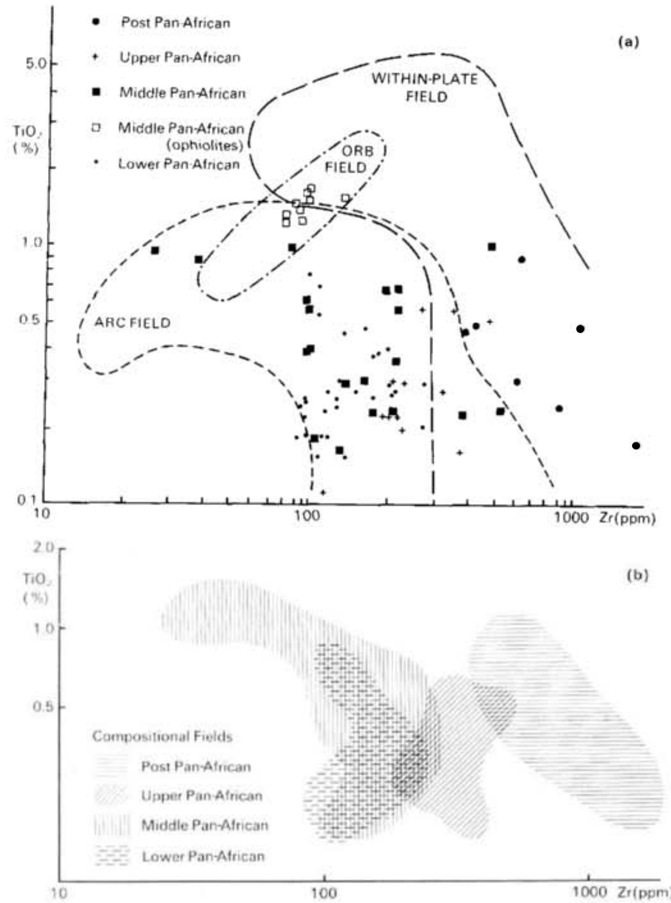

Fig. 15-2. a. Plot of TiO_2 against Zr for Pan-African igneous rocks. Compositional field for present-day volcanic rocks from island arc, ocean ridge basalts (ORB) and within-plate settings are after Pearce (1980). b. Graph depicts the variation of Ti and Zr abundance in Pan-African igneous rocks with time. Note the progressive increase in Zr whereas Ti remains relatively constant.

minor components. Gradually, the composition of the volcanic host became more siliceous, changing from andesitic through dacitic to rhyolitic, Volcanic types range from subaerial lavas, welded tuffs, breccias and agglomerates to water-lain ashes. There is rapid lateral and vertical variation in rock type which is characteristic of strato-volcanoes. Both Delfour (1975) and Greenwood et al. (1976) have suggested that these rocks were formed on the flanks of emergent volcanic island arcs. The sediments derived from these volcanic edifices range from fluviatile conglomerates to finer-grained arenaceous and argillaceous deposits laid down in shallow marine conditions. Commonly, the original sediments were reworked by turbidity currents and, in places, more than 10 000 m of sediments were deposited in subsiding basins.

This simplified description takes no account of local complexities. Rapid lateral and vertical facies change in sediments and the impersistence of compositional and textural volcanic rock types laid down under both subaerial and shallow marine conditions, provide ample criteria for erecting geological formations. Because local changes are many and complex, they often overshadow more significant regional variations. Nevertheless regional variations show through. For instance, in Saudi Arabia, there seems to be a regional N—S lateral variation from a northern, dominantly sedimentary zone, through a dominantly volcanic zone to a southern zone where sediments again predominate. Similarly, in the northeastern Sudan Gass (1955) and Neary et al. (1976) record that tracts of dominantly volcanic rocks give way laterally to, and interdigitate with, mainly sedimentary sequences of virtually identical composition. Generally, deposition seems to have been reworked intermediate and acidic volcanoclastics in shallow elongate basins. The environment was undoubtedly that of active volcanic arcs emerging from shallow peripheral seas.

Invading the Middle Pan-African volcanic—sedimentary sequences are numerous syn- and post-kinematic diapiric plutons of dioritic and granodioritic composition. As with the volcanics, these plutonic rocks tend to become more siliceous with decreasing age. Originally, it was thought that these plutons were emplaced during and after well-defined orogenic phases. Recently, however, it has been shown that in the southern Arabian Shield at least (Cooper et al., 1979) there was essentially an episodic continuum of plutonism. In Fig. 15-3, taken from the works of Cooper et al. (1979), Hashad (1980) and Fleck et al. (1976), the ages of various plutonic masses are plotted. The regularity of events throughout the 660—820 Ma period is evident. With relatively few reliable radiometric dates available, it seems more likely that the gaps in Fig. 15-3 are due to lack of isotopic data rather than significant pauses in plutonism.

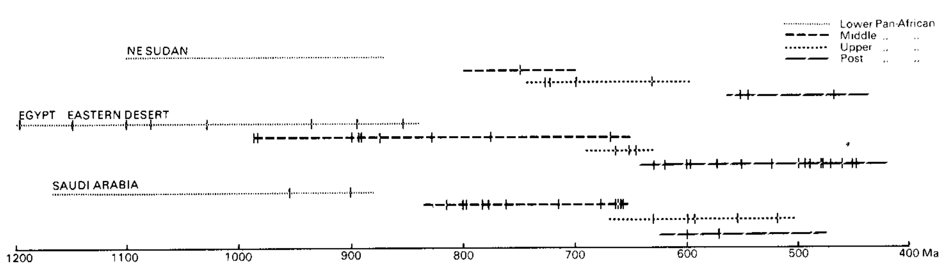

Fig. 15-3. Schematic representation of the age ranges of the Lower, Middle and Upper Pan-African and post-Pan-African activity in the Sudan, Egypt and Saudi Arabia. Vertical ticks are Rb-Sr whole-rock isochron dates determined mainly for plutonic bodies. Note the continuum of activity and the seemingly diachronous change from one division to another.

Another feature of Middle Pan-African magmatism is the spatial distribution of the plutons. Although initial impression on studying regional geological maps is of a random distribution, closer inspection suggests that many plutonic bodies lie in linear zones. Particularly clear examples of such zones are shown in Fig. 15-4a, b. In the southern Arabian Shield (Fig. 15-4a; Ramsay et al., 1979) and in the northeastern Sudan (Fig. 15-4b; Neary et al., 1976) there are N—S and NE—SW trending zones about 50 km wide where granitic masses are markedly more abundant than elsewhere.

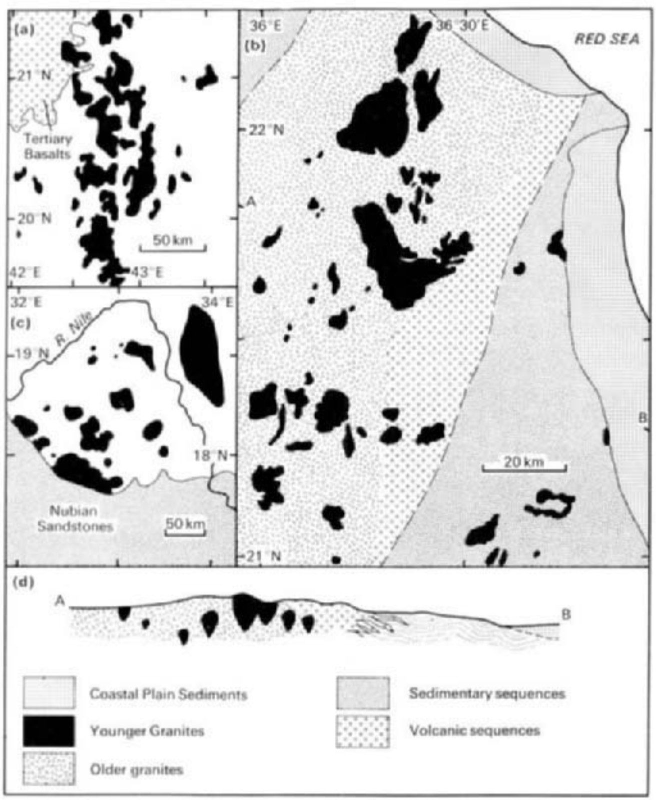

Fig. 15-4. a, b and c. Geological sketch maps showing granitic plutons in parts of Saudi Arabia and Sudan. Figures a and b depict linear Pan-African trends. Figure c' is a WNW—ESE section across Figure b showing the postulated relationship of plutons to volcano-clastic and sedimentary sequences.

Ramsay et al. (1979), in discussing the granitic zones of the Arabian Shield and other Middle Pan-African zones characterized by metamorphic grade or sedimentary/volcanic type, emphasize that although boundaries are approximate and arbitrary, the zones do represent discrete identifiable geological entities. Just what these zones or provinces represent in terms of geological processes is less clear — rifted sedimentary basins separated by Andean-type magmatic zones, and eroded ensimatic island arcs with granitic plutons flanked sequentially by volcanic rocks and reworked volcanoclastic sediments have been proposed. Here, the last explanation is preferred and an example of this relationship occurs in the northeastern Sudan (see Fig. 15-4b, c).

The Middle Pan-African was a period of numerous maturing ensimatic arc systems subsequently swept together by plate-tectonic processes to form, by the end of the division, larger "protocontinental" masses. As arcs matured through geological time the magmatic products became progressively more siliceous and $(^{87}Sr/^{86}Sr)_i$ ratios ripened from 0.7028 to 0.7035 (*according to Fleck et al., 1979, the initial ratios of the 660—820 Ma old intrusives vary between 0.7025 ± 5 and 0.7035 ± 7, ed.*), although all are calc-alkaline and on both a $Nb:SiO_2$ (Gass, 1979) and on the $TiO_2:Zr$ plot (Fig. 15-2) clearly fall in the volcanic arc magma field of Pearce (1980). In this model the original arcs were preserved whilst the intervening oceanic lithosphere was subducted. Occasionally, however, fragments of oceanic lithosphere were caught up between colliding arcs and these are preserved as ultramafic—mafic complexes of ophiolitic character (as defined by the 1972 Penrose Ophiolite Conference) that occur in zones across the region (Bakor et al., 1976; Gass, 1977; Frisch and Al-Shanti, 1977; Rehaile and Warden, 1978; Shanti and Roobol, 1979). So far about ten such ophiolite masses have been identified in Saudi Arabia, a further six to eight in the Egyptian Eastern Desert and two to three in the northeastern Sudan. Geochemical studies indicate that some of these bodies are probably fragments of oceanic lithosphere originally formed beneath back-arc marginal seas (Bakor et al., 1976). Although no reliable radiometric dates have yet been published on these ophiolite masses, structural relationships with dated formations suggest that there are at least three emplacement ages at c. 1000, 800 and 600 Ma (see Table 15-I) and therefore three phases of arc collision.

Upper Pan-African

On Table 15-I the time span of the Upper Pan-African is given as 670—600 to 600—500 Ma; a diachronous period up to 170 Ma long. The base of the division is readily recognizable as unmetamorphosed and relatively undeformed silicic volcanics and related sediments which overlie older formations with marked angular unconformity. A comglomerate of regional extent is usually the basal unit and stromatolitic limestones occur as minor components with interbedded high-energy fluviatile and shallow-water clastic sediments. Both syn- and post-kinematic calc-alkaline granites and granodiorites have been emplaced into the volcanic—sedimentary host.

Extensive tracts of these volcanic rocks occur in the northern part of the Arabian Shield, the Egyptian Eastern Desert and the northeastern Sudan where they are commonly preserved in down-faulted blocks. They seem to be absent in the south where the level of erosion is deeper. Rock types are mainly subaerial pyroclastics with ash-fall and flow units much more abundant than lavas. Extensive welded tuff units are particularly conspicuous and their unmetamorphosed character enables multiple eruptive components of a single cooling unit to be identified. Rock types present are mainly rhyolites and rhyodacites with subordinate andesitic flows.

Plutonic rocks of the Upper Pan-African are mainly calc-alkaline granites or granodiorites. In a detailed study of a small area in western Saudi Arabia Nasseef (1971) and Nasseef and Gass (1977) were able to identify three syn-kinematic and one post-kinematic phases falling in the range 610—525 Ma. Of the four magmatic episodes the last, post-kinematic, phase dated at 525 ± 20 Ma was the most voluminous. These findings are in general agreement with earlier and more extensive investigations by Fleck et al. (1976) based on K-Ar mineral studies. In the southern part of the Arabian Shield Fleck and his co-workers identified a major magmatic phase between 610 and 510 Ma with a major pulse between 610 and 540 Ma and minor episodes at 535 and 510 Ma. (*According to Fleck et al., 1979, recalculations of these data and new measurements change the duration of this phase to 610—650 Ma with the main pulse at 620—645 Ma ago, ed.*). For the Egyptian Eastern Desert Hashad (1980) reports an age range of 675—450 Ma for the Younger or Red Granites with maximum activity between 675 and 500 Ma, peaking at 600 Ma ago. A similar 600 Ma maximum is reported for the northeastern Sudan by Neary et al. (1976) (see Fig. 15-3).

Within the Upper Pan-African $(^{87}Sr/^{86}Sr)_i$ ratios range from 0.7032 to 0.7093. When plotted on a lime/alkaline index most of the Upper Pan-African rocks are calc-alkaline although only marginally so; some on Al_2O_3: $K_2O + Na_2O$ content are marginally peralkaline. However, the trace element abundance (e.g. $Nb:SiO_2$ and $Ti:Zr$ plots) has distinct characteristics of arc magmatism. This indicates that one or more subduction zones were active under the region, although the lack of metamorphism and deformation, other than block faulting and gentle folding, suggests that a well-developed, coherent and rigid sialic crust had been developed by this time. This contention is supported by the abundance of high-energy terrestrial and shallow-water arkosic sediments indicating a high-standing continental mass.

Just when the division ends presents problems for, in many respects, the region already had a continental character though with active subjacent destructive margin or margins. What can be deduced geochemically is when the subduction stopped. This change in tectonic setting is marked by the diachronous but well-defined order of magnitude increase in magmatic products of Nb, Y, Zr, U and Th. Although not strictly accurate, it is convenient to use the terms calc-alkaline and peralkaline in identifying this change in chemical composition. Pearce and Gale (1977) suggest that the

Nb content of peralkaline granites is high because the mantle source in such "within-plate" settings has been enriched in this element by migrating CO_2-rich fluids or interstitial melts. They argue that since Nb is not a hydrophyllic element, and since it is likely to be compatible with residual mineral assemblages in subducted oceanic crust, the enrichment of Nb is unlikely to occur in normal arc magmas. The incoming of peralkaline magmas that are taken to mark the end of the Pan-African cover a wide time span. In the extreme north of the Arabian Shield, adjacent to the Gulf of Aqaba, two peralkaline granites have been dated at 600 ± 24 Ma (Stoesser and Elliott, 1979) and 591 ± 8 Ma (Baubron et al., 1976; recalculated with new decay constant, ed.) respectively, whereas farther south in Saudi Arabia, east of Jeddah, the calc-alkaline Taif granites are dated at 514 ± 20 Ma (Nasseef and Gass, 1977; recalculated with new decay constant, ed.). In the northeastern Sudan two markedly peralkaline masses date at 500 Ma (Neary et al., 1976) whereas farther north in Egypt Hashad (1980) records that calc-alkaline activity continued until 500 Ma ago but by 450 Ma peralkaline magmatism was well established. The obvious interpretation of these data is that subduction beneath the region stopped earlier in some areas than in others. But whether this was along one zone or whether there were several zones that stopped at various times is as yet unknown. It is, however, pertinent to note that once established, peralkaline magmatism occurred at widely spaced intervals throughout the Phanerozoic.

PLATE-TECTONIC CONSIDERATIONS

Progressive cratonization of island arcs by processes analagous to those operating above present-day subduction zones is now widely accepted as the best hypothesis for the evolution of the Arabian—Nubian Shield. Within this generalized framework several attempts have been made to identify individual Pan-African arc systems on criteria such as zones of mafic—ultramafic (ophiolite) masses (Bakor et al., 1976; Gass, 1977 and 1979; Frisch and Al Shanti, 1977), regional variations in geochemistry (Greenwood and Brown, 1973; Gass, 1977; Gass and Nasseef, in press), zones of distinctive metamorphic and/or lithological character (Ramsay et al., 1979) and zones of characteristic mineralization (Al-Shanti and Roobol, 1979). The identification of arcs by these criteria alone is of dubious validity. For instance, although several mafic—ultramafic masses have been confidently identified as ophiolites as defined by the 1972 Penrose Ophiolite Conference (Bakor et al., 1976; Shanti and Roobol, 1979) and display back-arc ocean-floor geochemistry, it is by no means certain that these "ophiolite" zones accurately mark the sutures between arc systems. In the present SW Pacific arcs, ophiolites are preferentially emplaced along back-arc margins, but most Tethyan ophiolites are markedly allochthonous and have been moved tens, if not hundreds of kilometres during obduction. All Pan-African ophiolites so far identified have tectonic contacts and Shackleton et al. (1980) believe that the numerous mafic—ultramafic masses of the Egyptian Eastern Desert have been moved so far that the identification of linear ophiolite zones or sutures in that area is unjustified. Nevertheless, in Arabia the sutures may be more trustworthy and their position, tentatively marking arc margins, is shown on Fig. 15-5. Similarly, recent work on modern arc systems reveals that the supposedly simple relation between chemistry and depth to the subduction zone (> K, > Sr, < Rb: cf. Hutchinson, 1975) is not so simple. Variations along arcs are proving to be as great as, if not greater than, those across them. So, attempts to identify Pan-African arc systems from compositional variation must be distinctly suspect. Indeed, the variation of subduction zone models proposed, single easterly dipping (Greenwood and Brown, 1973), multiple easterly dipping (Gass, 1977), single westerly dipping (Schmidt et al., 1978), proves this point and emphasizes that at the present time there are insufficient data to justify this approach.

However, a simple arc system that has been eroded to near sea-level should have a plutonic core invading a zone of eruptive products that are in turn flanked by cannibalistic and shallow-water sediments. Of these features the plutonic core is the most likely to be preserved. Study of existing geological maps suggests that several linear granitic zones of appropriate dimensions exist in both Africa and Arabia. Traces of these zones, thought to mark arc axes, are plotted on Fig. 15-5, the inset rectangles in this figure identify two granitic zones shown in more detail in Fig. 15-4a, b. These granite zones seem to be best developed in the northeastern Sudan but elsewhere the extensive cover of Recent sediments and Tertiary volcanics leave major gaps, and subsequent deformation, such as that due to the Nadj Fault System in Saudi Arabia, present complications. Also, geochronological control is poor and the age of the granite zones is not known. It could be anticipated that such zones of magmatic activity would, once established, focus subsequent plutonism. So, plutons of varying ages could well be channelled along the same "hot" axial zones. Although it is premature to place too much reliance on these granitic zones, their abundance does reflect the complexity of the Pan-African arc system and their dominantly N—S orientation coincides with that of many ophiolite zones and, broadly, with the structural trend lines also shown on Fig. 15-5.

In Fig. 15-6 the evolution through time of the arc systems from immature to mature arcs and finally to continental conditions is depicted in cartoon form. The broad stages in evolution are well documented but the passage from Lower to Middle Pan-African is temporally vague and that between Middle and Upper Pan-African is diachronous. Similarly, the final stages in cratonization, marked by the switch from calc-alkaline to peralkaline magmatism, occurred earlier in some areas than in others. The line drawn on Fig. 15-5 is that of Stoesser and Elliot (1979) and separates 500—600 Ma Arabian peralkaline (to the east) from calc-alkaline products (to the west).

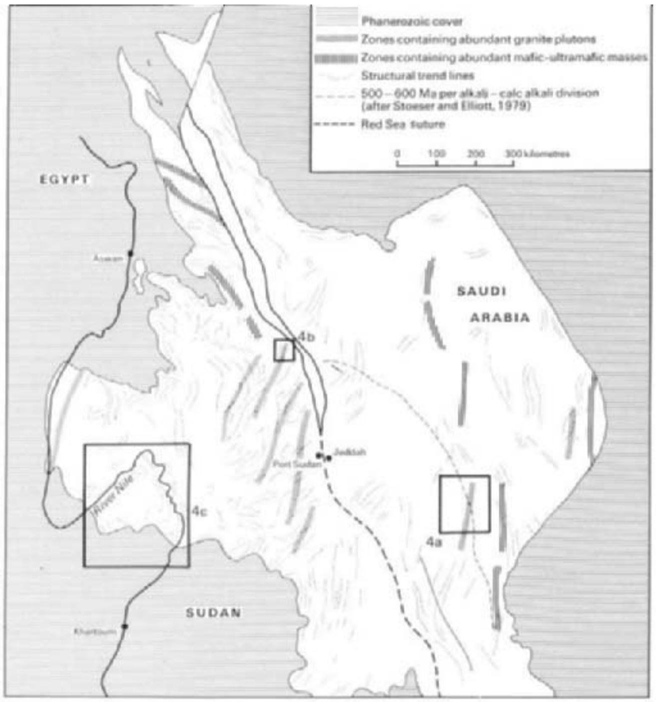

Fig. 15-5. Regional sketch map showing the disposition of mafic—ultramafic complexes (marking the approximate position of arc sutures), linear granitic zones (possible arc axes) and basement structural trends. The Red Sea has been closed to a Pre-Miocene position.

Seemingly subduction was still active west of this line whereas it had ceased to the north and east. There is no evidence as to which way this last subduction zone was inclined.

So far the data presented have come from northeastern Africa and Arabia. But can the arc cratonization model be applied to the Pan-African or Upper Proterozoic elsewhere? It is evident from the literature that in West Africa the Upper Proterozoic along the western margin of the West African Archaean to Lower Proterozoic craton has arc characteristics in parts (*see also Caby et al., this volume, Chapter 16, ed.*), and genetic models presented by others for the western Hoggar (Caby, 1970; Bertrand and Caby, 1978; Caby and Leblanc, 1973), Mali (Black et al., 1979), southern Morocco (Leblanc, 1976; *see also this volume, Chapter 17, ed.*), Tibesti (Ghuma and Rogers, 1978; Pegram et al., 1976) and Nigeria (McCurry and Wright, 1977; McCurry, 1976) have been, or can be, interpreted on an island-arc cratonization-accretion model such as that presented here. Indeed, the indications are that this model is also applicable to western and northern Ethiopia (Gilboy, 1970; Chater, 1971; Kazmin, 1976; De Wit and Aguma, 1977, De Wit and

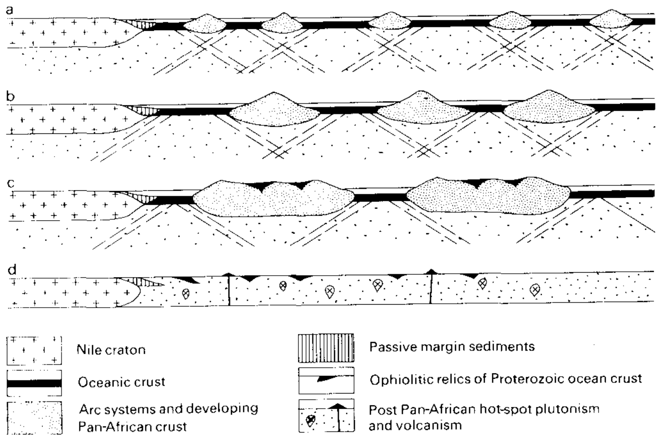

Fig. 15-6. A cartoon depicting stages in the development of the Arabian—Nubian Shield. (a) depicts the situation in the Lower Pan-African with many immature arc systems. By Middle Pan-African times (b) the arcs have matured and coalesced but have not attained continental dimensions. By Upper Pan-African times (c) the arcs have coalesced into continents but these still overlay subduction zones and magmatic activity had calc-alkaline affinity. Figure (d) depicts the post Pan-African (500—600 Ma) situation. When the continent was fully developed, subduction had ceased and magmatism was per-alkaline and of within-plate affinity.

Chewaka, 1978). Elsewhere in Africa the situation is less clear. The Mozambique and Damara belts could be ensialic features or, for the Mozambique belt, Proterozoic continent—continent plate collisions seems more likely (Shackleton, 1979; Kröner, 1979). Nevertheless, despite these uncertainties, the inescapable conclusion, based on Arabian—Nubian Shield data, is that plate tectonics and destructive margin processes essentially similar to that of the present day, operated between 1200—500 Ma ago in northern Africa—Arabia and were probably responsible for the formation of c. 5×10^8 km^3 of continental crust in the Upper Proterozoic.

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Chapter 16

PAN-AFRICAN OCEAN CLOSURE AND CONTINENTAL COLLISION IN THE HOGGAR—IFORAS SEGMENT, CENTRAL SAHARA

R. CABY, J. M. L. BERTRAND and R. BLACK

ABSTRACT

The late Precambrian to early Palaeozoic tectonic development of the Hoggar—Iforas region in the central Sahara is interpreted in terms of a complete Wilson cycle with ocean opening and closing between the West African craton and the Touareg shield.

Rifting along the eastern margin of the West African craton occurred around 800 Ma ago with a triple point in Mali, the Gourma being interpreted as an aulacogen. Continental fragmentation was accompanied by the injection, at the base of the crust and at high levels, of basic and ultrabasic magmas. The presence of basalts of possible oceanic origin, island arc and marginal trough volcano-clastic assemblages, widespread calc-alkaline plutonism, and paired metamorphic belts, including high-pressure eclogitic schists, strongly suggest that active subduction processes were at work. In the western part of the Touareg shield (Pharusian belt) widespread, prolonged continental Cordilleran conditions prevailed, the oldest being dated at 885 Ma. Three major tectonic events have been distinguished: (1) stabilization of the eastern Hoggar—Ténéré domain around 725 Ma ago; (2) N—S collision in the western Pharusian branch around 700 Ma ago and, finally, (3) oceanic closure around 600 Ma ago which led to the E—W collision between the passive continental margin of the West African craton and the active continental margin of the Touareg shield. The suture is marked by a string of positive gravity anomalies corresponding to the emplacement of ultrabasic and basic rocks. The 600 Ma collision was accompanied by the translation of foreland nappes onto the West African craton and affected the entire Touareg shield. The reactivation of older gneissic terrains in the central and eastern part of the shield is marked by greenschist metamorphic overprinting, intense intraplate deformation in N—S linear belts accompanied by crustal thickening, generation of granites and major lateral displacements along megashear zones. An analogy is drawn with Asia and the Himalaya fold belt.

The large amount of predominantly calc-alkaline rocks generated during the Pan-African s.l. (900—500 Ma) corresponds to a major period of crustal accretion which led to the cratonization of Africa.

INTRODUCTION

Ten years ago attention was drawn to the existence of Pan-African island arc and marginal trough volcano-detritic assemblages in the western Hoggar, associated with magmatism of oceanic affinities (Gravelle, 1969; Caby, 1970), and ophiolites were recognized at Bou Azzer (Morocco) along the northern edge of the West African craton (Leblanc, 1970). Burke and Dewey (1970) presented a plate-tectonic model for circum West African Pan-African belts, suggesting a Himalaya-type situation for most of the crust situated to the east of the West African craton. Bertrand and Caby (1978), in an attempt to synthesize the geology of the Hoggar, proposed a model involving subduction processes in the western Hoggar and showed the complexity and varying degree of Pan-African intracontinental deformation across the 800 Km wide shield. Recently, the preliminary results of a detailed geological and geophysical investigation of the critical Iforas—Gourma region in Mali, crossing the West African craton and the Pan-African belt to the east, showed convincing evidence of a Wilson cycle ending with collision between a passive continental margin in the west and an active continental margin in the east (Black et al., 1979b).

The major tectonic units in the region that we shall examine (Fig. 16-1) are, in the west, the West African craton stable since around 1700 Ma ago, and to the east, the Touareg shield (Hoggar, Iforas, Aïr) which belongs to the Pan-African mobile belt. Major N—S shear zones divide the Touareg shield from west to east into the following tectonic domains: (1) The Pharusian belt, comprising a western and an eastern branch, both characterized by the abundance of Upper Proterozoic volcano-detritic material; (2) the polycyclic central Hoggar—Aïr domain largely composed of ancient gneisses reactivated and injected by abundant granitoids during the Pan-African; (3) the eastern Hoggar—Ténéré domain which apparently was to a large extent stabilized at an early stage of the Pan-African episode around 725 Ma ago.

The aim of this chapter is to discuss the relationships between subduction—collision and ensialic processes within the Touareg shield, which underwent a multistage evolution during a period lasting several hundred million years. Whilst we feel that events marking the closing stages of the Pan-African episode related to collision with the West African craton are beginning to be well documented, our enquiry into the older events is much more speculative, but the tentative model presented is compatible with present-day plate-tectonic processes (Dewey, 1977).

For the geological background the reader may refer to regional syntheses by Bertrand and Lasserre (1976), Bertrand and Caby (1978), Bertrand et al. (1978), Black (1978), Black et al. (1979a), to detailed regional accounts by Gravelle (1969), Caby (1970), Bertrand (1974), Davison (1980), and early pioneering work of Lelubre (1952) and Karpoff (1960).

In this paper the term Pan-African covers the time period 900—550 Ma. All ages quoted are recalculated with the new decay constants recommended by Steiger and Jäger. Recent age determinations are given with 1 sigma error.

CONTRASTING PRE-PAN-AFRICAN HISTORY OF THE WEST AFRICAN CRATON AND THE TOUAREG SHIELD

When comparing the nature of the basement constituting the West African craton with that of the pre-Pan-African inliers occurring within the Touareg

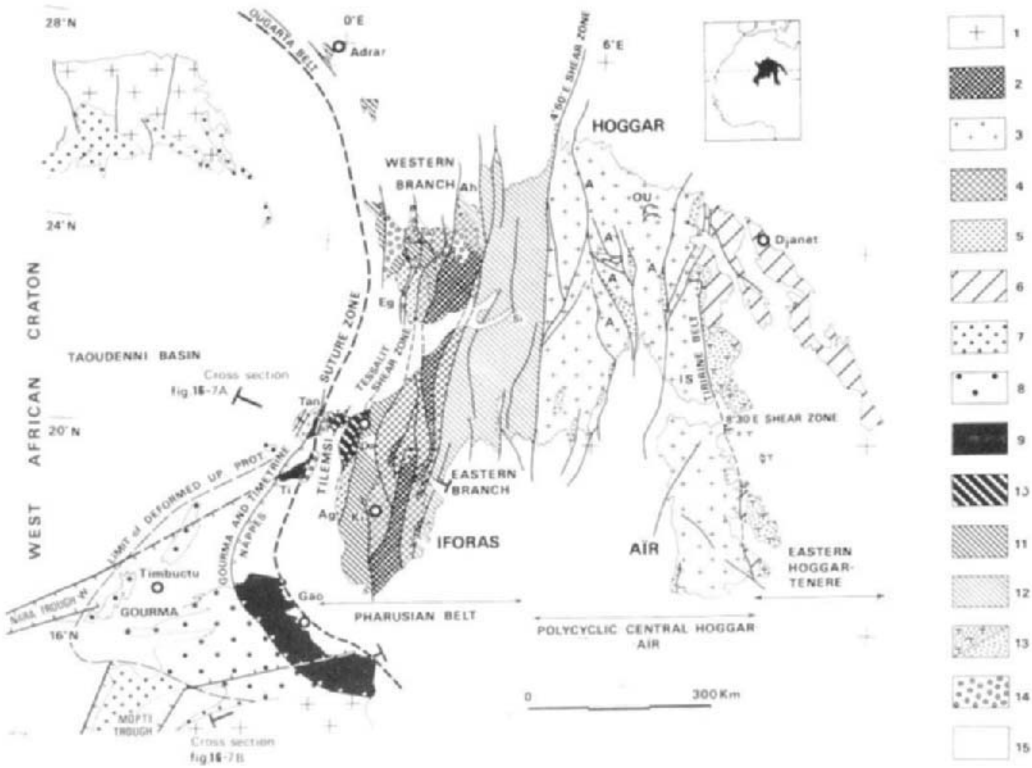


Fig. 16-1. Simplified geological map of the Touareg shield and adjacent areas. *1* = Reguibat and Leo Shields; *2* = slightly reactivated Eburnean granulites within the Pharusian belt; *3* = reactivated pre- Pan-African gneisses; *4* = undifferentiated gneisses highly reactivated during the Pan-African; *5* = *4*, affected by late high-temperature—low-pressure metamorphism; *6* = undifferentiated rocks of the eastern Hoggar (metamorphism at c. 725 Ma); *7* = Upper Proterozoic shelf sediments; *8* = slope-basin sediments of the Gourma aulacogen; *9* = Gourma and Timétrine nappes; *10*—*12* = late Upper Proterozoic greywackes and magmatic rocks of the Tilemsi accretion zone (*10*), of the western Pharusian belt (*11*), of the eastern Pharusian branch (*12*); *13* = late Upper Proterozoic volcano-detritic schist belts in the central and eastern Hoggar with *T* = Tiririne and Proche—Ténéré Groups; *14* = molassic "Série pourprée" (partly Cambrian); *15* = Palaeozoic and Mesozoic cover. *A* = Aleksod assemblage; *Ah* = Ahnet; *Ag* = Aguelhoc; *Eg* = Egatalis; *Tan* = Taounnant; *Ti* = Timétrine; *IS* = Issalane; *Ki* = Kidal; *OU* = Oumelalen; *Ta* = Taoudrart; *Ga* = Gara Akofou; *Ou* = Oumassène; *Si* = Silet.

shield, the most obvious feature is the widespread presence of high-grade rocks in the latter whereas the craton, at least its southeastern part, is composed of low-grade granite metavolcanic-greywacke assemblages reminiscent of Archaean greenstone belts.

Bessoles (1977) in his review of the West African craton underlined its division into: (1) a western domain of Liberian age ($\simeq$ 3000 Ma) almost unaffected by Eburnean reactivation ($\simeq$ 2000 Ma) and consisting essentially of granulites (western Mauritania, Sierra Leone, Liberia, western Ivory Coast), but including some Archaean low-grade greenstone belts (Sierra Leone); (2) the eastern domain is characterized by pre-Eburnean (2700 Ma) reactivated basement and NNE—SSW-trending Birrimian (Lower Proterozoic) belts of metavolcanic-sedimentary rocks, with widespread syntectonic and late tectonic granites emplaced in the time span 2000 $\pm$ 100 Ma. Except for some Eburnean molassic formations, late granites (1700—1600 Ma) and dolerites (1300 Ma), the Middle Proterozoic (c. 2000—1000 Ma) is generally absent in the West African craton, which has remained remarkably stable for the last 1700 Ma.

Pre-Pan-African basement remnants are known in most domains of the Touareg shield. All of these consist of high-grade rocks, often belonging to the granulite facies, metamorphosed during the Eburnean event (2150—2050 Ma). The largest inliers of Eburnean granulites (In Ouzzal—Iforas) display a dominant ENE—WSW structural trend. Their presence however is not due to Pan-African uplift as suggested by Shackleton (1976) as they are unconformably overlain by Middle and Upper Proterozoic formations. Some of the inliers are of Archaean age but the corresponding structure and metamorphism are largely unknown (Ferrara and Gravelle, 1966; Latouche and Vidal, 1974). Unlike on the craton, the Middle Proterozoic is well developed in the Pharusian belt where it is represented by alumina-rich quartzites and associated pelitic schists. They often contain intercalated sills and sheets of alkaline-peralkaline meta-igneous rocks which have yielded zircon Concordia dates of 1742 $\pm$ 12 Ma and 1843 $\pm$ 3 Ma (U. Andreopoulos-Renaud, unpubl. data). In the central Hoggar several sequences of supracrustal rocks have also been attributed to the Middle Proterozoic (Bertrand, 1974; Latouche, 1978).

These differences of the basement in the craton and in the Touareg shield are thought to be significant. They suggest that the position of the future Pan-African belt may have been predetermined by the presence of a high-grade Eburnean mobile belt so that oceanic opening eventually occurred along a line of weakness on the edge of the granulites.

On the craton the early Upper Proterozoic (c. 1000—800 Ma) is represented by flat-lying platform deposits, the best known and most representative being the Atar el Hank Group which consists of sandstones, overlain by stromatolite-bearing limestones (Bertrand-Sarfati, 1972; Trompette, 1972). Within the Pan-African belt a very similar sequence was first described in the northwestern Hoggar under the name "Série à stromatolites" (Fabre and Freulon, 1962; Caby, 1970) and this group, with similar shelf-type facies and stromatolite associations (Bertrand-Sarfati, 1969), has now also been identified in several areas in the western part of the Touareg shield. The early Upper Proterozoic marks a period of very widespread cratonic sedimentation, and we believe that there may have been continuous continental crust from the West African craton to the Touareg shield.

EARLY PAN-AFRICAN RIFTING DURING THE UPPER PROTEROZOIC

Development of a passive continental margin

A major change in sedimentation occurred about 800 Ma ago with rifting along the eastern margin of the West African craton, leading to the development of a passive continental margin. In Mali, a triple point coincides with an embayment in the craton: the Gourma (Fig. 16-1) is thought to be a failed arm which evolved as an aulacogen, a comparable situation with that of the Benue trough with respect to the Gulf of Guinea in the Cretaceous (Grant, 1971). The Gourma basin is characterized by deep subsidence with an accumulation of over 8000 m of sediments (Reichelt, 1972). The observed sedimentary sequences comprise an early terrigenous clastic phase at the base, followed by differentiated carbonate sediments indicating lateral passage platform—slope—trough, and end with prograde continental clastic deposits. These sediments thicken eastwards and are typical of deposits formed along the passive continental margin adjoining a mature wide ocean (Moussine-Pouchkine and Bertrand-Sarfati, 1978). The shape of the basin, as defined by the distribution of slope sedimentary facies and the gravity pattern, is that of a trough oriented WSW—ENE and representing a gulf within the West African craton. The aulacogen is marked by positive gravity anomalies which can be traced westwards across the craton where they split, defining the axes of the Nara and Mopti troughs. The Gourma formations have been affected by open folding parallel to the trend of the aulacogen, prior to late Pan-African regional folding which outlines the concave virgation of the Gourma.

To the south, in Togo-Benin, recent studies of the Upper Proterozoic (Simpara, 1978; Trompette, 1980) also indicate a passage from thin platform deposits to thick marine continental margin deposits as one goes eastwards towards the edge of the craton.

Magmatism associated with continental fragmentation

The first stage of rifting is preserved in the northwestern Hoggar where deformed pre-metamorphic dyke swarms of alkaline and peralkaline metarhyolites, undersaturated soda-trachytes and phonolites with associated

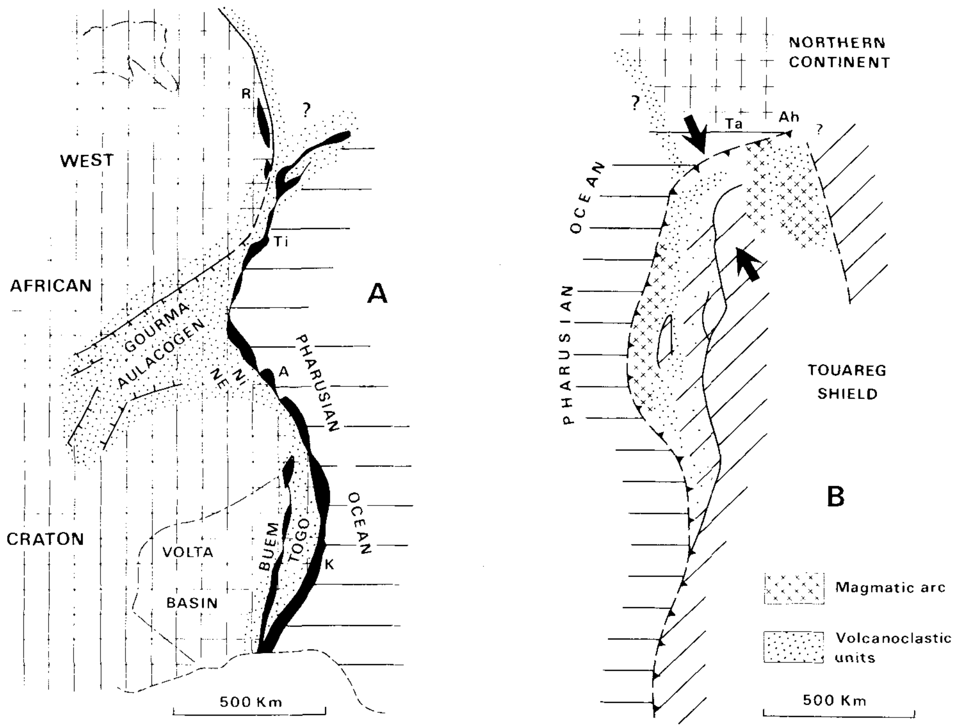


Fig. 16-2. Tentative palaeogeographical reconstructions.
A. Passive margin of the West African craton after ocean opening at c. 800 Ma. Dots represent the strongly subsiding areas; zone of basic and ultrabasic intrusions in the collapsed margin are shown in black; *Ti* = Timétrine; *A* = Amalaoulaou; *K* = Kandé; *R* = palaeorift. *Ni* and *NE* = inferred areas of sedimentation of internal (*Ni*) and external nappes (*NE*) compared to Togo-Atacorian and Buem, respectively.
B. Active margin of the eastern continent (Touareg shield) with its western accretion domain dying out southwards and its northern magmatic arc. Arrows outline the early NNW—SSE collision with a northern continent at c. 700 Ma. *Ta* = Tassendjanet nappe; *Ah* = Ahnet deltaic quartzites.

plutonic rocks are superimposed on earlier basic dyke swarms (Fig. 16-2A). On field, petrological and geochemical grounds, this complex is interpreted as representing the roots of a palaeorift (Dostal et al., 1979).

The intrusion of voluminous basic and ultrabasic rocks in the form of sills, laccoliths and stocks has occurred over wide areas of both continental margins. These complexes show some petrological and geochemical affinities with ophiolites but field observations (Caby, 1970, 1978) have shown these mantle-derived complexes to have been intruded into sediments prior to regional low-grade metamorphism. They belong to a large-scale magmatic event occurring around 800 Ma ago (Clauer, 1976; De la Boisse, 1979) and are thought to be related to the process of continental fragmentation (Fig. 16-2A). Three examples are given below:

(1) The Ougda—Tassendjanet complex ($\simeq 900\,km^2$), outcrops along one side of a major fault which may have controlled its emplacement into quartzites and limestones of the "Série à stromatolites". Garnetiferous metagabbros grading into garnet amphiclasites and pyriclasites may represent early stages of cooling of a gabbroic magma deep in the crust in granulite-grade conditions. Amphibole gabbros, fine-grained carbonate-rich serpentinites and foliated quartz diorites cut through stromatolite-bearing marbles and dolomites, producing high-grade contact metamorphism. Heterogeneous quartz-gabbros, tonalites and trondhjemites constitute the bulk of the complex, with lesser amounts of granodiorites. Many amphibolitized dykes of andesitic composition are also associated with the gabbros and may represent late sub-volcanic conditions. A recent geochemical study (C. Dupuy, pers. commun., 1979) concluded that the gabbroic rocks are LREE-depleted and display patterns typical of oceanic tholeiites and that younger rocks show progressive calc-alkaline affinities.

(2) The Amalaoulaou meta-igneous complex (Fig. 16-7B) coincides with a positive gravity anomaly marking the suture with the West African craton. It comprises layered amphibole-garnet pyriclasites, two pyroxene pyriclasites and serpentinites. According to De la Boisse (1979) metagabbros have primarily crystallized in granulite facies conditions at the base of the crust from a magma of tholeiitic affinity and have subsequently been tectonically emplaced along the suture. U-Pb dating on zircon (De la Boisse, 1979) has given an age of 810 ± 50 Ma, considered as that of magmatic crystallization. Undeformed quartz gabbros were intruded at 730 ± 40 Ma ago before the development of greenschist facies metamorphism with local blue amphibole.

(3) The Timétrine ultrabasic rocks occur as four elongated boudinaged massifs, interpreted as being incorporated in a nappe west of the suture (Fig. 16-3). They are composed of highly serpentinized rocks and represent metadunites and some metawherlites, lherzolites and harzburgites, displaying a foliation and strong chromite lineation. Lenses of metagabbro and diabase are found within the serpentinites. The presence of blue amphibole and aegyrine in the surrounding country rocks is restricted to some tens of metres close to the ultrabasic bodies (Karpoff, 1960). Overlying the chlorite-albite schists and the sericite quartzites which surround the serpentine massifs are found pillow basalts and their feeder dykes. Leblanc (1976) suggested that these rocks represent a dismembered ophiolite sequence obducted into a thick epicontinental sedimentary formation. On the other hand, Caby (1978) indicated that the contacts between fine-grained carbonated serpentinites and compact chlorite schists may be regarded as magmatic and that the ultrabasic rocks were intruded as sills and dykes into the schists.

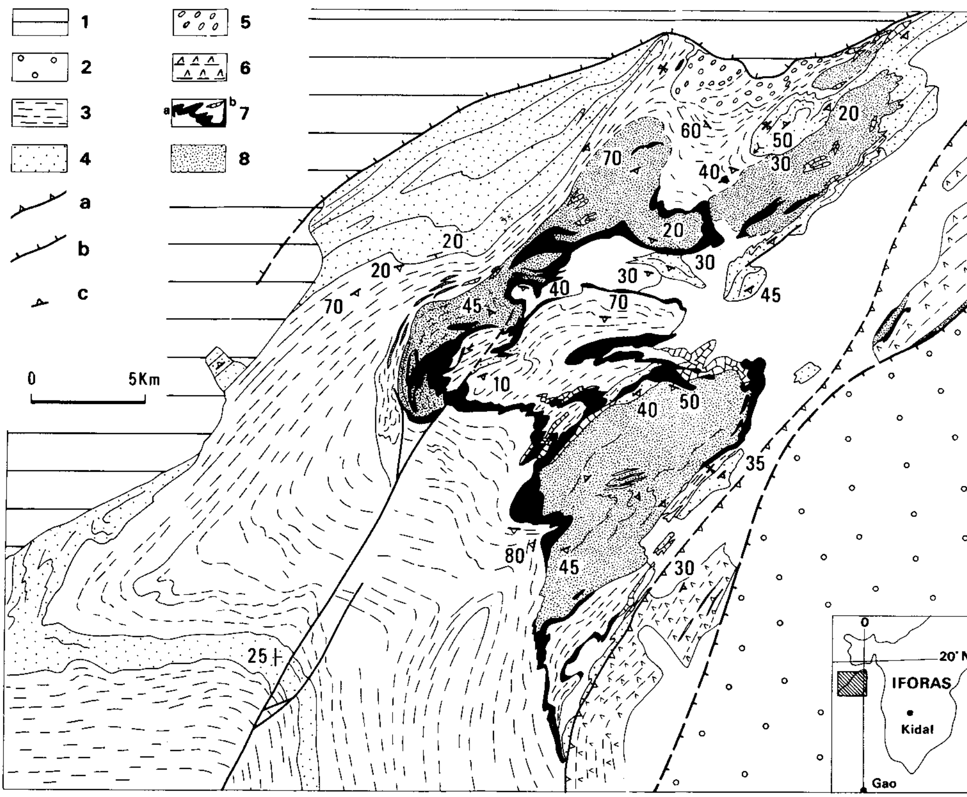


Fig. 16-3. Simplified geological map of the Timétrine area (presumably allochthonous). *1* = Cretaceous cover; *2* = molassic deposits (Nigritian), partly of Cambrian age; *3* = sericite-chlorite schists; *4* = sericite quartzites; *5* = pillowed metabasalt and related diabase, apparently conformable on *3*; *6* = diabase and metabasalt (blue amphibole bearing); *7a* = red hematite jaspers; *7b* = Ca-Fe-Mg carbonates; *8* = ultrabasic rocks; *a* = thrust; *b* = fault; *c* = cleavage.

INITIATION OF OCEAN CLOSURE AND SUBDUCTION: CORDILLERAN-TYPE EVOLUTION IN THE PHARUSIAN BELT

In the western part of the Pan-African Touareg shield, the late Upper Proterozoic deposits of the Pharusian belt are totally different from those of the West African craton. They have many characteristics comparable with island arc and modern active continental margins suggesting the existence of an open ocean separating two continents. The deposits reflect a very complex palaeogeography of basins, troughs and volcanic chains and were accompanied by widespread calc-alkaline volcanism and plutonism. Deposition was not everywhere synchronous and several cycles have been distinguished.

The oldest known rocks occur in the central part of the eastern Pharusian branch around Silet (Fig. 16-1). Two important lithological units separated by an unconformity (Bertrand et al., 1966) have been recognized by Gravelle (1969):

(a) A *lower unit* consisting of a volcano-detritic sequence with flows and sills of intermediate to acid composition, pillow basalts, sills and dykes of dolerite. It overlies marble and quartzites injected by basic and ultrabasic rocks. This assemblage is cut by a large composite batholith of quartz diorite, granodiorite, trondhjemite and adamellite which has yielded a U-Pb Concordia date on zircons of 887 ± 1 Ma (U. Andreopoulos-Renaud, unpubl. data). Early regional folding and greenschist facies metamorphism seem to be related to this major cycle.

(b) An *upper unit* comprising a volcano-sedimentary sequence including pelites and greywackes with a basal conglomerate and a volcanic sequence almost exclusively composed of andesites, dacites, pyroclastics, cut by dykes of andesitic composition.

During this magmatic cycle quartz diorites and adamellites dated at 831 ± 5 Ma (U-Pb on zircons, U. Andreopoulos-Renaud, unpubl. data) were emplaced in the form of elongate N—S-trending batholiths, up to 80 km long and 10—35 km wide, with numerous intraplutonic sub-volcanic rocks suggesting a high-level of emplacement, the country rocks being converted to hornfels. In the northwestern part of the eastern Pharusian branch thick accumulations of slightly deformed lavas of andesitic to rhyolitic composition are cut by calc-alkaline batholith and crop out over wide areas which are little known. Their relationships with adjacent poly-deformed metagreywackes and lavas and with the units distinguished by Gravelle (1969) have not yet been established. The vast amount of predominantly calc-alkaline volcanic and plutonic rocks emplaced during a period exceeding 200 Ma suggests a long Cordilleran-type evolution in the eastern Pharusian branch (Fabriès and Gravelle, 1977). Detailed geochemical studies of the volcanics from the Silet region show that the majority of samples display calc-alkaline trends with moderate LREE enrichment and La/Yb ratios between 5 and 8, typical of modern island-arc calc-alkaline rocks. They also resemble andesites and high-Al basalts from central-south Chile which are related to a relatively flat-lying Benioff zone (Chikhaoui et al., 1980). The old continental crust underlying Cordilleran-type assemblages is known from some high-grade slices along major faults and has been sampled as sialic granulite xenoliths in recent volcanoes (M. Girod, pers. commun., 1979).

In the western Pharusian branch (Fig. 16-1) one can distinguish late Upper Proterozoic volcano-detritic sequences which have been affected by two major phases of Pan-African deformation (e.g. the "Série verte"), and others which have only been subjected to late Pan-African tectonism (e.g. Taoudrart, Gara Akofou and Oumassène). The earlier (pre-700 Ma) "Série verte" occurs in synclinoria superimposed on subsiding basins already initiated during the Middle Proterozoic. It consists of a 6000 m thick sequence of flysch, volcanic greywackes and pebble conglomerates with intercalated andesites, dacites and dacitic breccias which may be derived from volcanic islands. Basic and calc-alkaline plutonic rocks (diabase, quartz diorite and granodiorite) have invaded the greywackes in the form of sills and stocks, but similar rocks are also found in the conglomerates suggesting an autocannibalistic evolution. The chemical composition of the greywackes is similar to modern examples and points to a lack of terrigenous and clay components even in fine-grained rocks (Caby et al., 1977). In contrast, the later *Taoudrart volcanics* (5000 m thick) are composed of andesite flows and intercalated pyroclastics with dacites and continental polygenic conglomerates on top, laterally interfingering with agglomerates, tuffs and greywackes. The *Gara Akofou volcanics* consist of a 2000 m thick sequence of basic and intermediate lava flows with intercalated pyroclastics, preserved in a fault-bounded monocline within the granulitic In Ouzzal unit. It probably represents the relics of an extensive volcanic cover. The *Oumassène volcanics* (2000 m) in the Iforas are composed mainly of andesite flows with intercalated pyroclastics. Trace element data for the volcanics from these three occurrences (Taoudrart, Gara Akofou and Oumassène) are consistent with derivation by partial melting from an upper-mantle source enriched in LILE (Chikhaoui et al., 1978). The rock suites display typical calc-alkaline patterns with pronounced LREE enrichment and La/Yb ratios between 12—14, typical of modern continental margin andesites associated with steeply dipping Benioff zones. Such similarities can be extended to other trace elements such as Zr, Nb, Hf and P (Chikhaoui et al., 1980).

The Tilemsi strip (Fig. 16-1) close to the West African craton is believed to represent a zone of accretion largely devoid of ancient basement. It is composed of greywackes with turbidites and associated conglomerates as well as occasional Al-rich pelitic layers, which may represent deep-sea clays overlying dacitic breccias and a lower metabasalt unit. These rocks are injected by a large volume of pretectonic plutonic rocks in the form of sills, dykes and massifs, a pretectonic adamellite having yielded a U-Pb Concordia age on zircon of 633 ± 3 Ma (U. Andreopoulos-Renaud, unpubl. data). These plutonic rocks exhibit a marked E—W polarity: whilst quartz gabbros abound close to the suture, eastwards the terrain is invaded by large massifs of quartz diorite. As one approaches the margin of the Iforas, granodiorites and adamellites predominate. A polarity is also displayed by the nature of the sediments, the amount of clay and terrigenous material increasing eastwards. In contrast to the deep trough environment in the west, syn- and late-tectonic flysch-molasse deposits and associated andesites occur along the eastern margin of the Tilemsi. Farther east, in the western Pharusian branch underlain by ancient basement, we have already pointed out the presence of thick andesite sequences displaying typical geochemical characteristics of continental margin, Cordilleran-type volcanics which we relate to an easterly dipping subduction zone.

We are not yet in a position to formulate a clear plate-tectonic reconstruction of the Tilemsi accretion zone. However, pending detailed mapping and geochemical studies, our field observations suggest a transition between an island arc in the west and a Cordilleran-type continental margin to the east, although a back-arc suture has not yet been discovered. The permanent contribution of mantle-derived rocks to this zone is exemplified by some syn- and late-tectonic high-level gabbro-peridotite bodies.

To conclude, the sharply contrasting sedimentological environments which developed in the late Upper Proterozoic, correspond to the formation of a passive continental margin on the edge of the West African craton and an active continental margin along the western margin of the Touareg shield. This has led us to question the hypothesis of a fixed Gondwanaland throughout the Proterozoic (Piper et al., 1973), which has strongly influenced geological thinking in Africa (Shackleton, 1976). The geological evidence from West Africa shows that continental fragmentation occurred around 800 Ma ago as was the case around the North American continent (Stewart, 1976). Critical reappraisal of the palaeomagnetic evidence supports the idea that important horizontal displacements occurred in the late Upper Proterozoic (Burke et al., 1976; Briden, 1977; Morel-à-l'Huissier and Irving, 1978; Black, 1978). There are very few palaeomagnetic data from West Africa and new measurements are urgently required in order to estimate the possible width of a Pharusian ocean situated to the east of the West African craton.

MULTISTAGE STRUCTURAL EVOLUTION OF THE PAN-AFRICAN TOUAREG SHIELD

Structural domains in the Touareg shield

In contrast to the West African craton, which has been stable throughout the Middle and Upper Proterozoic, parts of the Pan-African Touareg shield have had a complex history of deformation prior to collision with the West African craton around 600 Ma ago. A striking feature of the shield is the spectacular development of N—S shear zones which delimit the following structural domains (Fig. 16-1) from west to east:

The Pharusian belt

The Pharusian belt, as we have seen, is characterized by thick accumulations of late Upper Proterozoic volcano-detritic assemblages and associated calc-alkaline plutonism. We distinguish:

(a) The *Tilemsi strip*, lying immediately to the west of the Iforas which is believed to represent an accretion zone apparently largely devoid of ancient sialic basement. It has been affected by pronounced E—W shortening.

(b) The *western Pharusian branch*, floored by Eburnean granulites with a cover of Middle and Upper Proterozoic strata and their products of reworking during the Pan-African event. It has been affected by two major phases of Pan-African deformation, producing early WSW—ENE-trending structures and late N—S-trending folds.

(c) The *eastern Pharusian branch*, which is also floored by an ancient substratum. Lithological and metamorphic gradients are N—S and there is evidence for a pre-885 Ma tectonic event; the main orogenic imprint produced intense folding with a N—S trend.

The polycyclic central Hoggar—Aïr domain

The polycyclic central Hoggar—Aïr domain consists mainly of pre-Pan-African basement rocks (Archaean and Lower Proterozoic), injected by abundant Pan-African granitoids. The existence of a presumed "Kibaran" event (Bertrand, 1974; Latouche, 1978) and of narrow volcano-sedimentary belts of presumed Upper Proterozoic age is reminiscent of the pattern described in south- and northwestern Nigeria (Grant, 1978; Holt et al., 1978; Hubbard, 1978).

The eastern Hoggar—Ténéré domain

The eastern Hoggar—Ténéré domain, apparently stabilized around 725 Ma ago, includes along its western margin a late Pan-African ensialic linear belt developed along a shear zone (Tiririne belt, Bertrand et al., 1978). The eastern edge of the polycyclic central Hoggar—Aïr domain is affected by this late Pan-African event.

The "Kibaran" event (1200—900 Ma ago)

This event, recorded in Nigeria (Grant, 1972; Ogezi, 1977), is thought to have affected the Aleksod (Bertrand, 1974) and Gour Oumelalen (Latouche, 1978) regions of the Hoggar and may have had a wider extension in the central Hoggar—Aïr polycyclic domain (Fig. 16-1). Structural relationships with the early event in the eastern Pharusian branch are still unknown. In both the Aleksod and Gour Oumelalen areas a supracrustal sequence, forming large scale recumbent folds of Pennine-type overturned to the NW, has undergone Barrovian metamorphism in high-grade amphibolite facies with widespread migmatization. The supracrustal sequence is structurally unconformable on basement of Eburnean or Archaean age. The "Kibaran" ages, however, should be accepted with caution as both in Nigeria and in the Hoggar they have been obtained using Rb-Sr whole-rock and K-Ar methods (Picciotto et al., 1965; Grant, 1972; Bertrand and Lasserre, 1976) on metamorphic rocks only.

The event around 725 Ma ago

Poorly known tectonic events of this age are developed in the eastern Hoggar—Ténéré domain (Fig. 16-1). This region was metamorphosed under low greenschist to upper amphibolite facies conditions and has suffered strong E—W shortening, responsible for tight upright folds with N—S to NNW—SSE trend and associated b-lineations. Synkinematic sub-alkaline granitoids and a late granodiorite batholith have yielded a U-Pb Concordia age on zircon of 726 ± 22 Ma (U. Andreopoulos-Renaud, unpubl. data). A similar age was obtained on some granites west of the 8°30′E shear zone (Latouche and Vidal, 1974) but their extent and significance is still unknown. Note that the lithologies, metamorphism and structures differ completely on either side of the 8°30′E shear zone, indicating that this lineament may be the locus of a cryptic suture hidden to the north beneath the late Upper Proterozoic Tiririne Group. Thus, the eastern Hoggar—Ténéré domain which, in the absence of geochronological data, had previously been called the eastern craton (Bertrand et al., 1978), now appears to represent an early Pan-African mobile belt stabilized around 725 Ma ago. The Tiririne Group (or Proche—Ténéré Group in Niger) is the molasse related to this early Pan-African orogenic belt.

The event around 700—680 Ma ago

A major orogenic event, affecting both the early Upper Proterozoic "Série à stromatolites" and the volcano-clastic "Série verte", occurred in the western Pharusian branch. It is marked by NNW—SSE movements causing overthrusting of older crust and important crustal thickening.

In the northwestern Hoggar very high structural levels are exposed in the Tassendjanet nappe (Caby, 1970) (Figs. 16-2B and 16-4). This nappe of

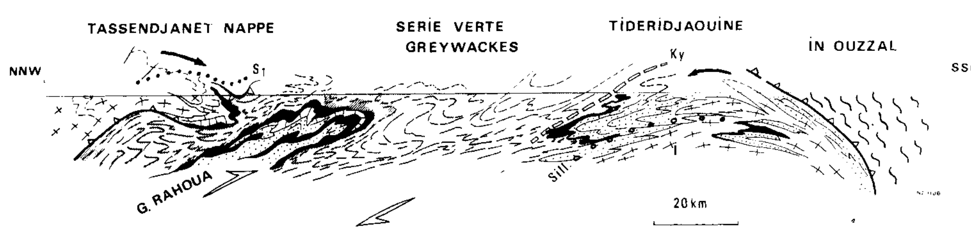


Fig. 16-4. The early Pan-African collision in the western Hoggar at c. 700 Ma. The Tassendjanet nappe is considered as part of a northern continent. S_1 is the upper limit of slaty cleavage. Ultrabasic rocks (in black) associated with quartzites may represent a lower thrust unit (Gour Rahoua). The Tideridjaouine quartzites (dotted) and gneisses (crosses) belong to a deeper structural level of isoclinal folds facing north. Both units moved synchronously and are interpreted as related to the closure of a sub-oceanic domain with a basified crust covered by many thousand metres of greywacke of the "Série verte".

Eburnean granites, with its overlying cover of the "Série à stromatolites" tectonically accumulated at its front, moved south-eastwards over more than 40 km as demonstrated by the presence of a tectonic window and frontal klippen. The underlying para-autochthone is composed of greywackes and plutonic rocks of the "Série verte". Deeper structural levels are gradually exposed to the south below the "Série verte" in the underlying Middle Proterozoic quartzites which are isoclinally folded in amphibolite facies conditions of Barrovian type. These folds, generated at depth, face northwards in contrast to the southward directed movement of the Tassendjanet nappe.

Contemporaneously, an elongated N—S unit of Eburnean granulites was thrust to the north-northwest farther south in the Iforas (Boullier et al., 1978; Boullier, 1979; Wright, in press). The unit has been later affected by lateral mylonite zones which sharply cut Eburnean and early Pan-African structures. The lithologically similar but little deformed In Ouzzal granulite unit of the northwestern Hoggar is also laterally cut by important vertical faults and contrasts tectonically with all levels of the adjacent, highly deformed Pan-African edifice. To the south, the In Ouzzal unit is progressively incorporated in the tangential structures of the Iforas granulite nappe.

Following Boullier et al. (1978), the Iforas granulite unit is underlain by a complex gneissic unit of high-grade amphibolite facies of Barrovian type with widespread migmatites underlying the granulite nappe in the Iforas. Its main structural characteristics are a well-defined flat-lying foliation or banding, often refolded, and a ubiquitous NNW—SSE-trending lineation parallel to the motion direction of the nappe (Boullier, 1979). All the contacts between the heterogeneous components of the assemblage are either tectonic or intrusive, but lateral transition to a less-deformed assemblage shows that it consists essentially of retrogressed Eburnean granulites, a cover of Upper Proterozoic age and pre-tectonic magmatic rocks.

As we have seen this major event involving refolding of huge crystalline nappes during high-grade, Barrovian metamorphic conditions exhibits opposite transport directions in the northwestern Hoggar and in the Iforas. We tentatively link this crustal shortening event to plate motion along NNW—SSE direction and to the closure of an E—W-tending sub-oceanic domain along the northern edge of the In Ouzzal granulite block and possible non-identified oceanic crust hidden beneath greywacke and trough deposits (Fig. 16-2B).

The precise dating of this early Pan-African collision around 700—680 Ma has still to be confirmed as it is based on only two results: a U-Pb determination on zircon from a syntectonic granite in the Iforas which gave a Concordia age of 693 ± 1 Ma (Ducrot et al., 1979) and a Rb-Sr date of 685 ± 15 Ma (Clauer, 1976) obtained on clay minerals from the "Série à stromatolites", incorporated in the Tassendjanet nappe.

The fact that the events around 1200—900 Ma, 725 Ma and 700—680 Ma are each confined to a particular structural domain implies that important lateral displacements occurred along the major shear zones delimiting the domains and suggests that the Touareg shield consists of an amalgam of microplates (Black, 1978).

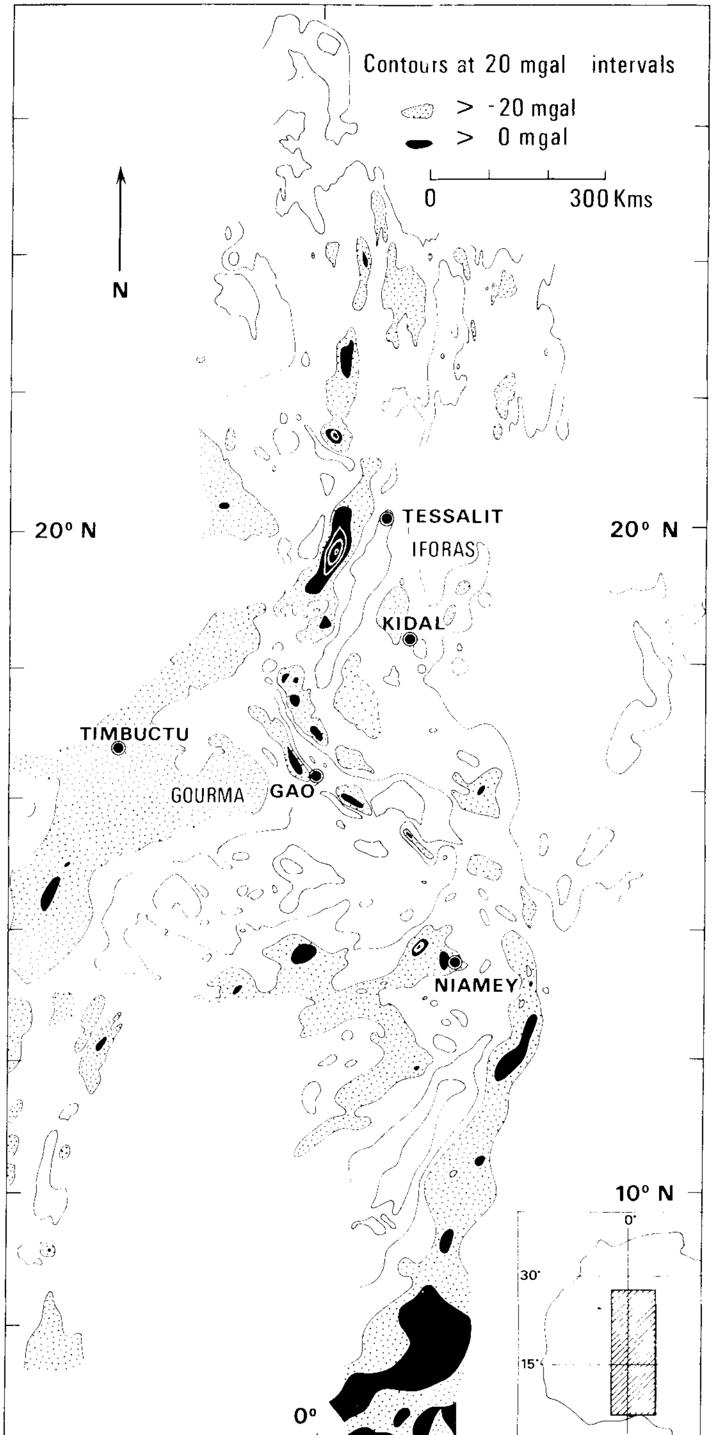
PAN-AFRICAN COLLISION OF THE WEST AFRICAN CRATON WITH THE TOUAREG SHIELD AROUND 600 Ma AGO

This major event affects the entire Touareg shield and overprints the earlier tectonic patterns. East—west shortening varies in intensity and is probably diachronous across and along the Pharusian belt.

The suture

The presence of a suture zone was first suggested in southern Morocco by Leblanc (1970), who interpreted the Bou Azzer basic and ultrabasic complex as obducted Upper Proterozoic ophiolites, now squeezed in a major vertical lineament (*see also Leblanc, this volume, Chapter 17, ed.*). The lineament was considered to extend, without interruption, from Morocco to the western Hoggar and to represent the eastern limit of the West African craton (Caby, 1970) where it is outlined by elongated positive gravity anomalies (Crenn, 1957; Louis, 1970). In Mali these anomalies have been shown to correspond to basic and ultrabasic rocks, but overthrust sheets were also emplaced west of the lineament upon the edge of the West African craton (Caby, 1978, 1979; Wright, in press). The most striking feature, however, is the abrupt juxtaposition, along the lineament, of terrigenous sediments deposited on a passive continental margin to the west, with volcano-clastic plutonic assemblages formed by accretion processes in a typical active continental margin to the east. This led us to consider the lineament as a suture zone (Black et al., 1979a, b). Reactivation of this lineament occurred during the Hercynian event (Ougarta belt) and during the Mesozoic—Tertiary (Gao trough).

Detailed gravimetric surveys (Fig. 16-5) have shown the existence of a string of positive gravity anomalies with amplitudes of over 30 mgal, locally reaching 80 mgal, which may be followed over a distance of 2000 km. The anomalies are of two types: (a) in Togo-Benin positive and negative anomalies of long wave-length which, taking into account the lighter sediments of the passive continental margin (Buem—Atakorian), can be interpreted in terms of thickening of the West African craton towards the suture and a steep contact with an even thicker, dense Pan-African block (Louis, 1978); (b) positive anomalies of shorter wave-length correspond to more superficial structures which display subvertical to easterly dipping geometries in accord with the general vergence and thrusting motion towards the craton.



Contours at 20 mgal intervals
> -20 mgal
> 0 mgal
0
300 Kms
N
20° N
TESSALIT
IFORAS
20° N
KIDAL
TIMBUCTU
GOURMA
GAO
NIAMEY
10° N
0°
30°
15°
0°
0°

These correspond to the position of basic and ultrabasic complexes, and the interpretation of gravity profiles shows these bodies to be rootless and to continue to depths of 6 to 20 km (Bayer and Lesquer, 1978; Ly, 1979).

The location of the two types of anomalies seems to be related to the geometry of the craton which we believe reflects the original shape of the continent before collision. The first type of paired gravity anomalies in Togo-Benin occurs where the craton forms a promontory. This situation led to the complete disappearance, by subduction, of the oceanic floor, thus locally bringing into direct contact by flat underthrusting the low-grade metasediments along the passive continental margin of the craton and the high-grade gneisses of the eastern continent. In contrast, the second type of anomalies is located in an embayment where island-arc and Cordilleran volcano-clastic assemblages of the active margin of the eastern continent have been preserved (Black et al., 1980).

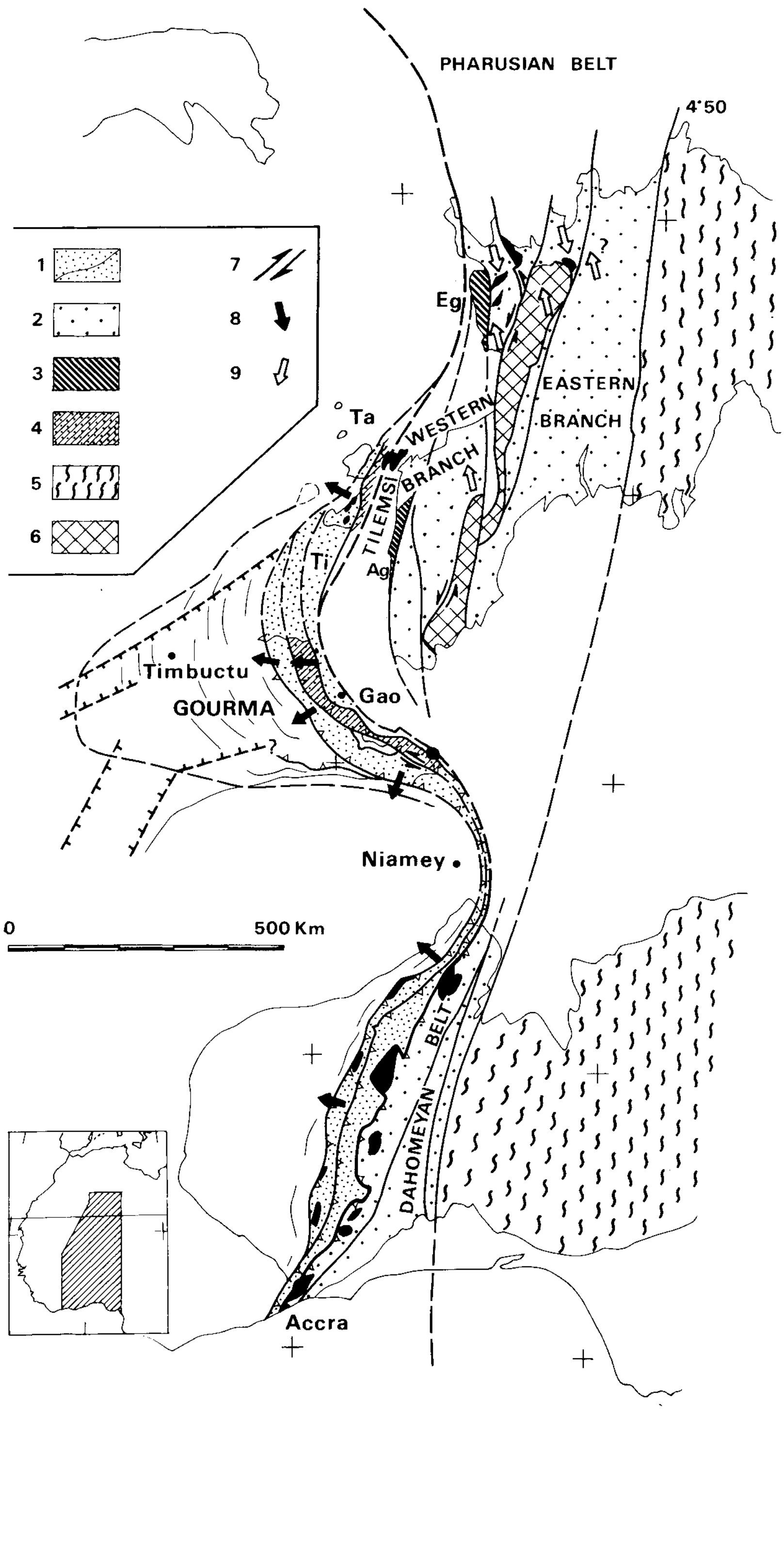
Foreland nappes west of the suture

Nappes have been translated westwards onto the West African craton. In the Taounnant region (Figs. 16-1, 6 and 7A) mylonitized quartzites belonging to the passive margin are overlain by a nappe of blue amphibole-bearing metabasalts, associated with lenses of serpentine which may represent oceanic material. These rocks rest directly upon Eburnean rocks of the craton. Farther south in Timétrine the quartzites, phyllites, serpentines and metabasalts are also probably allochthonous.

In the Gourma aulacogen the sediments have been strongly deformed, implying shortening of several tens of kilometres (Fig. 16-7B). In the east, towards the mouth of the aulacogen, one observes a pile of pellicular nappes which have been transported westwards and south-westwards over about 80 km (Caby, 1978, 1979). The schists forming the external nappes belong to the passive margin and have been metamorphosed in the greenschist facies. In contrast, the quartzites and schists of the internal nappes have undergone high-pressure metamorphism with development of eclogitic mineral assemblages. Farther south the front of the Dahomeyan belt also shows a succession of units thrust westwards which have been described by African (1975), Simpara (1978) and Trompette (1980).

The striking feature of this external zone of the Pan-African belt west of the suture and on the passive margin of the West African craton is the complete absence of autochthonous Pan-African magmatism.

Fig. 16-5. Simplified Bouguer anomaly map of the eastern margin of the West African craton.


PHARUSIAN BELT
4°50
1
2
3
4
5
6
7
8
9
Eg
Ta
WESTERN
BRANCH
TILEMSI
EASTERN
BRANCH
Ti
Ag
Timbuctu
GOURMA
Gao
Niamey
0
500 Km
DAHOMEYAN
BELT
Accra

Paired metamorphic belts

The Tilemsi zone of accretion immediately east of the suture has been interpreted as representing the relic of an island arc and a marginal cordillera. It is flanked by zones of contrasting metamorphic type (Fig. 16-6). To the west allochthonous rock units resting on the West African craton close to the suture display high-pressure—low-temperature metamorphism, whereas to the east, along the border of the western Pharusian branch, zones of low-pressure—high-temperature metamorphism are developed. The former are characterized in the northeastern Gourma by eclogitic micaschists which contain jadeitic pyroxene. Probe-data indicate formation under pressure of the order of 15 kb and a temperature of about 650° C (Reichelt, 1972; De la Boisse, 1979). The blue amphibole-bearing metabasalts of Timétrine and Taounnant may also belong to this metamorphic unit. The latter zone of low-pressure—high-temperature metamorphism, represented by the Aguelhok gneisses (Iforas) and the Egatalis gneisses (northwestern Hoggar), displays fresh mineral assemblages of cordierite-garnet-sillimanite-spinel, and these gneisses are sometimes associated with hypersthene-bearing rocks. They represent zones situated along the western edge of the Touareg shield which were subjected to a high-thermal regime and to strong uplift following collision, as confirmed by young biotite cooling ages.

Deformation in the Pharusian belt east of the suture

In the Tilemsi accretion zone polyphase deformation in amphibolite facies grade close to the suture passes eastwards into N—S trending open to tight folds of high structural level.

In the western Pharusian branch E—W shortening is negligible in the Tassendjanet nappe but gradually increases southwards (Caby, 1970). N—S-trending structures with recumbent to isoclinal folds and later open folds with similar trend represent the main structures, both in the "Série verte" and the high-grade rocks of Egatalis. The structural evolution of the northwestern Hoggar is bracketed within the time range 625—570 Ma (Allègre and Caby, 1972); the first figure corresponds to syn-kinematic granite emplacement and the second figure to late granitoids, believed to be roughly synchronous with the peak of metamorphism at Egatalis. Late stages

Fig. 16-6. Simplified sketch-map of the Pharusian—Dahomeyan belt after collision between the West African craton and the Touareg shield. *1* = Gourma and Buem—Atacora nappes; *2* = undifferentiated rocks mainly of Pan-African metamorphism (*Eg* = Egatalis and *Ag* = Aguelhoc); *4* = high-pressure—low-temperature metamorphism of the Gourma nappes; *5* = undifferentiated reactivated pre-Pan-African rocks in the central Hoggar and Nigeria; *6* = Eburnean granulites; in black: metabasic to ultrabasic rocks; *7* = strike-slip fault; *8* = late movements related to the collision with the West African craton; *9* = early collision in the Northern Iforas and western Hoggar; *Ta* = Taounnant; *Ti* = Timétrine.

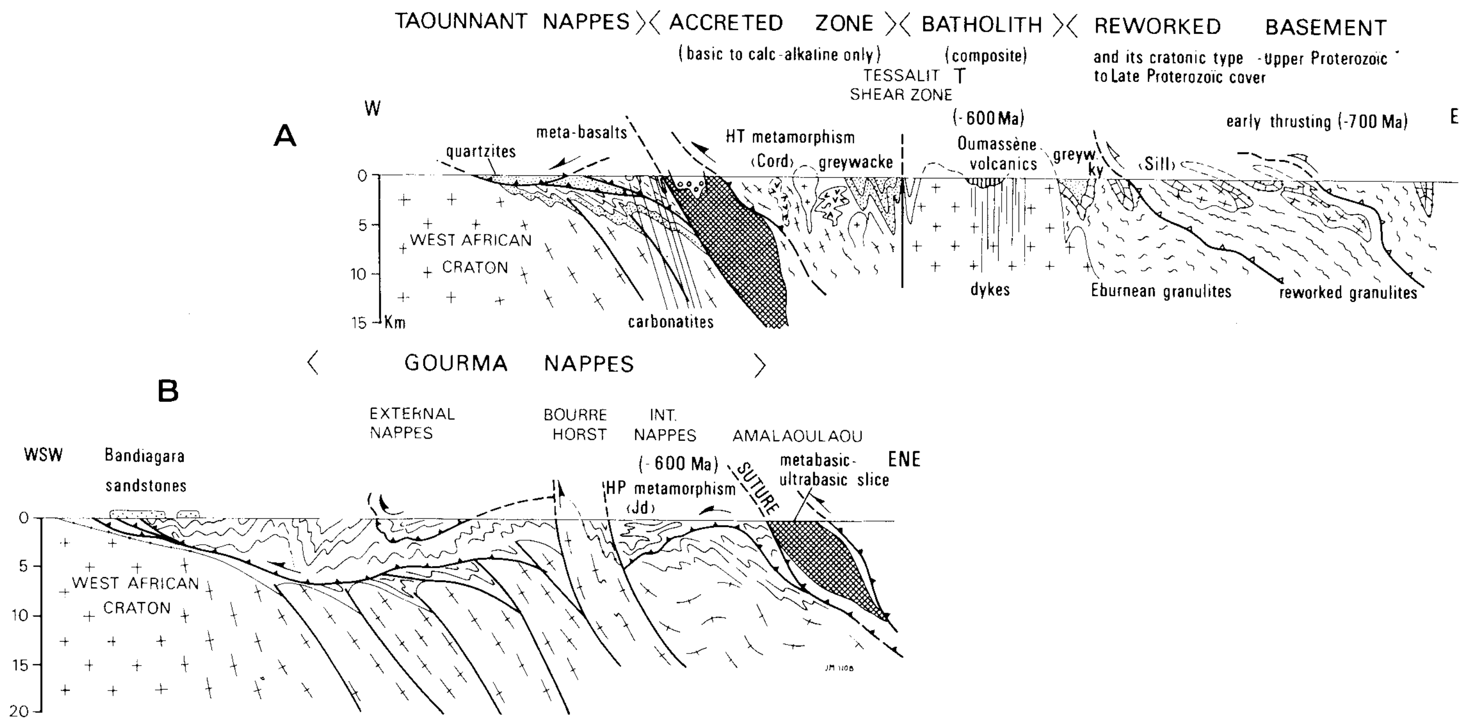


Fig. 16-7. Schematic geological cross-sections A. northern Adrar des Iforas; B. the Gourma nappes. The location of the sections is indicated in Fig. 16-1.

of this deformation are related to the main N—S-trending sinistral shear zones and branching strike-slip faults (Caby, 1970), some of which are outlined by ultrabasic to basic and peralkaline rocks. Farther south, in the central Iforas, E—W shortening is less pronounced but important shearing took place at that time. N—S-trending sinistral movement along shear zones and gneissic doming are followed by the emplacement of a composite late- to post-tectonic calc-alkaline batholith with spectacular intrabatholitic dyke swarms between 615 and 590 Ma ago (Ducrot et al., 1979). The zone was subjected to continuous uplift, leading to unroofing and outpouring of flat-lying Nigritian rhyolites, followed by the intrusion of alkaline ring-complexes. ^{39}Ar-^{40}Ar whole-rock and U-Pb Concordia data on a sheared granite yielded 560 and 540 Ma, respectively, and suggest late movement along the edge of the granulite nappe (Lancelot et al., in press).

In the eastern Pharusian branch N—S-trending open folds of very high structural level in the north gradually pass southwards into tight folds with associated greenschist facies metamorphism, grading to amphibolite facies and migmatization around synkinematic granites (Lelubre, 1952; Gravelle, 1969). U-Pb zircon dating of such granites (Picciotto et al., 1965) has given recalculated ages between 647—620 Ma. Later folds and thrusting locally occurred after the deposition of molasse-type sediments. Late- to post-tectonic granites yield ages in the range 600—550 Ma (Boissonnas, 1973).

The molasse of the Pharusian belt is represented by the "Série pourprée" which partly has a Lower Cambrian age (Caby, 1970; Clauer, 1976).

Reactivation of the polycyclic central Hoggar—Aïr domain and the Tiririne linear intracontinental belt

The 4°50'E shear zone, which is a transcontinental lineament that can be traced to Benin, marks the eastern limit of the Pharusian belt. The amount of lateral movement along this major strike slip fault is unknown. Neither lithological nor structural correlations between the Pharusian belt and the adjoining central polycyclic Hoggar—Aïr domain are possible, except for the latest events related to collision with the West African craton. The Pan-African in this domain is characterized by a greenschist metamorphic overprint, intense deformation localized along N—S schist belts, large-scale granite emplacement and lateral movement along shear zones.

The linear Tiririne intracontinental belt (Bertrand et al., 1978) is situated along the eastern margin of the polycyclic Hoggar—Aïr domain and provides a model for the reactivation process (Fig. 16-1). Here two different basement units (polycyclic central Hoggar—Aïr and eastern Hoggar—Ténéré domain) face each other along the 8°30'E shear zone which corresponds to a dextral transcurrent fault. The orogenic event at 725 Ma, which affected the eastern Hoggar—Ténéré domain (see page 419), was followed by deposition of the flat-lying molasse-type Tiririne Group, containing sills of diorite, diabase and gabbro intruded around 660 Ma ago. A trough filled by more than 8000 m of sediments developed in proximity to the shear zone and subsequently evolved into a linear fold belt with greenschist metamorphism locally reaching amphibolite facies grade with anatexis. A late-tectonic granite yielded a U-Pb Concordia age on zircon of 604 ± 13 Ma and dates this event (Bertrand et al., 1978). Reactivation of the polycyclic central Hoggar—Aïr basement to the west (Issalane) is indicated by: (1) a westerly dipping cataclastic to mylonitic new foliation in an area up to 10 km wide, displaying a strong stretching lineation at high angle to the 8°30′E shear zone; (2) a zone of recumbent folds, and (3) emplacement of a 585 ± 14 Ma old high-level batholith (Bertrand et al., 1978). The model proposed for this ensialic linear belt involves westward underthrusting of the eastern Hoggar—Ténéré domain with its overlying Tiririne formation, producing local crustal thickening, anatexis and generation of granites at depth and subsequent strike-slip movement along the shear zone.

The narrow N—S-trending belts of low-grade metasediments of assumed Upper Proterozoic age associated with predominantly acid volcanics and occurring within the central Hoggar—Aïr domain (Fig. 16-1) could represent diachronous equivalents of the Tiririne belt and, by analogy, they may also be associated with early underthrusting, causing crustal thickening as well as lateral shear. In this connection re-examination of some of the major, now vertical, shear zones indicates that they have had a complex history and bear the imprint of an early deformation when they behaved as thrusts (A.M. Boullier, pers. commun., 1979). McCurry and Wright (1977) regard such belts of transcurrent faulting in Nigeria as part of a fossil collision zone between two continental blocks. Abundant Pan-African granitoids about 650—600 Ma have invaded large areas of the central polycyclic Hoggar—Aïr domain and have played an important role in reactivation, locally producing low-pressure—high-temperature metamorphism superimposed on earlier mineral assemblages (Bertrand, 1974; Vitel, 1979). Although it is difficult to relate these granitoids to subduction processes there is an overall zonality in the Touareg shield: diorites and granodiorites which may be related to a subduction zone predominate in the Pharusian belt, whereas adamellite and K-rich varieties abound in the polycyclic central Hoggar—Aïr domain. Although isotopic studies are still in their infancy, preliminary results indicate low initial Sr ratios for diorites in the Pharusian domain (Davison, 1980), whereas values ranging between 0.707 and 0.72 have been obtained in the central polycyclic domain (Bertrand, 1974; Vialette and Vitel, 1979) and in Nigeria (Van Breemen et al., 1977), suggesting that crustal fusion related to crustal thickening may have played an important role.

Attention has already been drawn to the striking similarities between the shear pattern of the Touareg shield, characterized by branching and sub-parallel shear zones (Bertrand and Caby, 1978; Ball, 1980), and the Cenozoic fault pattern of northeastern Asia, produced by the collision of

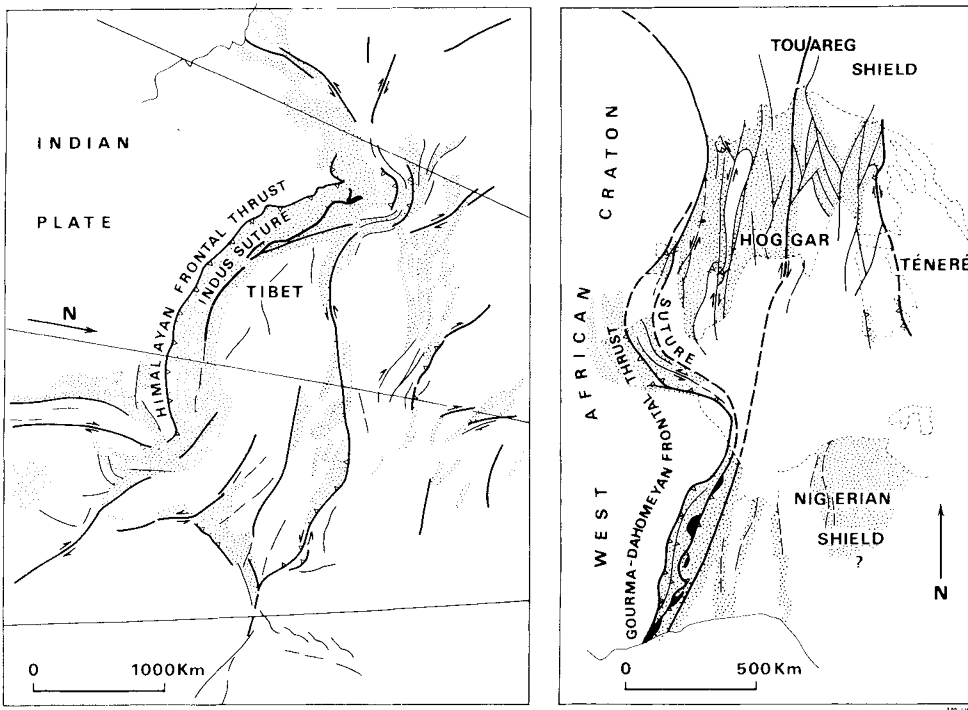


Fig. 16-8. Comparison between Cenozoic to Recent fault-thrust pattern of NE Asia (after Molnar and Tapponnier, 1975) and Pan-African tectonics in northwestern Africa. Dotted areas represent Tertiary and late Pan-African deformation, respectively.

India with Asia (Molnar and Tapponnier, 1975; Fig. 16-8). We suggest that the Pan-African reactivation of the central polycyclic Hoggar is related to collision with the West African craton, and the present surface may represent the deep structural level of the same processes which are now at work in northeast Asia.

CONCLUSION

Evidence from many earth science disciplines suggests the existence of a Wilson cycle ending around 600 Ma ago with collision and suturing between the passive continental margin of the West African craton and the active margin of the Touareg shield to the east.

The geochronological data, although still fragmentary, show the time range (900—550 Ma) for Pan-African events in the central Sahara to be comparable to that of the Arabian—Nubian Shield (Greenwood et al., 1976; Kröner, 1979). However, one must distinguish tectonically quiet periods of island-arc and Cordilleran development and periods of intense deformation related to continental collision within the above time range. The large amount of predominantly calc-alkaline magmatic rocks generated in this time interval corresponds to a major period of crustal accretion which led to the cratonization of Africa.

The long pre-collisional Cordilleran-type evolution in the western part of the shield and the multistage collisional history, involving stabilization of the eastern Hoggar-Ténéré domain at around 725 Ma, N—S collision in the western Pharusian branch at around 700 Ma ago and, finally, E—W collision against the West African craton at around 600 Ma, suggest that the Touareg shield is composed of an amalgam of plates. Thus the collision of a cold rigid craton in the west against an amalgam of relatively hot and more ductile micro-plates in the east produced extensive reactivation over a width of several thousand kilometres in northeastern Africa.

The most pertinent argument in favour of Pan-African active subduction processes operating in this part of Africa, besides the presence of rock assemblages characteristic of island-arc and continental Cordilleran environments and the results of geochemical investigations on associated andesites, is the existence of distinctly paired metamorphic belts. The high-pressure—low-temperature eclogitic rocks thrust onto the passive continental margin display mineral assemblages which strongly indicate that they have formed in a subduction zone and were subsequently thrusted.

It has been pointed out that the reactivation of the entire Touareg shield as a consequence of collision with the West African craton is marked by greenschist metamorphic overprinting of older gneissic terrains, intraplate deformation in linear belts with pronounced E—W shortening accompanied by crustal thickening, and generation of granite. If the analogy which we draw with the tectonic model proposed for the Himalayan—Tibetan region (Molnar and Tapponnier, 1975) is correct, the amount of lateral displacement along shear zones in the central Touareg shield may well exceed the dimensions of the shield.

The basification process through intrusion of ultrabasic and basic bodies of tholeiitic affinities, followed by calc-alkaline magmatism, is thought to have played an important role in the history of the central Sahara.

Our conviction is that detailed studies of the Pan-African edifice, beautifully exposed at a deep structural level in the Sahara region, will throw new light on the processes that were at work in ancient active continental margins, on the nature and degree of intracontinental deformation and, more generally, on the geochemical evolution of the crust in late Precambrian times.

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Chapter 17

THE LATE PROTEROZOIC OPHIOLITES OF BOU AZZER (MOROCCO): EVIDENCE FOR PAN-AFRICAN PLATE TECTONICS

M. LEBLANC

ABSTRACT

The Bou Azzer ophiolites are amongst the first Precambrian ophiolites ever to have been described. They possess the typical characteristics of modern ophiolites, namely:

(1) An assemblage composed of an ultramafic complex with a tectonite fabric, a cumulate gabbro complex, a basic volcanic complex with pillow lavas and dyke swarms, trondjhemite and quartz-diorite intrusions and an overlying volcano-sedimentary series.

(2) The presence of chromite pods and asbestiform chrysotile in the ultramafic complex and copper sulphide deposits in the basic lavas.

(3) An allochthonous tectonic setting (obduction onto continental margin).

The ophiolites of Bou Azzer are inferred to be fragments of Upper Proterozoic oceanic crust (788 ± 9 Ma) on the northern margin of the West African craton. The ophiolites were obducted onto the craton with a north to south translation, during the major Pan-African deformation phase (dated at about 685 Ma in the Bou Azzer region). One can thus consider that plate-tectonic mechanisms existed in northwest Africa during Upper Proterozoic times.

INTRODUCTION

The most conclusive evidence for a plate-tectonic model is often the presence of ophiolites. They are interpreted as sheets of oceanic crust obducted onto continental margins and testify to a sea-floor spreading stage followed by collision (Dewey and Bird, 1970). The first complete ophiolitic complex described in the Proterozoic was that of Bou Azzer (Leblanc, 1972a, 1975, 1976a), located in the Pan-African segment of the Anti-Atlas, Morocco (Fig. 17-1).

After giving an outline of the geological setting I shall describe the Bou Azzer ophiolitic complex and compare its petrography and its metallogeny with that of modern ophiolites. Finally I shall discuss its tectonic setting and how this allows us to propose an Upper Proterozoic plate-tectonic model for the evolution of the northern margin of the West African craton.

THE BOU AZZER OPHIOLITIC COMPLEX

Regional geology

The Precambrian basement of the Anti-Atlas (Choubert, 1947) comprises

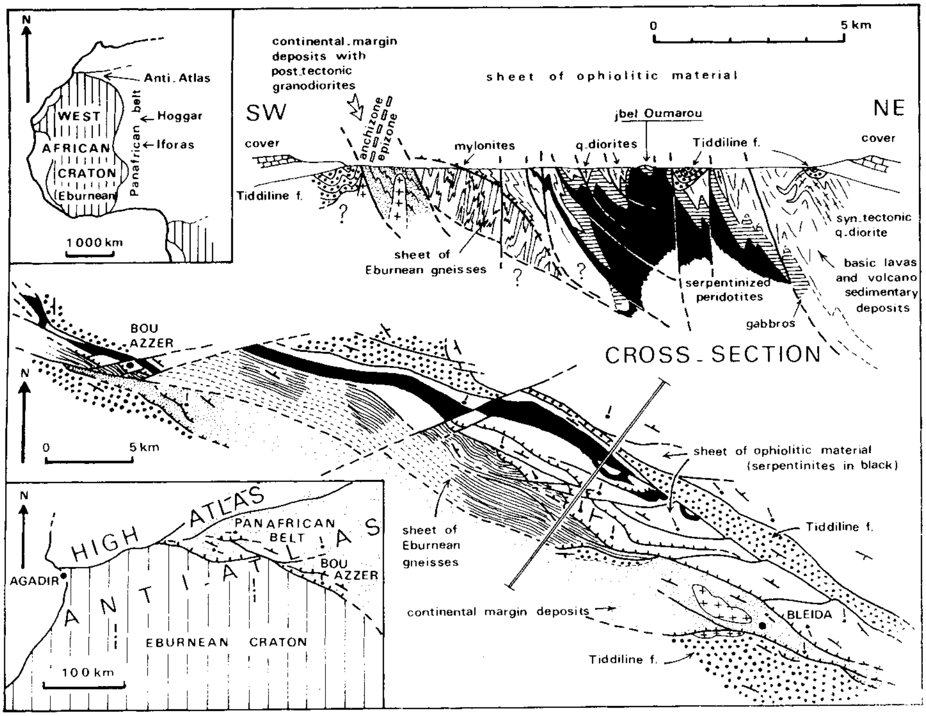

Fig. 17-1. Schematic map of Bou Azzer (cover rocks removed) with a cross section and inset maps for localization. The fold axis belong to the major tectonic event B_1 post-dated by post-tectonic granodiorites and by deposition of the Tiddiline formation.

two major structural units: (a) the northern border of the West African Eburnean craton (2000 Ma, Charlot, 1978) to the southwest; and (b) a segment of a Pan-African orogenic belt (680—580 Ma) to the northeast (Caby, 1970; Clauer, 1974, 1976; Leblanc, 1975; Charlot, 1978). The Bou Azzer district (Fig. 17-1) is situated along the contact between these two units in the central part of the Anti-Atlas (Major Anti-Atlas Fault, Choubert, 1947). The structural trend of the district is WNW—ESE. The Precambrian basement is exposed over an area of 70 km × 10 km, elongated parallel to this structural grain.

The gneissic rocks of the *Eburnean craton* (Choubert, 1960, Charlot et al., 1970) are exposed 20 km west of Bou Azzer and are overlain unconformably by shelf sediments (quartzites and stromatolitic limestones) which were folded during the Pan-African orogeny.

On the southern border of the Bou Azzer district the Bleida zone, which is limited on both sides by major thrusts, is interpreted as *an ancient continental margin* (Leblanc and Billaud, 1978). This zone of subsidence is characterized by a 2 km thick rock sequence including quartzites and stromatolitic limestones, successively overlain by basic lavas, the thick volcano-sedimentary formation of Bleida and, finally, arkosic sandstones. Quartz-diorites and quartz-gabbros intrude the Bleida formation.

The major Pan-African deformation (B_1, Leblanc, 1973b) produced recumbent folds with an axial planar slaty cleavage and thrusts which involved a slice of presumed Eburnean gneisses. Metamorphic clay minerals (schistosity S_1) progressively appear northwards in the Bleida zone and have been dated at 685 ± 15 Ma (Rb-Sr on metamorphic illites, Clauer, 1976). The whole Bou Azzer complex was affected by greenschist facies metamorphism.

The post-tectonic granodiorite of Bleida is dated at 615 ± 12 Ma (U-Pb, Ducrot and Lancelot, 1978). The overlying Tiddiline Formation, which rests on a peneplained surface, consists of detrital deposits (conglomerates, greywackes, a tilloïd[1] and feldspathic sandstones) preserved in grabens. The greywackes are attributed to insular arc-system turbidite deposits with metalavas (spilites and keratophyres) and associated intrusions of plagiogranite. Ultramafic horst blocks were subjected to weathering at the end of this period. The late Pan-African brittle deformation (B_2) is post-dated by the Infracambrian and Palaeozoic cover (Choubert, 1958; Leblanc, 1973a). These strata rest unconformably on the Precambrian basement and have escaped most of the Hercynian tectonics (Leblanc, 1972b). The base of this cover sequence is composed of the volcanic Ouarzazate Formation (580 ± 15—565 ± 20 Ma, U-Pb, Juery et al., 1974; Juery, 1976) and of the sedimentary Adoudounian Formation (Boho Adoudounian lavas dated at 534 ± 10 Ma, U-Pb, Ducrot and Lancelot, 1977).

At the scale of the Anti-Atlas there remain several correlation problems which need complementary studies: for example, the Askaoun granodiorite (Siroua) which intrudes a greywacke sequence resembling the Tiddiline Formation is dated at 699 ± 10 Ma (Rb-Sr, Charlot, 1978). Nevertheless a coherent succession of Pan-African events has been obtained at the scale of the Bou Azzer district (Leblanc and Lancelot, 1980).

The Bou Azzer mafic-ultramafic complex

A number of good exposures in different parts of the Bou Azzer district have allowed a coherent stratigraphy of the complex to be reconstructed (Leblanc, 1972a, 1975) in spite of tectonic contacts and upright folds (Fig. 17-1). The Bou Azzer mafic-ultramafic complex is 4 to 5 km thick. It comprises from bottom to top (Fig. 17-2): (a) ~ 2000 m of serpentinized peridotites; (b) ~ 500 m of ultrabasic and basic cumulates (layered gabbros); (c) large stocks of quartz-diorites; (d) ~ 500 m of basic lavas with pillow lavas; and (e) ~ 1500 m of a volcano-sedimentary series.

The serpentinites (40% of the total complex) are derived from dunites

[1] Tillite-like rock of uncertain origin.

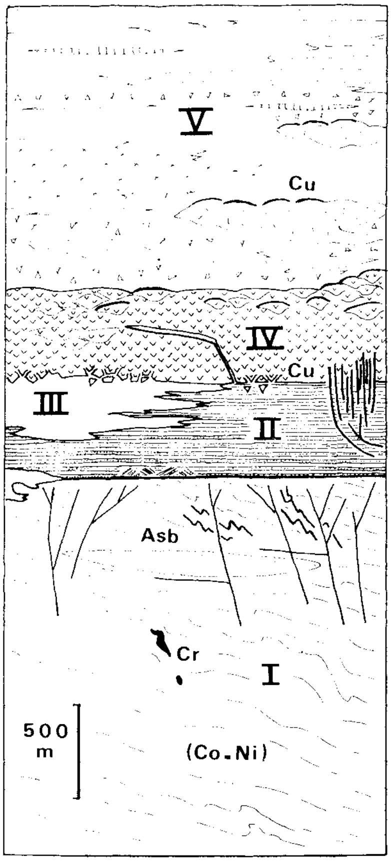

Fig. 17-2. Schematic section through the ophiolites of Bou Azzer showing: I = serpentinized peridotites with their foliation and various dyke generations; II = layered gabbros; III = quartz-diorite stock; IV = basic lavas with magmatic breccias, leucogabbroic sill and dyke swarm at their bottom and pillow lavas at their top; V = volcano-sedimentary series comprising lavas, greywackes (triangles) and tuffs of basic, spilitic and keratophyric composition and horizons of jaspilites, calcareous tuffs, jaspers, quartzites and limestones. The mineralizations within the Bou Azzer ophiolites are chromite pods (*Cr*), asbestiform chrysotile (*Asb*), copper sulphides (*Cu*) and cobalt-nickel arsenides (*Co-Ni*).

and harzburgites showing a tectonic fabric. Their average Ni content (2650 ppm) and the geochemistry of their accessory chromites are similar to those from mantle peridotites of modern ophiolites (Irvine, 1967; Leblanc et al., 1980). The serpentinization (lizardite, chrysotile) was accomplished during a multi-stage history which terminated before the end of the major Pan-African deformation. The rodingitized basic and microdioritic dykes interfolded and boudinaged within the serpentinites are cut by quartz-microdioritic dykes.

The layered gabbros (10—15%) are locally discordant upon a pre-existing foliation in the serpentinized peridotites (primary discontinuity). The igneous layering is disturbed by magmatic breccias and sometimes obliterated by pegmatitic recrystallization. Near the base ultramafic cumulates are predominant and exhibit decimetre-scale layering with dunite-wherlite-gabbro sequences. Towards the top hornblende microgabbros become the preponderant facies, with associated quartz-microdioritic horizons. Throughout this gabbroic complex sills and dykes of diabase, keratophyre and trondhjemite are quite frequent.

The quartz diorite stocks (10%) occur between the layered gabbros and the overlying basic lavas. Their maximum thickness reaches 500 m compared with several kilometres of lateral development, i.e. they display a stratiform shape. Diffuse interfingering of the stocks with the layered gabbros and association with quartz-gabbros are observed while intrusive contacts are evident with the basic lavas (agmatite). These stocks are considered part of the ophiolites. Nevertheless, other quartz diorite massifs have intruded the upper part of the ophiolite complex during the major B_1 Pan-African deformation event and caused contact metamorphic aureoles.

The basic lavas overlie the layered gabbros with a gradational contact through a massive microgabbro, or with a sharp contact marked by magmatic breccias with a leucogabbroic matrix. The top of the lava succession usually contains pillow lavas and hyaloclastites. At the base diabase, leucogabbro and quartz-diorite dykes and sills are abundant.

Injections of diabase, microgabbro and keratophyre material within the layered gabbros develop upward from sills to dykes. The dykes have an orthogonal disposition with regards to the igneous layering and are often chilled at their margins. In the basic lavas above, the dykes (20 cm—1 m thick) can locally occupy up to 70% of the total volume and thus constitute a sheeted dyke complex.

The volcano-sedimentary series contains again pillow lavas, often spilitic and (quartz) keratophyric in composition. Greywackes and tuffs comprise the major part of the sedimentary rocks along with associated jaspilites and calcareous tuffs. Near their base the greywackes locally display a conglomeratic facies with fragments of gabbro, microgabbro, basic lava and also quartz-diorite. Arkosic sandstones appear in the uppermost part of this series.

Evidence for ophiolites at Bou Azzer

The assemblage described above is very similar to Mesozoic and Cenozoic ophiolite assemblages as regards both its thickness and the spatial arrangement of the different petrographic units (Penrose Conference, Geotimes, September 1972).

The dyke swarms are poorly developed in comparison to the sheeted dyke complexes characteristic of the Troodos and Hatay massifs. But many modern ophiolites (e.g. Antalya, Pindos, New Guinea) also have no sheeted dyke complex.

Located between gabbros and basic lavas and reworked within the greywackes of the volcano-sedimentary series, the quartz-diorites are part of the Bou Azzer ophiolites. The relative abundance of these quartz-diorites (10%) and keratophyres (5%) is a particular feature of the Bou Azzer complex and contrasts with the observation that leucocratic rocks are generally present only in small volume (less than 10%, Coleman and Peterman, 1975) in most modern ophiolitic complexes. However, the country rocks in the southern part of the Troodos sheeted dyke complex are mainly diorites, quartz-diorites and granophyres (Desmet et al., 1980) and there are quartz-diorite stocks, one hundred metres thick, at the top of the gabbros within the North Oman ophiolites (Allemann and Peters, 1972). Furthermore, keratophyre lavas supersede basic lavas in some ophiolitic sequences where leucocratic intrusive rocks are abundant (Bayley et al., 1970; Jackson et al., 1975). These acid rocks might represent the most differentiated part formed during the magmatic evolution along a mid-oceanic ridge (Coleman and Peterman, 1975; Dixon and Rutherford, 1979). An alternative interpretation of the Bou Azzer keratophyres may be that these rocks are related to an island-arc-type development which was superimposed on the ophiolite complex (Leblanc, 1975).

When plotted on an AFM diagram the representative rock types show a similar trend to that of the Troodes and Pindos ophiolitic complexes (Fig. 17-3). This "ophiolitic trend" differs from that of a tholeiitic trend by displaying a higher Mg/Fe ratio. Major elements are of dubious validity due to alteration and the transition trace elements (Ti, Cr, Ni, V) have therefore been used (Table 17-1) by reference to the values of modern basic lavas (Andriambololona, 1978). The basic lavas of Bou Azzer appear to be similar to mid-ocean tholeiitic olivine basalts. Furthermore, when plotted on the Ti/Cr vs Ni diagram (Beccaluva et al., 1979), they lie within the ocean-floor tholeiite field (Fig. 17-4) together with the sub-effusive basic members ($SiO_2 < 55\%$) from the quartz-diorite assemblages. The trondhjemite dykes and the massive quartz-diorite of the Bou Azzer Mine display a low $100\,K_2O/(K_2O + Na_2O)$ ratio ($< 5\%$) similar to that of the oceanic plagiogranites from the Cyprus ophiolite (Coleman and Peterman, 1975). Nevertheless, some Bou Azzer quartz-diorites are a little more potassic. On the basis of these data it is suggested that the Bou Azzer ophiolite may have been generated at a mid-oceanic ridge (Type I — ophiolite of Rocci et al., 1975), but it also displays some characteristics of the type II — ophiolite formed in or near an immature island arc (abundance of diorite, trondhjemite and tonalite). Additional data on trace elements (REE, Y, Zr) would be necessary to confirm the calc-alkaline affinities of the acidic lavas and of some dykes (Church, 1980) from the Bou Azzer ophiolites.

On the continental margin (Bleida zone) the basic lavas consist of Ti-rich

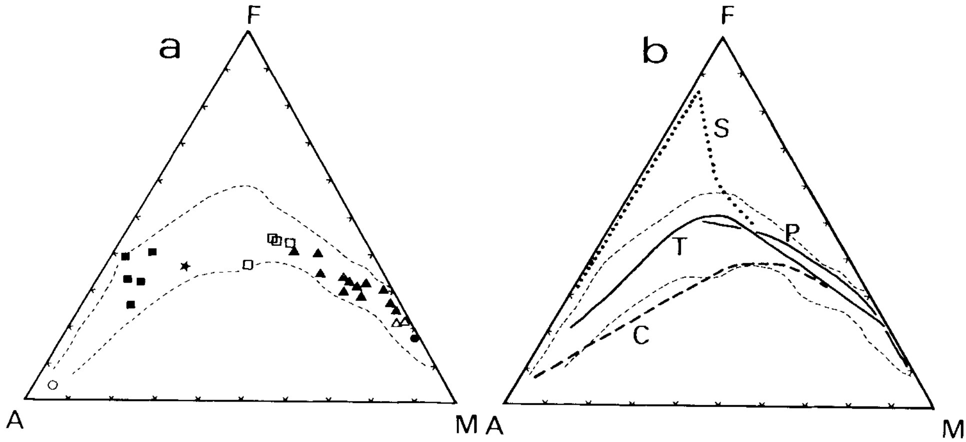

Fig. 17-3. AFM diagrams showing (a) the domain of 90 Bou Azzer ophiolitic rocks. The 26 data points plotted here represent the continuous cross-section of Jebel Oumarou (Fig. 17-1): serpentinized peridotites (black dots), ultramafic cumulates (white triangles), layered gabbros (black triangles), basic lavas (white squares), quartz diorite (star), keratophyres and quartz-keratophyres (black squares) and trondhjemite (circle). This domain displays an approximately linear repartition, especially in the right part of the diagram (b) which conforms to the trends of typical ophiolites: T = Troodos (Thayer, 1967); P = Pindos (Parrot, 1967). It differs from the (C) calc-alkaline trend (Hess, 1960) and even more again from the (S) Skaergaard trend (Wager and Deer, 1939).

TABLE 17-I

Average values for some basic lavas from the Bou Azzer district compared with modern volcanic series (Andriambololona, 1978). Number of analyses are given in brackets

	TiO_2 (%)	FeO^+/MgO	Cr(ppm)	Ni(ppm)	V(ppm)
From the Jebel Oumarou (Fig. 17-1)(4)	0.7	1.23	390	160	275
From the whole Bou Azzer complex (16)	1.04	1.71	270	120	133
From the continental margin (4) (Bleida zone)	1.92	1.8	270	140	138
Ref. continental tholeiites	1.8—2.3	0.5—2.	300—1300	180—1100	—
Ref. abyssal ol-tholeiites	0.7—1.5	0.6—1.27	300—1000	100—420	200—290
Ref. abyssal pl-tholeiites	0.8—1.4	0.8—1.4	190—390	80—140	200—300
Ref. calc-alkaline margin basalts	1.3	1.48	140	70	—
Ref. calc-alkaline arc basalts	0.97	1.73	70	35	255
Ref. arc tholeiites	0.88	1.79	50	30	330
Ref. alkaline basalts	2.1—3.2	0.7—1.7	110—700	90—500	240—280

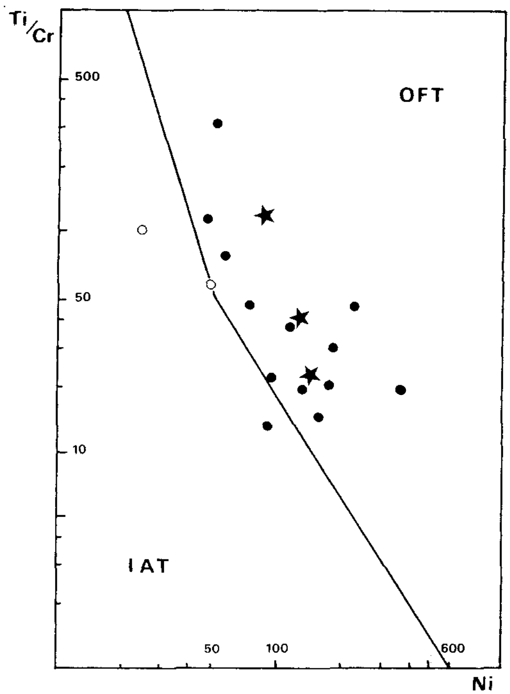

Fig. 17-4. Ti/Cr versus Ni diagram for basic rocks ($40\% < SiO_2 < 55\%$) of the Bou Azzer ophiolite. Black dots: basic lavas (Leblanc, 1975); circles: basic dykes (Church, 1980); black stars: microdiorite related to quartz-diorite (Leblanc, 1975). The oceanic-floor tholeiite field (OFT) and the island-arc tholeiite field (IAT) have been defined by Beccaluva et al. (1979) on the basis of 262 analyses from present-day volcanic rocks. The basic lavas from the continental margin (Bleida zone) lie also in the OFT field (not represented here).

tholeiites, intermediate between abyssal and continental tholeiites, and are typical of rifting and spreading stages (Strong and Williams, 1972).

Small pods of chromitite exist within the serpentinized peridotites. If we examine the diagram Cr/(Cr + Al) vs Mg/(Mg + Fe) the representative points of massive chromite (microprobe analyses) lie within the domain which is characteristic of ophiolite podiform chromite deposits (Irvine, 1967; Leblanc et al., 1980).

Elsewhere in the serpentinites chrysotile asbestos has been exploited which is especially developed along the borders of a diallagite horizon and those of tectonized serpentinite massifs.

The famous Bou Azzer cobalt orebodies (50,000 tons Co metal) contain cobalt arsenides in a quartz carbonate gangue and are located on the margins of the serpentinite massifs. Their genesis is complex and for the most part they post-date the tectonic emplacement of the ophiolite (Leblanc, 1975). The quartz-carbonate gangue is derived from the silica and carbonates of the weathering carapace which developed above the ophiolites during the late Upper Proterozoic. This carapace was generated in small endoreic basins on a peneplained surface of weathered basic rocks and serpentinite enriched in Ni (5500 ppm) and Co (260 ppm). Concentration of cobalt probably occurred contemporaneously with the leaching of the serpentinites (dissolution of magnetite which is the main cobalt host mineral) which released cobalt and nickel. The latter is more soluble and was removed from the system whilst cobalt was concentrated in situ (cavities filled with silica and Mn-Co hydroxides, pockets of red clay containing Fe-Co arsenides). These cobalt concentrations have been strongly remobilized during the late Pan-African and the Hercynian phases of deformation. This apparently unique type of cobalt deposit is clearly related to the serpentinites of the Bou Azzer ophiolite.

Some occurrences of copper sulphides are found in the keratophyric and spilitic rocks at the top of the Bou Azzer ophiolite. This mineralization is poorly developed and without large lateral extent. On the continental margin (Bleïda area) massive sulphide orebodies are interbedded within the volcano-sedimentary formation of Bleida which is interpreted as a lateral equivalent of the upper part of the Bou Azzer ophiolitic complex (Leblanc and Billaud, 1978). The Bleida mineralization (2,500,000 tons at 8% Cu) was controlled by sedimentation of silty greywackes, underlain by a jasper marker bed. This mineralization is related to a discrete, acidic volcanic event with associated quartz-diorite intrusions, following the basic volcanism developed on the continental margin.

In the Bou Azzer ophiolitic complex copper mineralization has also been noted at the bottom of the basic lavas; in the breccias and the basic dyke swarms below the pillow lavas. This is a typical position for ophiolite sulphide mineralization but here it is not of economic concentration. Finally, there are some indications of copper at the bottom of the gabbroic cumulates and within a chromite- and sulphide-rich horizon (chalcopyrite, bornite, linnaeite) in the serpentinites.

The Bou Azzer ultramafic—mafic complex is an ophiolite with respect to its original geometry, petrography, geochemistry and metallogeny. Some problems are still unresolved, particularly the petrological evolution and the initial structural setting, but these problems are those of all ophiolites.

TECTONIC SETTING OF THE BOU AZZER OPHIOLITE

The southern limit of the Pan-African segment of the Anti-Atlas is in part demarcated by the Bou Azzer ophiolite which may mark a major suture zone. Unfortunately, it is difficult to make an accurate reconstruction of the regional tectonic setting: the Pan-African basement outcrops are few and far between and the main part of the belt is hidden to the north under the palaeozoic and Mesozoic cover of the High Atlas domain (Fig. 17-1).

From seismic data the crustal thickness is the same for the Eburnean craton and the Pan-African segment of Southern Morocco (Hatzfeld, 1978). Gravimetric data (Van den Bosch, 1971) show no linear, strong positive anomalies but only a weak anomaly in the Bou Azzer district which may correspond to an unrooted and a low-angle north-dipping ultramafic—mafic body (A. Lesquer, pers. commun., 1975). This is in agreement with obduction of slabs of oceanic crust on a continental margin.

The ophiolite has been dismembered into several sheets which have been stacked up against the continental margin (Bleida) which is also deformed. The tectonic emplacement took place during the major Pan-African deformation (B_1) and was accompanied by Barrovian-type greenschist metamorphism.

A large thrust sheet of gneisses of assumed Eburnean age delimits the junction between the ophiolite and the continental margin (Fig. 17-1). The gneisses, with a vertical foliation, are covered by subhorizontal mylonites representing the only high strain zone in the Bou Azzer district. Structurally overlying these mylonites is part of the ophiolitic sequence which is locally overturned (younging directions being derived from pillow lavas and graded-bedding in the greywackes).

The recumbent B_1 folds have an axial planar slaty cleavage and their axes trend NNW with their axial planes now upright due to later events. This geometry would suggest thrusting towards the WSW, obliquely onto the margin of the West African craton (WNW—ESE). The lineation of the mylonites is trending N—S and the stretching direction observed affecting pyrite crystals from the Bleida copper deposit (Billaud, 1977; Leblanc and Billaud, 1978) is almost parallel to the B_1 axes. The orientation of slip planes and rotation of pyrite crystals (sigmoidal crystallisation of quartz-chlorite-sericite in the pressure shadow zones) indicate a southerly movement towards the West African craton. The same movement has been found in the folded and mylonitized cover of the West African craton some 25 km west of Bou Azzer (Horrenberger, 1973). The tectonic emplacement of the ophiolites against the border of the West African craton was accompanied by a complete serpentinization of the ultrabasic rocks. Injections of serpentinite along the thrust planes may be smeared out some 2 km away from the serpentinite massifs and must have facilitated movement along these planes. The actual erosion level is almost the same as that which existed at the end of the major tectonic phase (B_1) and corresponds to moderate post-Pan-African uplift. This could be the reason why the ophiolitic sheets are still preserved.

The ultimate Pan-African deformation B_2 (Leblanc, 1973b) produced WNW—ESE-trending, upright chevron folds with axial planes and fracture cleavage parallel to the border of the West African craton. Late brittle structures are manifested in the form of reverse (reactivated B_1 thrust planes) and transcurrent faults which are mainly sinistral (WSW—ENE). This deformation has determined the actual geometry of the basement in the Bou Azzer district (Fig. 17-1).

In conclusion the Bou Azzer ophiolite occurs in a allochthonous setting and has been obducted towards the south onto a continental margin, a typical structural position for ophiolites (Gass et al., 1975).

DISCUSSION

The Bou Azzer ultramafic—mafic complex represents an Upper Proterozoic ophiolite (Leblanc, 1972a) and is interpreted as a fragment of oceanic crust, obducted during the major Pan-African deformation (Caby and Leblanc, 1973; Leblanc, 1975, 1976a). Other Upper Proterozoic ophiolites have subsequently been described in Anglesey, Wales (Baker, 1973; Wood, 1974; Thorpe, 1978); Saudi Arabia (Neary, 1975; Bakor et al., 1976; Shanti and Roobol, 1979) and Timetrine, Mali (Leblanc, 1976b) but no certain ophiolites older than Upper Proterozoic have so far been identified.

It is now interesting to speculate whether the Upper Proterozoic oceanic crust might have been generated and then destroyed by processes similar to modern plate tectonics.

Spreading of oceanic crust at Bou Azzer

The sialic continental margin (Bleida) is overlain by deposits with lateral facies and thickness variations indicating a gradually subsiding basin and an open sea to the north (Leblanc and Billaud, 1978). The ocean opening can perhaps be dated using the age of 788 ± 9 Ma which is that of basic volcanism on the continental margin (Rb-Sr whole-rock isochron related to a metamorphic aureole of basic dykes, Clauer, 1976). This may well have been coeval with continental rifting.

If one assumes that the major Pan-African deformation was that of a body translation without significant rotation one may reconstruct the original orientation of the palaeo-ridge. Using the model of Juteau et al. (1976) the fabric of the cumulate gabbros and tectonite peridotites may delimit the orientation of ocean ridges. In the favourable case of the Jebel Oumarou (Fig. 17-1) and after reorientation of the igneous layering in the gabbros to the horizontal we can observe: (a) E—W dykes dipping steeply north; (b) N—S lineation in the gabbros; (c) E—W foliation dipping north with a N—S chromite lineation in the peridotites. From the above observations it would seem that the oceanic crust of Bou Azzer was situated to the south of an E—W palaeo-ridge and to the north of the northern continental margin of the West African craton.

Only indirect and debatable arguments are available for the estimation of the size of the proposed palaeo-ocean. If we accept 50 Ma long spreading considering that the ocean opened and closed in about 100 Ma (788—685 Ma), and furthermore assume a slow and constant spreading half rate of 0.5 cm/a, as suggested by the Ti-content of the basalts (Nisbet and Pearce, 1973), we obtain a 1000 km wide oceanic domain. The slow spreading rate advocated here should explain the relative abundance of leucocratic rocks by a long, subcrustal magmatic differentiation process (Coleman and Peterman, 1975) and the relative scarcity of sheeted dykes (Sleep, 1975).

However, in this hypothesis the starting age, the duration and the rate of spreading are broadly unknown. The one field evidence is the presence of arkoses near the top of both the continental margin sequence and the ophiolite sequence, suggesting that these two domains were at one time in close proximity. Furthermore, pretectonic quartz-diorite and thick volcano-sedimentary series are found in these two domains (Leblanc and Billaud, 1978).

The Bou Azzer ophiolite is presumed to have been a fragment of N—S-spreading Upper Proterozoic oceanic crust which developed between the West African craton and a hypothetical northern craton. The size of this palaeo-ocean remains unknown but I suggest, without conclusive evidence, a marginal sea rather than a large Atlantic-type ocean. A similar setting has been proposed for the Tethyian ophiolites (Church and Stevens, 1971; Mercier et al., 1975; Brunn, 1980; Hawkins, 1980).

A subsequent island-arc-type evolution has been suggested to explain part of the quartz-diorites and acidic lavas (Leblanc, 1975).

Correlations are available with the Upper Proterozoic ophiolites of the North Atlantic domain (Strong, 1979) as well as the sutured belt of mafic rocks along the eastern edge of the West African craton (Caby and Leblanc, 1973; Caby et al., this volume) among which dismembered ophiolites have been described (Leblanc, 1976b).

There are probably all transitions from true ophiolites to continental rifting zones invaded by mafic and ultramafic material as assumed by Dewey (1980).

Bou Azzer ocean closing

There is no convincing evidence of pre-tectonic obduction by gravity-sliding (e.g., olistostromes, mélanges). The Bou Azzer ophiolite has been obducted southerly onto a sialic continental margin during the major Pan-African deformation event (685 ± 15 Ma). The thrust contact of the ophiolite sequence with presumed Eburnean gneissic basement is demarcated by subhorizontal mylonites. All tectonic movements were from north to south, in agreement with the contemporaneous (around 700 Ma) early suturing along the eastern edge of the West African craton (Bertrand and Caby, 1978).

The Pan-African segment of the Anti-Atlas is characterized by the absence of crustal thickening, high-grade metamorphism, syntectonic anatectic granitization and important post-orogenic uplift, and this implies some form of moderate collision without continental overlap. The Anti-Atlas may have been a re-entrant along a lobate collision boundary where obducted ophiolites have been preserved. The enrootment of the corresponding oceanic crust — e.g., the suture zone — should be found in the inner part of the belt somewhere towards the north.

The precise mechanism of obduction is not exactly known but in modern cases it is generally related to subsequent subduction (Gass et al., 1975). The slip zone may dip either away from, or under the continental margin and may produce a calc-alkaline magmatism. At Bou Azzer the development of calc-alkaline volcanism (Tiddiline Formation) with associated greywackes and plagiogranite ($K_2O > 1.5\%$) above the obducted ophiolite suggests a south-dipping subduction zone under the West African craton (Leblanc and Lancelot, 1980).

The ultimate shortening (B_2) probably took place when the West African craton collided directly with the hypothetical northern craton. The subsequent post-collisional volcanism of the Ouarzazate formation (580—560 Ma) produced calc-alkaline lavas (Boyer et al., 1978), associated with tholeiitic basalts, trachytes, abundant ignimbrites, hypabyssal granites and Cu-Mn-Mo mineralizations. This calc-alkaline post-collisional volcanism was generated after a time delay not exceeding that (25 Ma) observed for the Alps (Dal Piaz et al., 1979). Unfortunately, the narrow width of the Anti-Atlas (Fig. 17-1) does not allow to reconstruct a magmatic zonality from which the dip of the supposed subduction zone could be deduced.

CONCLUSION

The Bou Azzer ophiolite provides proof for the assumption that mechanisms similar to those of modern plate tectonics have operated during Upper Proterozoic times: (1) opening of a presumed marginal oceanic domain by spreading along an E—W constructive plate margin on the northern edge of the West African craton; (2) closing of the oceanic domain during a polyphased collision event, corresponding to N—S shortening with obduction of oceanic crust (ophiolite) onto the continental margin and subsequent subduction under the craton; and, finally, (3) an ultimate shortening corresponding to a moderate continent—continent collision (without continental subduction). It still remains to determine the plate movements all around the West African craton and to integrate the different Pan-African collision types in a continent-wide dynamic model.

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5. Geochemistry, isotope geology, petrology and geophysics: Constraints and models for Precambrian crustal evolution

Chapter 18

EARTH TECTONICS AND THERMAL HISTORY: REVIEW AND A HOT-SPOT MODEL FOR THE ARCHAEAN

R. St. J. LAMBERT

ABSTRACT

A variety of geophysical, petrological and geochemical models for the earth are reviewed. Most deal with either the initial or present states of the earth and do not attempt a comprehensive theory. Convection models and thermal history models assume an initial hot, partly molten earth state with the exception of some two-layer convection models for a cool earth, constructed but rejected by McKenzie and Weiss. Petrological and geochemical models almost all emphasize mantle heterogeneity but are divided into varieties advocating either early major continental growth or steady (perhaps episodic) continental growth. Isotopic constraints have been interpreted in widely differing ways, but the preferred view is that Sr, Pb and Nd isotopes now all indicate progressive growth of the continents with a maximum growth rate in the late Archaean. Recycling has been postulated on chemical and isotopic abundance grounds, but does not appear to be important until the Phanerozoic. These ideas, taken together with a philosophy which equates high heat flow with high rates of continental growth, leads to the author's model of a hot-spot non-plate-tectonic model Archaean, in which interaction of hot-spots (plumes) with primitive oceanic or mini-continental crust can lead to either the granite—greenstone or ancient gneiss type of Archaean crustal complex.

INTRODUCTION

Models of the earth and its history generally fall into two classes; essentially geophysical or geochemical-petrological. The former are developed from heat-flow histories or convection models, the latter from either accretion theories, lunar analogy, or data from Pb, Sr and Nd or stable isotopes. In some cases ideas from several different approaches have been mixed. Faced with the problem of the nature of tectonic processes in the Archaean, exemplified by the variety of solutions proposed (e.g. by various authors in the first sections of Windley, 1976, and Windley, 1977, chapter 3) I decided to consider the constraints imposed by the presence of a conductive lid or lithosphere at the earth's surface, which has fairly easily defined properties. In 1976 I argued that Archaean crustal thermal gradients were not extreme, using some indifferent evidence of *P-T* gradients from metamorphic complexes and calculated geotherms for a 30 km-thick basaltic crust (see also Davies, 1979). In 1978 (published 1980) I calculated the maximum possible thickness for an Archaean crust as 45 km, assuming 5 km K-rich granite/sediments at the top, 18 km of mixed schists, amphibolites and gneisses and a 22 km basal layer of pyroxene-granulites. In the same paper it was argued that there is a one-to-one correspondence between high surface heat flow (q) today and the degree of present tectonic activity and that there was also a correlation of crustal fractionation processes leading to the production of sialic rock with such active tectonic zones. Arguing that an Archaean regime of very high q (say 200 mW/m^2) would produce continental crust at a very rapid rate and that major recycling of sial is not likely, the Sr, Pb (and now Nd) isotopic evidence was held to indicate progressive continental growth under comparatively low q ($\leqslant$125 mW/m^2). This process culminated 2600 Ma or so ago during a period of maximum q, since when continental growth has been comparatively minor. To account for these observations, it was argued (Lambert, 1980) that the two-layer mantle-convection system proposed by McKenzie and Weiss (1975) could provide an appropriate heat-flow pattern, if the earth was initially cool (i.e., essentially never partially molten).

Such a model is consistent with a heterogeneous accretion model for the earth and with the Sr and Pb isotopic evidence from modern oceanic floors and islands. A discussion and integration of these lines of evidence was presented to the Geodynamics Commission Working Group 5 of the IUGG in 1979 (Am. Geophys. Union, Spec. Vol., in press).

Whether or not this model is correct as a whole, the implications for the Archaean are clear. In the following I will discuss other earth models, consider the heat-flow problem afresh in the light of recent work (which has suggested raising q (today) from 60 to 80 mW/m^2) and develop a non-plate, hot-spot model for the Archaean.

GEOPHYSICAL AND GEOCHEMICAL MODELS

Prior to the recognition of plate-tectonic theory and its corollary, that the mantle must convect, purely conductive models for the earth's thermal history were prevalent. Hanks and Anderson (1969) ushered in a new series of calculations involving a convecting mantle, showing that core formation must have occurred early if the earth was not to be so molten as to inhibit crustal formation and preservation in the Archaean. As we now know that crustal rocks are preserved which are 300 Ma older than Hanks and Anderson assumed as a limit, their arguments for a very short accretion period and early core formation are reinforced. However, the accretion process is very complicated (Kaula, 1979; Smith, 1979, pp. 15—20) and Kaula's preferred accretion time is longer than Hanks and Anderson's by a factor of no less than 10^3. Vitjazev and Majeva (1977) have also considered the effect of core formation on thermal history: their model shows that even with core formation lasting 2 Ga the calculated surface heat flow reached a maximum of five times that of today and the earth must have a mantle thermal gradient close to the melting point. There are numerous difficulties with all such models, but one firm conclusion emerges: the energies associated with accretion and core formation after accretion are both so high that it is difficult to escape the conclusion that the earth must have been extensively melted at some stage. This conclusion stands for the combination of a homogeneous accretion process and core formation after accretion.

The rather different approach of McKenzie and Weiss (1975) was based on two-layer mantle convection (Richter, 1973; McKenzie et al., 1974). In this model, involving internally heated convection cells, the 700 km discontinuity in the earth is taken as a major boundary and various initial temperature states were assumed without discussing their cause at length. Computations involving instantaneous thermal equilibrium were made for initial states of 0°C, 1000°C, 2000°C and a completely convective earth. McKenzie and Weiss preferred the latter model, which has a declining heat-flow pattern, whereas the other models had heat-flow maxima. Values of these maxima depend on assumed K, U and Th abundances and the time of the maxima on the assumed initial temperature. On the simple assumptions that the core generates negligible energy, that accretional and core formation energy were negligible today and that all heat flow today (q = 60 mW/m^2) is due to radioactive heat, I estimated (Lambert, 1980) a heat-flow maximum of 125 mW/m^2 at 2800 Ma ago if the earth was initially at 1500°C throughout. Part of the reason for arriving at this result was, however, a specific geological perspective of the nature of the Archaean crust.

By contrast, a thermal history with whole mantle convection was proposed by Sharpe and Peltier (1978). They showed that a homogeneous accretion process could generate enough energy to provide a surface heat flow of about 27 mW/m^2 today which, together with today's crustal heat production, roughly equalled the Lee (1970) heat flow of 60 mW/m^2. This model requires zero radioactivity in the interior of the earth and produces an internal geotherm which is close to the melting point throughout the mantle for all of geological time since 4 Ga ago. It is therefore a geologically steady-state model, with q declining from 125 mW/m^2 3.8 Ga ago to a near-plateau at 75 mW/m^2 in the Proterozoic, followed by a slow decline to today. This model also predicts a mean mantle viscosity of 10^{22} poise today, the maximum value reached. Higher mean viscosities correspond to higher rates of convection at the surface and higher heat losses. The model depends on homogeneous accretion, periods of time when surface heat loss is by conduction only and negligible internal radioactivity, with none of which I would agree.

Most geophysical models, therefore, tend to treat the earth as being comparatively uniform, predict a unidirectional or even constant surface history and emphasize high initial temperature states. By contrast, geological, petrological or geochemical models are much more variable and emphasize heterogeneity. Included in this general class are expanding earth models, which are generally invoked to remove anomalies in plate-tectonic or sea-floor-spreading hypotheses. For instance, Glikson (1977) has proposed that the earth was small until 1 Ga ago, when expansion led to the present plate-tectonic regime. This model would explain the extraordinary contrast between Proterozoic and Phanerozoic tectonics which is widely recognized (Glikson, 1977; *see also Glikson, this volume, Chapter 4, ed.*). However, application of uniformitarianism in heat production and heat flow requires that a small globe must have higher surface heat flow. Glikson's Proterozoic earth with half its present radius would have had one-quarter of its present surface area and four times its present heat flow. No sialic crust could stabilize under such conditions. The early Proterozoic crust has all the characteristics of being exceptionally stable and this extreme model must be rejected.

Petrological or geochemical models, like the geophysical models, usually address themselves to one aspect of the earth to place constraints on evolutionary patterns. Sr, Pb and Nd isotopes and lunar analogies play a large role in these models. An early anorthositic + granitic crust has been advocated by Shaw (1976) assuming homogeneous accretion, early high heat flow and invoking lunar analogy. W. S. Fyfe (pers. commun., 1980) also thinks some early anorthositic crust likely. Early granitic crusts have been advocated by Hargraves (1976) and Ringwood (1979) although their models differ considerably in detail. These models are unsatisfactory in that no known method of recycling granite or anorthosite exists, yet neither type of crust has left any isotopic or geochemical mark on the present surface. Further, Moorbath (1975, 1976, 1977) and O'Nions and Pankhurst (1978) have argued, chiefly from Sr isotope evidence, that continents have grown steadily (or episodically) from small beginnings c. 3.8 Ga ago and that there cannot have been a large-scale primordial crust unless it has been completely recycled in some singular event. S. Moorbath (pers. commun., 1979) has extended this argument by including Pb isotopes, using a large number of whole-rock analyses from the Amitsoq and Nuk gneisses of western Greenland which developed sequentially as primary mantle-derived units (*see also Moorbath and Taylor, this volume, Chapter 20, ed.*). Nd isotopes also indicate negligible reworking of sialic material, insofar as data are available, excepting the case of the Churchill Province of the Canadian Shield which has a primary age in the latest Archaean (De Paolo and Wasserburg, 1976; McCulloch and Wasserburg, 1978; Hart, 1979). The McCulloch and Wasserburg data led these authors to conclude that "the period 2.5 to 2.7 AE was a major epoch of formation of new continental crust — (also) manifest from the studies of previous workers" and that "periods of continental growth are sharply episodic".

Episodic continental growth requires episodic depletion of the mantle, so it is not surprising that Sr and Pb isotope studies clearly indicate mantle heterogeneity (Carlson et al., 1978; Hofmann and Hart, 1978). There is a division of opinion as to whether the Rb-Sr and Pb data on modern basalts and related rocks from oceanic areas indicate genuine Proterozoic ages for mantle fractionation episodes or whether the data reflect mixing of two or more discrete source regions in the mantle. The problem has been discussed extensively by Tatsumoto (1978) whose conclusion in favour of mixing I find acceptable (cf. Moorbath, 1977, p. 174). Whichever conclusion is correct, the broader conclusion from these studies, that there are regions of the mantle with different histories, isolated from one another until partial melting and crust formation occurs, is highly significant. No geophysical model for the mantle can ignore this conclusion. Further, the isotopic anomalies are thought to be related in some way to the hot spot or plume hypothesis of mantle behaviour (Schilling, 1973; Tatsumoto, 1978) although the models proposed by these two authors are essentially the converse of one another (Tatsumoto, 1978, pp. 80—82). Another general conclusion from these studies is that, even if the direct age interpretation of mantle isochrons is invalid, the data *require* a differentiation of the mantle at or prior to the apparent age. This requirement can be derived from Tatsumoto's analysis of the Pb isotope data (Tatsumoto, 1978, fig. 10). The problem has been analyzed physically by Richter and Ribe (1979) who show that isotopic heterogeneity followed by mixing of isotopes from more than one source is consistent with, and permitted by, current convection theory. Their result is consistent with either a plume model or a two-scale flow model for the upper mantle, but is developed in terms of the latter. Another geophysical model in which advection in the upper mantle is held to be primarily responsible for the genesis and mixing of LIL-element depleted and enriched regions of the mantle is that of Anderson (1979). This model involves depleted eclogite and/or garnetite cycling through the upper mantle, enveloping and occasionally mixing with enriched, relatively primitive mantle. Neither Richter and Ribe's or Anderson's models have been tested against earlier conditions in the earth: in particular, it seems difficult to apply Anderson's model to the Archaean (or perhaps even the Proterozoic) as eclogite is not a likely feature of the earth at that time (Green, 1975; Lambert, 1976).

All petrological and geochemical models and any geophysical model which attempts to explain constraints from such models run into the vexed question of the degree of recycling. Analysis of this question has been attempted by O'Nions, Evensen and Hamilton (O'Nions, 1979) and Veizer and Jansen (1979). The O'Nions model accounts for major continental growth in the late Archaean, using exchange between a continental crust plus atmosphere reservoir and the mantle controlled by different upward and downward time constants. This model is a satisfactory mathematical expression for the gross characteristics of Sr and Nd distribution as at present known, but is difficult to relate to an earth with an episodic unidirectional (non-cyclic) history without modification, at the very least by making the time constants themselves time-dependent. However, the model does require that there be a considerable degree of recycling from earliest recorded geological time. The model of Veizer and Jansen starts from a constant or slightly increasing continental growth rate with time and develops a high-level (non-mantle) recycling model which accounts for exponential growth of area of stable continents, of thicknesses of sedimentary sequences and volumes of mineral deposits with time. The first of these data bases (area of continents stabilized per unit time) seems particularly poorly defined and the curve shown (O'Nions, 1979, fig. 1) is impossible to reconcile with either O'Nions' or McCulloch and Wasserburg's statements concerning episodic continental growth concentrated into the Archaean (see also Lambert, 1980).

The preceding discussion shows that earth evolutionary models are very much at the hypothesis stage of scientific development. Conclusions likely to be of lasting consequence are few; questions and challenges are dominant.

THE HEAT FLOW AND HEAT GENERATION PROBLEM

The revision of the mean surface heat flow of the earth from 60 mW/m^2 to close to 80 mW/m^2 (rounded-off here for discussion purposes) by Williams and von Herzen (1975) and Sclater et al. (1979 — lecture at Working Group 5 meeting of IUGG, London, Ontario) necessitates some re-examination of the arguments which I have used earlier (Lambert, 1980). If we assume heat production from the core of 2×10^{12} W (modified from Loper, 1978) then I estimate the following heat productions today (cf. 1979 calculation): continental crust 6.4×10^{12} W, oceanic crust 0.4×10^{12} W and mantle 32×10^{12} W; total 40.8×10^{12} W. 2600 Ma ago the corresponding figures were 12.6×10^{12} W, 0.8×10^{12} W and 64.6×10^{12} W for a total of 80×10^{12} W given constant core heat production. Such a mantle heat production requires 305 ppm K, 0.035 ppm U and 0.100 ppm Th. However, if there is a time-constant of (say) 0.5 Ga involved in the transfer of thermal energy to the surface, these abundances (which are maxima) would be reduced to 240 ppm K, 0.028 ppm U and 0.080 ppm Th. I have argued (Lambert, 1980) that such figures are eminently reasonable from a petrological point of view. Smith (1979) gave 20 ppb U for the whole earth from cosmochemical arguments which corresponds to 23 ppb in the mantle if there is zero in the core and 1.91 ppm (Lambert, 1980) in the continents (of mass 1.49×10^{25} g). This figure of 0.023 ppm is respectably close to the 0.028 ppm derived above: my estimate is partly controlled by choice of a mantle K/U of 8880.

Given equilibrium and a zero time constant in the heat production/heat flow pattern we end up with a heat flow of 157 mW/m^2 at 2600 Ma and 236 mW/m^2 at 3600 Ma. If there is a significant time constant involved (e.g. 0.5 Ga) then the "equilibrium" heat flow would be greater at corresponding past times: about 185 mW/m^2 at 2600 Ma and 290 mW/m^2 at 3600 Ma. Consider, however, the lower figures: the 2600 Ma value of 157 mW/m^2 is only reached today in contemporary volcanic areas (mid-ocean ridges and oceanic crust < 10 Ma old; parts of Japan; the Vesuvian province, etc.). I argue still, as in 1979, that the converse rule holds good; such high heat flows can only be produced where magmas are close to the surface. Such areas are places of crustal production, including continental crust: it is not surprising that the Archaean continents grew quickly. But why did they not grow before 3000 Ma or so? In my opinion the answer remains that the earth was cool before that time and that it reached a thermal maximum in the late Archaean. The high average heat flow today of 80 mW/m^2 reinforces the argument put forward on the basis of only 60 mW/m^2.

There may well be a case for revising the details of the thermal history I have suggested (Lambert, 1980, fig. 1) which was based on 60 mW/m^2 today. Suggested thermal maxima, following McKenzie and Weiss (1975) were 120 mW/m^2 at 3500 Ma and 130 mW/m^2 2800 Ma. Interpolation of McKenzie and Weiss' model for an initial temperature of 1500°C and my estimates for K, U, Th gave less preferred maxima of 130 mW/m^2 at 3600 Ma (too early in my opinion) and 100 mW/m^2 at 2400 Ma. The two maxima correspond to the onset of upper and lower mantle convection, respectively. Modifications to these estimates are of degree only and not of kind and will be discussed elsewhere. The hot-spot model does not depend critically on the exact heat flow, at least not in the present tentative form of the model.

A HOT-SPOT MODEL

Surface heat flows of the order of 120 mW/m^2 to 180 mW/m^2 appear to require magma at shallow depth wherever they occur. Supposing that the range of values, from hottest to coolest regions at any one time, is of the order of a factor of two (roughly today's situation, away from actual volcanic areas), a suitable range for the late Archaean would be 100 to 200 mW/m^2. In today's terms this corresponds to 20 Ma old oceanic crust (Davis and Lister, 1977, fig. 11) and Iceland, respectively. Comparing these limits with Oxburgh and Parmentier's (1977) conclusion "Oceanic lithosphere is gravitationally stable on the asthenosphere until it is 40—50 Ma old" we see that the return flow from Archaean upwelling centres must have been in different form from today's. The laws of conduction do not change; nor do mantle liquidus or solidus temperatures change with time (although secular variation of mantle composition would slowly cause such a change). The oceans were almost certainly present in the late Archaean, averaging 2—8 km deep (Hargraves, 1976): heat exchange and hydrothermal processes were presumably similar. Present-day 40 Ma old crust has a mean heat flow of $70\,mW/m^2$, well below the suggested late Archaean range.

If all the oceanic crust and lithosphere was buoyant, there could be no subduction of slab-like portions of the lithosphere, a feature *essential* to what we call the plate-tectonic regime. Roughly speaking, plate tectonics cannot begin until some oceanic lithosphere becomes gravitationally unstable: there would have been a lithosphere prior to this time, but the geodynamic process called plate tectonics requires subduction of slabs in order for all its manifestations to be present.

The only alternative is a *hot-spot model* with the return flow being a body flow, the simple depression of crust back into the mantle. The *P-T* curves of Green (1975) permit an estimate of the nature of this process. All geotherms were at or above today's average: basalt would descend along a geotherm (if in thermal equilibrium) through the garnet-granulite or amphibolite fields until melting intervened at 40—70 km, depending on H_2O content. The partial melting of garnet-granulite has not (to my knowledge) been studied in detail, but basalt in such a facies will probably melt at or close to an invariant point or cotectic to a liquid close to its own composition. Amphibolite will yield small volumes of tonalite plus granodiorite, leaving a denser residue to return to the mantle. Simple isostatic calculations show that some 20 km of tonalite must be added laterally to a 50 km basalt pile before a continent will emerge from a 2.8 km deep ocean; simply melting the basalt/amphibolite in situ without removing the products laterally will of course not change the depth of the ocean floor appreciably other than by heat transfer and expansion changes. Continent production (in the sense of continent with freeboard) will therefore be a slow process if it is preceded by ocean formation and will (if involving tonalites) take place in the older parts of the oceanic crust — the cooler parts. However, by Iceland analogy, a thick basalt pile at a hot-spot should also produce primary sialic differentiate. If the pile was submarine and contained appreciable H_2O, the differentiation process should tend towards the calc-alkaline rather than tholeiitic variety, producing perhaps the partly subaerial central volcanoes which have been identified in the Canadian greenstone belts (see papers by Goodwin, Ayres, M. B. Lambert and by Wilson and Morrice in Baragar et al., 1977).

My hot-spot model therefore contains two active components, namely the upwelling and downwelling regions. These could at first sight be arranged as central sinks and linear upwellings in a honeycomb pattern, or vice-versa. The latter pattern better solves the space problem. An Archaean upper mantle and crust of upwelling at hot-spots and honeycomb patterned downwelling areas, both probably spatially transitory, is therefore proposed.

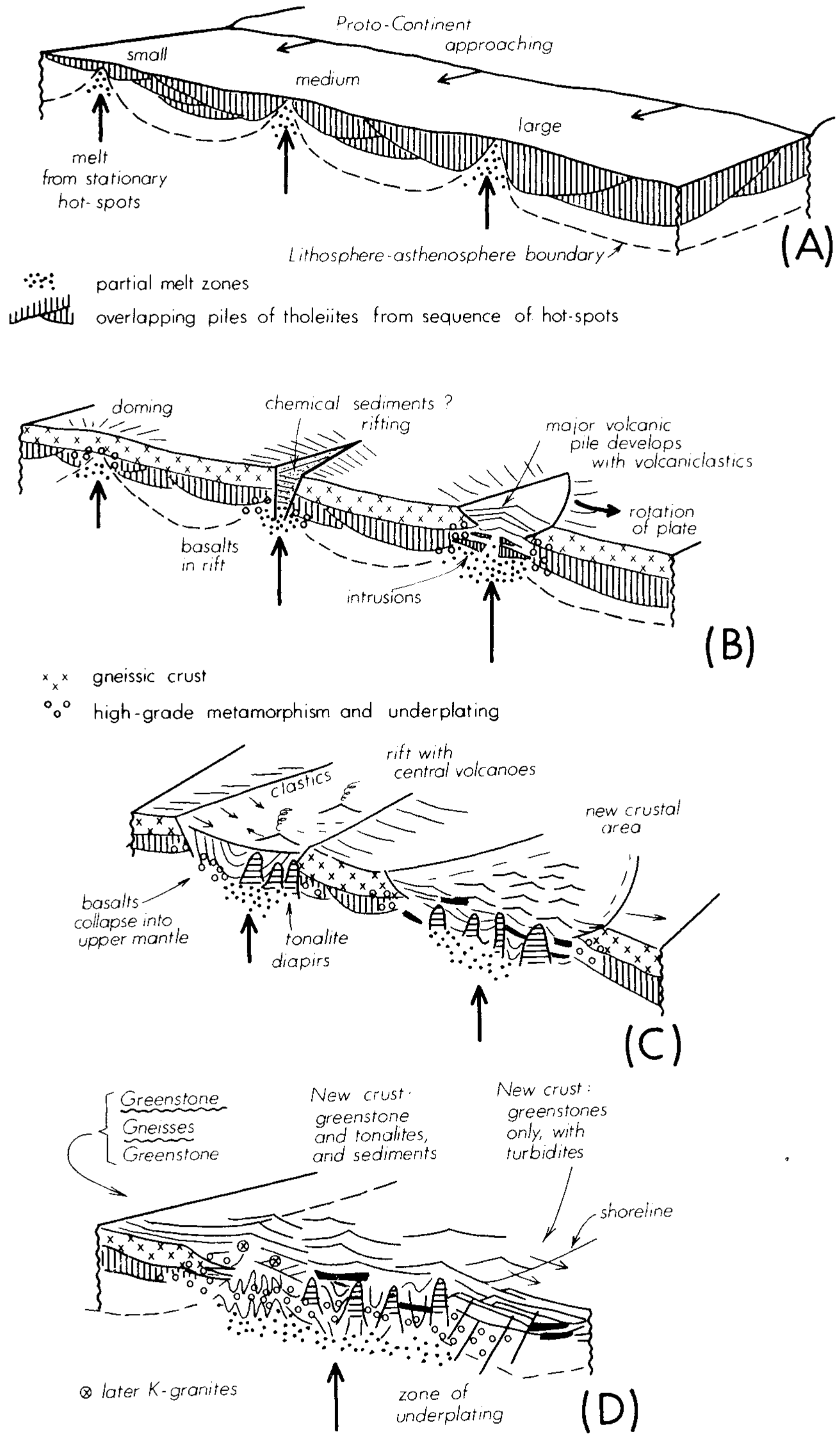
In this model greenstone belts are remnants of oceanic crust. Their complex tectonics and relationships to adjacent sialic units as summarized by Windley (1977, chapter 3) can be explained simply by this model. Consider a series of hot spots which are generating large, overlapping submarine volcanic piles (Fig. 18-1). Within this oceanic crust differentiation will sooner or later produce sialic intrusives, which will form continental nuclei. As centres of hot-spot activity migrate with respect to the more consolidated parts of the crust and its embryonic subjacent lithosphere (as they do at present), a variety of interactions may ensue.

Before discussing these further I would note that continental crust will only be created and preserved where an appropriate mix of felsic and mafic crustal components is developed. No felsic component means no continent: too much will raise the continental crust above sea level and trigger erosion until equilibrium is restored. Continental crust formation up to some 40 km is therefore more or less automatic, given the relative densities of the four main units: ocean, sialic and mafic differentiates and barren upper mantle. The omnipresent asthenosphere is the zone of fundamental isostatic compensation.

The features of greenstone belts which must be explained by any model are the almost invariably synformal and metamorphic nature of the greenstones; their thickness, much greater than modern Layer 2 of the oceanic crust, or even, in some cases Layers 2 and 3; that no base is ever seen except in rare cases where greenstones overlap onto pre-existing sialic complexes; their invariable association with volcanoclastic turbidites and chemical sediments but no other kind in significant abundance. They are usually cut by tonalite-granodiorite complexes which are probably mantle-derived and by late potassic granites which appear to have a metasedimentary component. The minicontinents of ancient sialic crust are usually gneissic, dominated by tonalites and amphibolites in the amphibolite or (rarely) the granulite facies. The relationship between these two main components of the Archaean continents cannot be simple: Windley (1977, pp. 58—61) has shown that there is no concensus of opinion.

To examine the problem, consider a model in which enough protocontinent has aggregated to form a discrete unit, cohesive but not yet thick enough to break the oceanic surface. It is a mini-plate: relative to the mantle and its plethora of hot-spots it can migrate. Suppose there are three varieties of hot-spot (plume) over which it can pass — small, medium and large. Fig. 18-1A shows the scene before there is any significant interaction. Figs. 18-1B and 1C show interaction and Fig. 18-1D shows a possible final result for one sector of the system.

The oceanic crust is shown as consisting of overlapping structureless piles of (pillow) lavas, domed over the plumes. The latter are partially melted near the surface and are contributing to the oceanic crust. As the oceanic crust with its thin continental slab drifts over the plumes (Fig. 18-1B) the small plume warms up the crust; it may add melt as dykes and sills, perhaps causing underplating, but no other effect is envisaged. The metamorphism in the oceanic crust could easily reach granulite facies levels (800°C, 8 kb).


Proto-Continent
approaching
small
medium
large
melt
from stationary
hot-spots
Lithosphere-asthenosphere boundary
(A)
partial melt zones
overlapping piles of tholeiites from sequence of hot-spots
doming
chemical sediments ?
rifting
major volcanic
pile develops
with volcaniclastics
rotation
of plate
basalts
in rift
intrusions
(B)
gneissic crust
high-grade metamorphism and underplating
rift with
central volcanoes
clastics
new crustal
area
basalts
collapse into
upper mantle
tonalite
diapirs
(C)
Greenstone
Gneisses
Greenstone
New crust·
greenstone
and tonalites,
and sediments
New crust:
greenstones
only, with
turbidites
shoreline
later K-granites
zone of
underplating
(D)

The medium-scale plume causes rifting of the crust, but does not provide enough energy to cut it apart completely. The rift contains basaltic volcanism and there will be extensive lower-crust metamorphism as before. The large plume breaks the crust and is shown as rotating the end of the continental fragment away from the main body. The large body of low-density partial melt and the accompanying thermal expansion raise the crust, old and new, to relatively high elevations, but in the absence of partial melting and lateral transport the crust all remains submarine.

As the continental fragment crosses the main part of the plumes further rifting will occur (Fig. 18-1C). The small plume is ineffective beyond causing results already mentioned, but the other plumes are shown as developing new environments. The rift enlarges but remains an entity. The basalt pile is collapsing into the partial-melt region as dehydration reactions weaken it and its mean density increases. The plume is shown as shifting slightly off-axis of the rift, concentrating partial melting of the amphibolitized base of the new basalt pile, producing tonalite diapirs. These are shown as cutting the edge of the rift, confusing the local geology (particularly as the ancient gneissic crust shown is probably tonalitic). The tonalites may be subjacent to calc-alkali volcanoes (dacite being the commonest silicic lava in the Archaean), which provide abundant submarine pyroclastics and, occasionally, ore-bodies. Submarine erosion from cliffs along the edge of the rift may provide local breccias or fanglomerates to the trough in which chemical sediments will also accumulate. If erosion of this complex ever reveals the contact of the greenstone belt and the ancient gneisses, it will either be faulted or masked by intrusion. The large plume continues to rotate the edge of the ancient continent away and fills the space with thick basalts with abundant intrusions. The pile thickens until tonalite diapirs form as in the rift valley. In Fig. 18-1D the process continues (whereas the medium-size plume does not possess enough energy to develop the system any further in any one locality). As the large plume continues to provide partial melt and heat to the new crust, regional metamorphism will become pervasive to the point that partial melting of the ancient gneisses and the new sediments may occur, producing K-granites; the greenstones collapse into steep structural patterns as tonalite ($\sim$ granodiorite) diapirs rise; the crust thickens to the point where the surface becomes subaerial and granulites can develop extensively at the base of the pile. If the process continues long enough, the "new" tonalites may themselves become gneissic and very difficult to distinguish from parts of the pre-existing crust. The edge of the system will not be as complex, consisting solely of greenstones, basic intrusives and

Fig. 18-1. A. A proto-continent and associated thick basalt oceanic crust approaches three stationary hot-spots.
B. Beginning of hot-spot/continent interaction.
C. Main stage of development.
D. Final stage of new crust above the large hot-spot.

sediments above. This part of the complex will be typical oceanic crust and will eventually be destroyed. Note particularly that the left hand edge of the section in Fig. 18-1D will show the sequence ancient oceanic crust—ancient sialic layer—younger greenstone belt. The models shown provide us with the standard combinations of greenstones older than, younger than, faulted against and intruded by sialic crust (note that the ancient gneissic continent will itself be composite, having been formed by an earlier version of the same process). The model shows that high-grade metamorphism can be superimposed on older crust or on the new crust; that either old or new crust can be partially melted, and that almost every facet of the whole sequence will be submarine.

The process continues until the progressive addition of new crust of the appropriate density brings continents into existence which resemble their modern counterparts in size. The cessation of this process somewhere around 2500 Ma ago must also be explained by a successful model. Heat flow in the thermal model advocated here declines at this stage, but such a change must be gradual, whereas the cessation of Archaean continental growth appears relatively rapid. Plumes must have continued to exist; the only difference seems to be that continent/plume interaction ceased, at least in its surface manifestation. Perhaps the strength of the continents rose to levels which the plumes were unable to break, or as lithosphere thickness began to grow under the continents (by cooling) the partial-melt zone (asthenosphere?) was unable to develop and no melts were available in the continental areas. Consideration of a *P-T* diagram showing a peridotite solidus, an ascending diapir, and barriers at (say) 10 km and 50 km depth will show why partial melting in the mantle will eventually be suppressed under a thickened continent with subjacent lithosphere. This, rather than a mechanical explanation, or a coincidence such as the "drying-up" of the source region, is my preferred solution.

CONCLUSION

Models of earth history are still subject to an extraordinary degree of uncertainty, due to our lack of understanding of the present-day interior of the earth and the fragmentary crustal record for the ancient crust. Constraints on models are not many: the chief one chosen in this account is that isotopic data require a heterogeneous, fractionated mantle and a continuously growing continental crust, though episodic in that most of the growth occurred during the late Archaean. The chief assumption made is that high surface heat flow and a high rate of tectonic activity and crustal production are all directly related. Further, once sialic crust has been produced in tectonically active, hot areas, that crust does not return to the mantle at a significant rate. Differentiation of the mantle is therefore considered to be irreversible. It is therefore argued that a very high average heat flow is not possible in the earlier Archaean: an exact limit cannot be established, but values over 150 mW/m^2 are unlikely by modern analogy. At such levels of heat flux thick lithospheric plates cannot exist and all lithosphere is buoyant. Any return flow to the mantle is by general "body flow" of the lithosphere, which is probably quickly recycled above hot-spots. Fractionation through partial melting of the lower parts of amphibolite piles under high P_{H_2O} and some crystal fractionation in tholeiitic magma chambers, produces sialic crust. This crust slowly aggregates. When passing by or over stationary plumes a variety of tectonic events can be envisaged: doming, metamorphism and perhaps underplating of the primitive crust; rifting with formation of thick tholeiitic lava sequences (greenstone belts), later folded and intruded by locally generated tonalite diapirs; and in the event of massive extrusion of tholeiite from a major plume, the creation of new continental crust on a large scale due to the pervasive mixing of old and new sialic crust and old and new greenstones. No subduction need be invoked to explain all the major features of the later Archaean as preserved, given a hot-spot model for a comparatively cool earth with a thermal maximum some 2800 Ma ago.

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Chapter 19

PETROGENESIS OF ARCHAEAN ULTRAMAFIC MAGMAS AND IMPLICATIONS FOR ARCHAEAN TECTONICS

D. H. GREEN

ABSTRACT

The distinctive peridotitic lavas and intrusives of Archaean greenstone belts imply magma temperatures of $> 1600^{\circ}C$ at near-surface conditions. These magmas are probably produced by two or more stages of melting of rapidly rising peridotite diapirs originating at depths in excess of 200 km in the Archaean mantle. Immediate source compositions for individual lavas imply geochemical heterogeneity in the Archaean upper mantle, similar to that in the modern mantle. The high magmatic temperatures of Archaean magmatism imply differences in the Archaean geotherm. While it is possible that a lithosphere similar to that of the modern earth and a similar pattern of plate tectonics existed in the Archaean, an alternative, preferred model predicts a thinner lithosphere, more active plate movements, instability of eclogite at deep crustal depths and an absence of subduction of oceanic crust.

INTRODUCTION

This chapter focusses attention on the evidence for remarkably high temperatures of extrusion of the Archaean magmas commonly referred to as peridotitic komatiites or "spinifex-textured peridotitic komatiites" and discusses models for petrogenesis of such magmas and the implications of these models for Archaean tectonics. These Archaean magmas are compared with the most magnesian and highest temperature liquids among basalts erupted in the last 500 million years of earth history in an attempt to evaluate possible changes in mantle composition or mantle geothermal gradients with time.

The field occurrences of peridotitic komatiites, particularly the presence of pillow structures, brecciated flow tops and identifiable cooling units, demonstrate that these ultramafic compositions were emplaced as lava flows and shallow intrusives. The variety of quench textures, particularly the acicular, interlocking bladed olivine crystals forming the distinctive "spinifex" texture, and the evidence for crystallization, settling and accumulation of olivine phenocrysts, demonstrates that the original magmas have chilled rapidly but also that there has been, in some cases, opportunity for magma differentiation. The task of identifying liquid compositions, free of accumulate olivine yet sufficiently magnesian to precipitate the most magnesian olivine observed in the natural rocks, is an important first step. Once the compositions of Archaean liquids are established, the temperature of extrusion of the magmas can be determined. Previously, a peridotitic komatiite from Barberton, South Africa (Viljoen and Viljoen, 1969a, b) was investigated and it was shown that the magma was essentially anhydrous ($< 0.2\%$ H_2O) on eruption and erupted as a completely liquid peridotite magma at a temperature of $1650 \pm 20°C$ (Green et al. 1975). In this paper, quench-textured peridotites and a quench-textured magnesian basalt from Australia and Canada are compared with the Barberton example (49J).

QUENCH-TEXTURED PERIDOTITES

The samples studied experimentally were selected from among those for which major element and trace element data have been obtained (Sun and Nesbitt, 1978; Nesbitt et al., 1979). Samples include typical bladed or acicular olivine quench (49J, 422/95, 331/144) but also include samples containing remarkable dendritic or feathery quench olivine from Yackabindie (331/277, 331/346, 331/347)(Nesbitt, 1971). Rock compositions (calculated anhydrous) and compositional ranges of relict olivines are given in Table 19-I. The dendritic olivine crystals are surprisingly iron-rich, coexist with very magnesian tremolite and do not show compositional zoning from magnesian cores to more iron-rich rims as is observed in Barberton and Mt. Burges bladed olivine quench crystals. Chromium and calcium contents of these olivines are low and variable. In contrast, the bladed olivine crystals of 49J, 422/95 and 331/144 have magnesian core compositions, show slight normal zoning to more Fe-rich rims and consistently contain $> 0.1\%$ Cr_2O_3 and $\geqslant 0.15\%$ CaO.

Table 19-I illustrates the ultramafic compositions of the Barberton and W. Australian rocks, the approach of the Munro Township composition to a picritic composition and the much more siliceous composition of a spinifex-textured basalt (331/338) from the Negri volcanics in the Pilbara area of northwestern Australia.

EXPERIMENTAL PETROLOGY AND PETROGENESIS OF ARCHAEAN QUENCH-TEXTURED PERIDOTITES

The results of the experimental study of the Barberton sample (49J) (Green et al., 1975) are summarized in Fig. 19-1 and Table 19-I. Olivine is the liquidus phase and at 10 kb crystallizes alone (or with accessory chrome spinel) over a temperature interval of $\sim 220°C$ and a compositional range from $Mg_{93.5}$ at the liquidus to $Mg_{89.7}$ at orthopyroxene appearance (1450°C). At lower pressures this temperature interval and compositional range would be greater. Experimental studies on pyrolite and Tinaquillo

TABLE 19-I

Composition of Archaean quench-textured peridotitic komatiites (cols. 1—4), and quench-textured basalt (col. 5)

	Barberton 49J	Mt. Burges 331/144/5	Yackabindi 331/346	Munro 422/95	Pilbara 331/338
SiO_2	46.3	46.1	44.9	46.1	55.6
TiO_2	0.19	0.21	0.27	0.4	0.43
Al_2O_3	3.6	5.3	5.4	8.4	11.8
Cr_2O_3	0.43	0.41	0.43	0.46	0.27
NiO	0.27	0.21	0.21	0.13	—
FeO	10.3	10.4	10.0	10.9	9.3
MnO	0.21	0.21	0.20	0.22	0.22
MgO	32.9	30.9	33.1	23.6	12.2
CaO	5.1	5.9	5.2	8.8	7.8
Na_2O	0.49	0.40	0.28	0.84	1.92
K_2O	0.18	0.01	0.00	0.15	0.48
P_2O_5	0.01	0.00	0.00	0.01	0.04
100 Mg/(Mg + Fe)	85	83	85.5	79.4	70.3
Olivine in rock (Mg value)	93.6 → 89.0	92.3 → 74	87.4 → 71	92.7 → 92.3	nil
Liquidus olivine	⩾93.5	⩾93.6	—	⩾91.3	⩾86.3
Predicted liquidus olivine $K_D = 0.33$	94.4	93.6	94.6	92.0	87.6
CaO content of liquidus olivine	0.14	0.13	—	0.22	0.16
Cr_2O_3 content of liquidus olivine	0.17	0.15	—	0.15	⩽ 0.10
CaO content of most magnesian olivine in rock	0.19	0.15	n.d.	0.22	—
Cr_2O_3 content of most magnesian olivine in rock	0.18	0.15	n.d.	0.12	—.

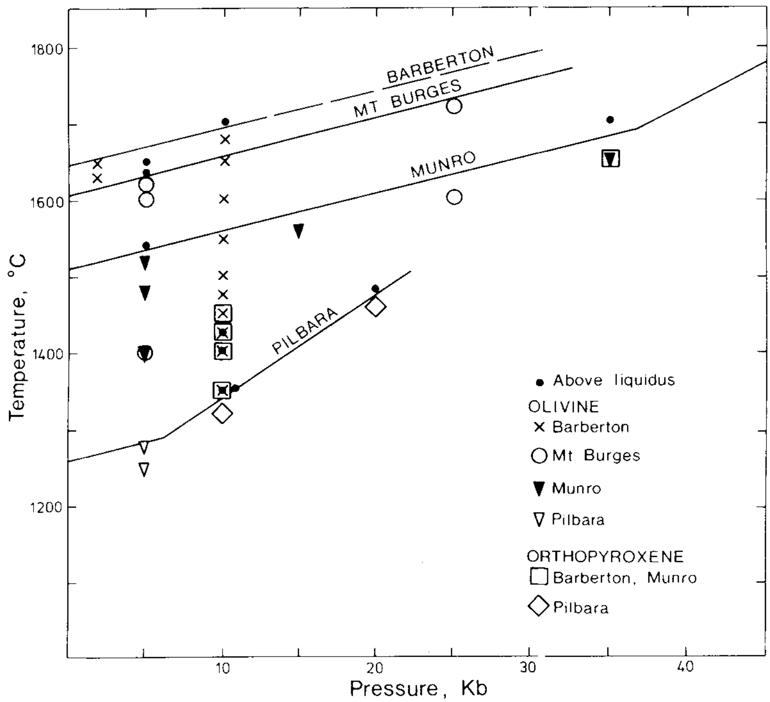


Fig. 19-1. Liquidi and liquidus phases for Archaean quench textured peridotites and basalts (Table 19-I). Symbols indicate the presence of olivine and orthopyroxene in near-liquidus runs.

peridotite composition (Jaques and Green, 1980) determine that orthopyroxene appears at 1270—1320°C at 2 kb.

The experimental studies of Fig. 19-1 and Jaques and Green (1980) are equilibrium partial-melting studies and it should be noted that crystal fractionation of early-formed olivine from peridotitic liquids will produce iron-enrichment and more rapid change in Mg-value of olivine with decreasing temperature.

The crystallization behaviour of Mt. Burges, Munro and Pilbara samples of Table 19-I has also been studied experimentally, using the methods previously described. The use of graphite capsules for sample containers results in lower oxygen fugacity in the charges than is apparently appropriate for the natural rocks. Thus the experimental conditions do not permit assessment of Ni-partitioning between olivine and liquid nor of the precipitation of chrome-spinel since both these aspects of crystallization are sensitive to f_{O_2}. The results of the experimental studies are summarized in Fig. 19-1 and Table 19-I. Olivine is the low-pressure liquidus phase of all samples, and the composition of the olivine obeys, within the experimental uncertainties, the partition relationship $(K_D^{Ol/liq})_{Fe/Mg} = 0.33$ (Table 19-I). Three samples, Barberton, Munro and Pilbara, give K_D values of 0.39, 0.37 and 0.38 using the compositions of the highest temperature olivine observed and the bulk composition of the charge, assuming all Fe as Fe^{2+}. The inability to quench liquids to glass in Barberton and Mt. Burges compositions prevents direct analysis of olivine-liquid pairs in these peridotitic liquids. For the Pilbara sample, analyses of olivine/glass pairs at 5 kb, 1280°C and 1250°C give $K_D = 0.33$ and 0.32, respectively, and for the Munro sample at 5 kb, 1520°C, an olivine/glass pair gives $K_D = 0.33$. Lower temperature experiments contain much quench olivine and glass compositions are iron-enriched during quenching (cf. Jaques and Green, 1979). The data obtained do not provide unequivocal evidence for variation of $(K_D^{\mathrm{ol/liq}})_{\mathrm{Fe/Mg}}$ with temperature, or liquid composition, over a temperature range from 1250°C to 1700°C and for liquids with MgO contents ranging from 12% to 33%. Thus it is appropriate to use $(K_D^{\mathrm{ol/liq}})_{\mathrm{Fe/Mg}} = 0.33$ in testing for equilibrium relationships between observed olivine and postulated liquid compositions or for evaluation of the composition of residual olivine in mantle source regions which have yielded particular magma compositions as partial melts.

The experimental study, taken in conjunction with the field evidence for absence of phenocrysts and for extrusion and rapid chilling of the magmas, defines the extrusion temperatures of the magmas. All magmas were essentially anhydrous. Water dissolves in these magmas under high pressure (Green et al., 1975) and saturation with water depresses the liquidus to lower temperature (e.g. to 1510°C for Barberton 49J at 5 kb or to 1300°C at 25 kb). However, such water-bearing magmas cannot move to the surface without rapidly crystallizing olivine as they pass through the olivine + liquid + vapour field. Hydrous peridotitic komatiite magmas must arrive at the surface in partly crystallized state with porphyritic texture — the samples selected for study do not show such textures and were essentially anhydrous magmas. The extrusion temperatures were:

Barberton	(49J)	— 1650 ± 20°C;
Mt. Burges	(331/144/5)	— 1610 ± 20°C;
Munro	(422/95)	— $>$ 1510°C ($\ll$ 1610°C);
Pilbara	(331/338)	— 1270 ± 20°C.

Note that the liquidus olivine for the Munro composition studied is Mg_{92} [$K_D = 0.33$] whereas slightly more magnesian olivine cores occur in the natural rock ($Mg_{92.5 \pm 0.2}$). Thus the Munro composition is slightly less magnesian and lower in normative olivine than the parental magma which precipitated the observed ($Mg_{92.5}$) olivines. The extrusion temperature of this parental magma would be $>$1510°C but $\ll$1610°C. The spinifex-textured basalt from the Negri volcanics has a liquidus temperature within the range of modern primitive olivine-rich magmas (Green et al., 1979; Duncan and Green, in prep.).

The effect of pressure on all compositions is to increase the liquidus temperature at 4—5°C/kb for the range of pressure over which olivine is the liquidus phase. In the Pilbara sample orthopyroxene is the liquidus phase at 10 kb and 20 kb and the slope of the orthopyroxene-liquidus is ~ 15°C/kb. In the Munro sample, olivine and orthopyroxene occur together at the liquidus (35 kb, 1680°C) and it is expected that orthopyroxene would be the liquidus phase at higher pressure.

Primary magmas of mantle derivation, which have not suffered olivine or other fractionation between magma segregation at source and extrusion at the earth's surface, should be saturated with olivine and orthopyroxene at some pressure. The two samples, Pilbara and Munro, are potential primary magmas, segregating from residual harzburgite at ~ 6 kb, 1280°C and ~ 35 kb, 1680°C, respectively. In the Pilbara example the residual olivine is $Mg_{87.6}$ and since this is necessarily more magnesian than the original pre-melting source olivine composition, the inference that the Pilbara sample is a primary magma has the connotation that the source peridotite was more Fe-rich (~ Mg_{86}) than modern mantle peridotite (Mg_{89-90}). This inference is *not* supported by the very magnesian liquidus olivine and orthopyroxene of the other samples and an alternative interpretation of the Pilbara sample is preferred. The Pilbara spinifex-textured basalt is inferred to be a derivative magma, having fractionated olivine (15—20%) from a parental magma in which the Mg-value would be ~ 75. Such a parent magma would have separated from residual harzburgite (Mg_{90-91}) at ~ 15 kb, 1500°C.

The Barberton and Mt. Burges compositions have olivine as the liquidus phase up to very high pressures. The compositions are themselves peridotitic and in considering these liquids as partial melts segregating from source peridotite at low pressure, the only residual silicate phase could be highly magnesian olivine (~ Mg_{94}). These liquids may be either formed by very high degrees of melting (60—70%) of source peridotites containing 3—4% CaO, Al_2O_2 (i.e. model peridotites matching those which suffice to account for the spectrum of Phanerozoic natural mantle-derived basaltic compositions and natural mantle-derived peridotites) or, alternatively, by a "normal" (~ 30%) degree of melting of a refractory source peridotite (1.5% CaO, Al_2O_3).

The melting behaviour of two peridotite compositions, pyrolite (Ringwood, 1966) and Tinaquillo lherzolite (Green, 1963) has been studied experimentally (Jaques and Green, 1980) and the trends of liquid compositions for increasing degree of melting have been defined at 2 kb, 5 kb, 10 kb and 15 kb. Figure 19-2 is the (jadeite + Ca Tschermaks silicate) — olivine-quartz diagram (Green, 1970) which provides a convenient summary of the data and a comparison with the Archaean magma compositions from Table 19-I. Barberton, Yackabindie and Mt. Burges compositions must represent sufficiently high degrees of melting of mantle peridotite such that olivine only is the residual phase. The Pilbara composition has an unusually low CaO/Al_2O_3 ratio and its bulk composition may reflect low-temperature secondary alteration. Assuming no such problem, however, the magma composition plots to the

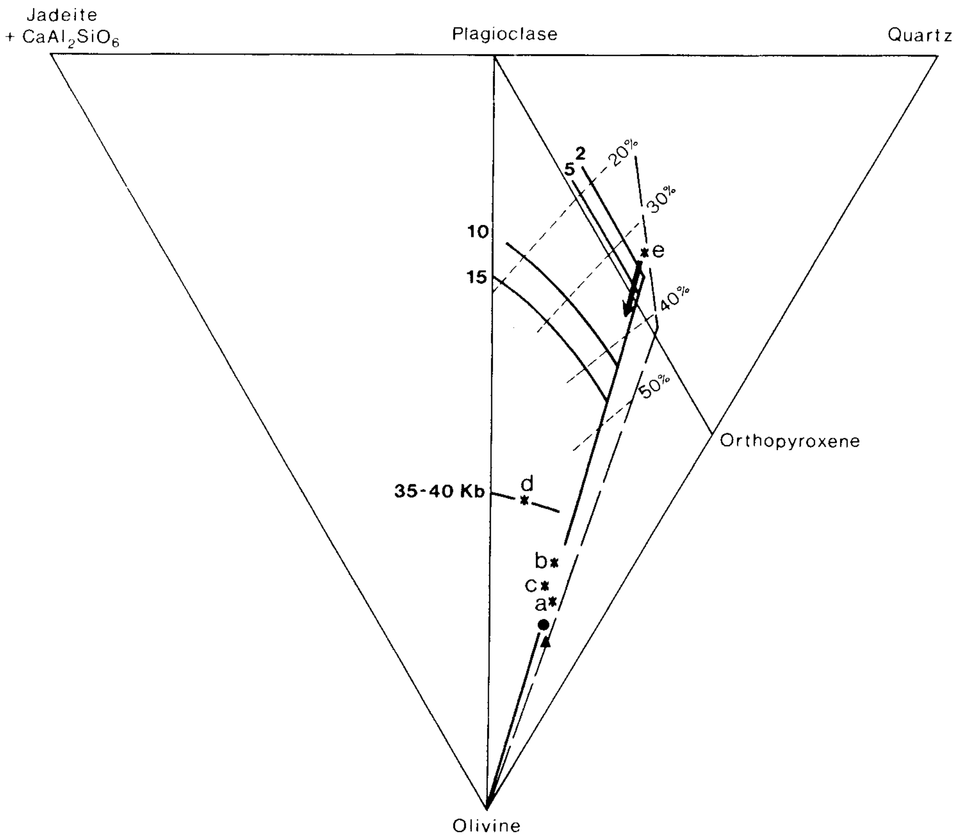


Fig. 19-2. Compositions of liquids derived by equilibrium melting of pyrolite (dark solid lines) at 2, 5, 10 and 15 kb pressure, plotted in the system (jadeite + Ca-Tschermaks silicate) — olivine — quartz (Green, 1970; Jaques and Green, 1980). Short dashed lines indicate percent melting of pyrolite composition, long dashed line shows the compositions of liquids from Tinaquillo lherzolite at 2 kb. Note that the loci of melt compositions follow the ol-opx cotectic until opx is eliminated and then lie on an olivine control line. Co-ordinates calculated from a high-pressure norm in which compositions are cast in terms of the components of the figure together with diopside, ilmenite, magnetite, chromite, orthoclase and apatite. ● = pyrolite; ▲ = Tinaquillo lherzolite; *a* = Barberton; *b* = Mt. Burges; *c* = Yackabindi; *d* = Munro; *e* = Pilbara; (see Table 19-I).

quartz-rich side of the 2 kb and 5 kb olivine + orthopyroxene saturation boundaries but the experimental study indicates olivine and orthopyroxene saturation at ~ 6 kb. This inconsistency, together with the iron-rich nature of the liquidus olivine, suggests that the magma composition results from olivine removal at < 2 kb from a peridotitic komatiite or picritic komatiite melt segregating on the olivine + orthopyroxene saturation line at ~ 15 kb, 45% melting (Fig. 19-2). The Munro composition does not plot on the olivine control line (or on the olivine + orthopyroxene saturation boundary) for the 2, 5, 10 or 15 kb studies but instead is consistent with separation on an olivine + enstatite co-saturation boundary at $\gg$ 15 kb. The experimental study suggests that magma segregation at a pressure of 35—40 kb with ~ 50% melting of a pyrolite of Tinaquillo peridotite-like source would be appropriate.

THE PROBLEM OF SOURCE COMPOSITION AND HIGH DEGREES OF MELTING OF UPPER-MANTLE PERIDOTITE

Although the highly magnesian and most olivine-rich magmas of Table19-I are peridotite in composition and, in three examples, not very different from estimates of the modern upper-mantle composition, it is not easy to suggest processes of complete melting of the Archaean upper mantle. Apart from impact melting there is no known process which can take a body of upper-mantle peridotite from near-solidus (~ 1000°C at 1 atmosphere pressure, 1500°C at 30 kb) to its liquidus (~ 1700°C at 1 atmosphere, 2000°C at 30 kb) so rapidly that separation of crystals and liquid will not occur during either the heating or decompression (diapirism) process. The strong density contrast between olivine and pyroxene crystals ($\rho \approx 3.3$) and olivine tholeiitic or picritic liquid ($\rho \approx 2.7$) ensures that olivine and pyroxene crystals will rapidly settle within a crystal + liquid magma as soon as the percentage of liquid is sufficient to destroy grain to grain contact, i.e. > 25—30% liquid. In a previous paper (Green, 1975) it was suggested that very rapid diapiric ascent could entrain olivine and pyroxene crystals in initially picritic melts, and as the entrained crystals moved to shallower depths they would dissolve, moving the liquid composition towards peridotite. Rapid ascent of magma (sufficient to entrain peridotite xenoliths ($\rho = 3.3$) from depths in excess of 30 km) does occur in modern basaltic volcanism but it is characteristic of small volume, very under-saturated magmas containing high proportions of dissolved (at high pressure) $(OH)^-$ and $(CO_3)^{2-}$ and apparently does *not* occur in olivine tholeiitic or tholeiitic picrite magmas, i.e. it is characteristic of small-volume partial melts of the modern mantle and magmas which characteristically show explosive eruptions.

Arndt (1977) has carefully considered the problem of liquid segregation from an ascending crystal + liquid diapir and has argued that peridotitic liquids represent remelting or continued melting of residual peridotite following extraction of picritic or basaltic melts at deeper levels. There is sufficient flexibility within this general model to account for the variation in both major element and trace element characteristics. Effective separation of a small melt fraction (< 5%) at pressures within the garnet peridotite stability field will produce residual peridotite which may then melt to ~ 30% to yield olivine-rich magmas which are strongly depleted in incompatible elements but not depleted in heavy rare earths. By contrast, a high degree of melting (~ 30%) initially produces liquids with the same relative incompatible elements as the source peridotite and leaves only residual olivine and orthopyroxene. If a small amount of this liquid does not leave the source peridotite but remains as an intergranular film, then continued ascent of the now very refractory peridotite *bulk* composition may yield a second-stage magma of high Mg-value but still with incompatible element contents which are similar to the source. i.e. two or more processes of high-degree melting with incomplete magma extraction do not produce strong fractionation of incompatible element contents, whereas complete extraction of small melt fractions produces maximum fractionation within the incompatible element group. Two or more processes of high-degree (20—30%) melting and incomplete magma extraction produce large differences in the major element compositions of liquids, successive melts being increasingly dominated by olivine and orthopyroxene as the only residual phases contributing to the later stage melts.

Sun and Nesbitt (1978) have considered the implications of incompatible element contents in Archaean magmas, particularly the REE, and have argued that the Archaean mantle was chemically heterogeneous and that the degree of heterogeneity was similar to that deduced for the source regions of modern ocean-basin basalts. The experimental study reinforces this conclusion since only olivine and orthopyroxene are potential residual phases for all magmas and the REE contents of these phases are negligible. The source region for the Pilbara spinifex-textured basalt was enriched in light rare-earth elements (LREE) relative to heavy rare earth (HREE) but the converse is true for Mt. Burges, Yackabindi and Munro — the latter being the most strongly depleted in REE (Fig. 19-3). The Barberton sample is almost chondritic in its REE relative abundances, with only slight LREE/HREE enrichment. There is no simple correlation between REE and major element abundances of the immediate source regions. The implications of hypothetical simple batch melting of source compositions [all of which had $2 \times$ chondritic HREE abundances (Dy $\rightarrow$ Yb)] are illustrated in Fig. 19-3. The La abundance of the source regions would then range from $10 \times$ chondritic to $\sim 0.45 \times$ chondritic, the Ca contents from $\sim 2\%$ to $\sim 4.5\%$ and the % melting of the source regions from 25% to 90%.

To summarize, the compositional variation of the Archaean magmas is such that inhomogeneity of the immediate source regions is required with no apparent simple relation between major element abundances and incompatible element abundances. If we accept the reasonable hypothesis that melting in excess of $\sim 30\%$ of any source peridotite must result in segregation of magma but that the efficiency of separation of liquid from residue is variable and further that magma may segregate from residual crystals at lower degrees of melting, then we have a model for the Archaean mantle which has the flexibility to produce the observed variations in Archaean ultramafic and basic liquids. The major conclusion

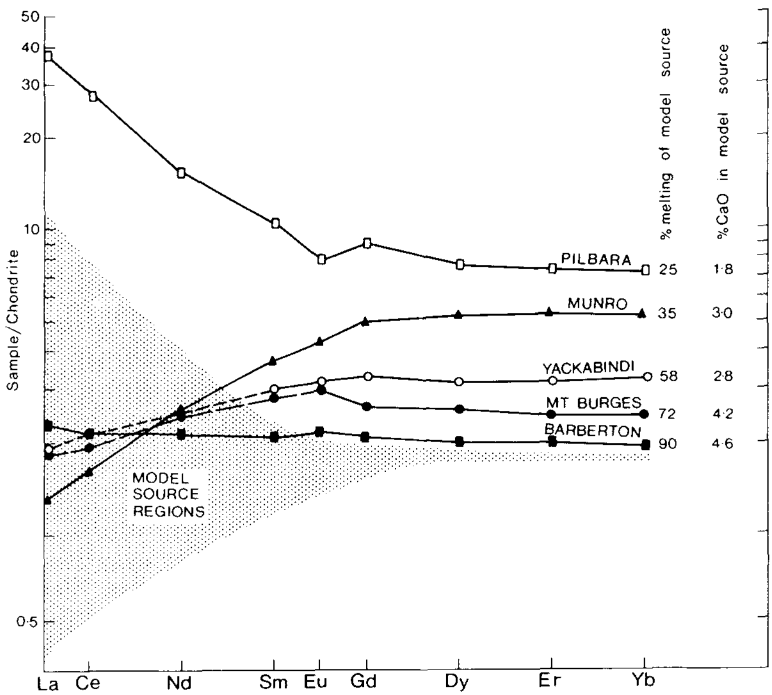


Fig. 19-3. Rare earth abundance data for composition of Table 19-I (from Sun and Nesbitt, 1978). Note that there is little fractionation among the heavy rare earths (Dy → Yb). The shaded area illustrates the range of rare earth abundances in model source regions *assuming* that all source regions have twice chondritic abundances of Dy, Er and Yb. This model would require the source regions to vary in CaO content (see figure) and imply degrees of partial melting from 25% to 90%. The figure illustrates the necessity for the immediate source peridotites of these magmas to be chemically inhomogeneous as the possible residual phases, olivine and orthopyroxene have negligible REE contents.

is that the Archaean upper mantle is not obviously less geochemically complex than the modern earth's mantle in its variability of major element and incompatible element abundances and that processes of incompatible element (particularly LREE) enrichment and depletion by migration of low % melting liquids in the mantle were operative in the Archaean mantle.

COMPARISON OF ARCHAEAN MAGMA GENESIS WITH PHANEROZOIC, PERIDOTITIC AND BASALTIC MAGMA GENESIS

The discussion of the previous section incorporated the concept of diapirism of partially molten mantle peridotite as the immediate cause of magma segregation and ascent in the Archaean upper mantle. In addition the details of magma chemical composition require chemical inhomogeneity of the source peridotite compositions. These are concepts which have been applied to the modern earth and it is important to compare the nature of modern and Phanerozoic magmatic activity with that of the Archaean to understand differences and similarities and their significance.

Diapirism and emplacement of peridotite "magma" at upper mantle and crustal levels has been demonstrated for "high-temperature peridotite" intrusions such as Mt. Albert, Lizard, Tinaquillo, Serrania del Ronda and Beni Bousera (see Wyllie, 1967 for summary). The chemical compositions of these peridotite diapirs are similar to, or slightly more refractory than, the compositions of Barberton, Mt. Burges or Yackabindi of Table 19-I. Typically these intrusions, although at high temperature ($> 1000°C$) causing dynamothermal metamorphism of adjacent country rocks to pyroxene granulite or pyroxene hornfels facies assemblages, are largely or completely crystalline. The emplacement process has resulted in cataclasis, recrystallization and production of distinctive augen and foliated textures within the peridotites which preserve evidence of initial crystallization of the peridotite mineralogy at pressures in excess of 10 kb and temperatures in excess of 1200°C.

Recent publications (e.g. Kirby, 1979) equating these complexes with "ophiolites" may well be correct if the term "ophiolite" is applicable to any mechanically emplaced (?overthrust, ?obducted) sliver of oceanic crust juxtaposed against continental crust. This nomenclature, however, obscures the important distinction between the high-temperature and relict high-pressure lherzolite mineralogy of these ultramafic complexes and the highly refractory harzburgite tectonite and layered, accumulative dunite, peridotite, pyroxenite and gabbro sequences of the deep-seated portions of "classical" ophiolites. The latter are characteristically low-pressure magmatic sequences and relate to segregation and fractionation of picritic and magnesian quartz tholeiite magmas derived from mantle peridotitic source rocks.

The concept of mantle peridotite diapirism is also an integral part of most discussions of modern basalt petrogenesis. The primary magmas which are parental to the common mid-ocean ridge basalts are considered to be picritic (Table 19-II), i.e. 20—30% melt fractions segregating from residual peridotite at pressures of ~ 20 kb and temperatures of around 1450°C (Green et al., 1979). Extrusion temperatures of common magmas are normally nearer 1250°C (Table 19-II, column 1) and reflect olivine crystallization and removal from primary picritic magmas (liquidus $\sim 1350°C$, 1 atmosphere). There are, however, examples of picritic melts rich in olivine which have quench textures ("spinifex-textures", leading to application of the term komatiitic by some authors) and must have been emplaced at and quenched from temperatures of $\sim 1350°C$ (Gansser et al., 1979). Continued ascent of diapirs of residual peridotite after segregation of tholeiitic picrite magma may produce second stage melts, segregating from residual harzburgite at pressures of 5—8 kb and a temperature of $\sim 1350°C$ (Table 19-II, column 3) (Duncan and Green, 1980, in prep.). Extrusion temperatures for such second-stage melts range from $< 1300°C$ to 1350°C,

TABLE 19-II

Compositions of very magnesian Phanerozoic extrusives

	Olivine tholeiite glass from Atlantic ocean floor (1)	Inferred picritic parent magma for common MORB (2) (cf. column 1)	2nd stage melt of mantle peridotite (parental magma for Troodos upper pillow lavas, Cyprus (3)	High-Mg andesite or "Boninite" (Cape Vogel, Papua-New Guinea) (4)	Howqua Peridotite (Cambrian) Victoria, Australia (5)
SiO_2	49.7	48.3	52.4	57.6	50.9
TiO_2	0.72	0.60	0.30	0.2	0.01
Al_2O_3	16.4	13.7	11.7	7.5	0.73
FeO	7.9	7.9	8.4	10.7	10.4
MnO	0.12	0.12	0.15	0.2	0.19
MgO	10.1	16.7	15.8	17.1	37.2
CaO	13.1	10.9	10.7	5.1	0.58
Na_2O	2.0	1.65	0.70	0.6	0.03
K_2O	0.01	0.01	0.10	0.4	0.03
Cr_2O_3	0.07	0.06	—	—	—
NiO	0.03	0.08	—	—	—
$\frac{100\,Mg}{Mg + \Sigma Fe}$	71	79	77	74	86
Observed liquidus olivine	Fo_{89}	$Fo_{91.5}$	Fo_{91}	[Clinoenstatite Mg_{92}]	Fo_{94}
Liquidus temperature at 1 atmosphere	1230°C	1350°C	1350°C	1360°C	?

(1) DSDP3-18 from Frey et al. (1974); (2) Green et al. (1979); (3) Duncan and Green (1980, in prep.); (4) Dallwitz et al. (1966); (5) Peridotite with cumulate olivine, clinoenstatite and bronzite phenocrysts in a rapidly quenched devitrified matrix (Crawford, 1980).

the lower-temperature magmas being porphyritic with olivine and/or orthopyroxene phenocrysts or fractionated by removal of olivine.

There are other Phanerozoic extrusive rocks which are characterized by very magnesian phenocrysts of clinoenstatite (replacing protenstatite), bronzite or, less commonly, olivine in a quenched matrix. This group includes the "high-magnesium andesites" of Cape Vogel, Papua-New Guinea (Dallwitz et al., 1966), and the Marianas trench (Dietrich et al., 1978) and the "boninites" of the Bonin Islands. These magmas are silica-rich relative to other primary, mantle-derived basalts (cf. Table 16-II). It appears probable that these magmas have their liquidus temperatures lowered by dissolved water at their conditions of origin (work in progress) and their porphyritic character reflects crystallization during ascent, accompanied by boiling-off of water and crystallization en route to the surface. An extreme example of this magma type is possibly the Howqua peridotite (Crawford, 1980). This remarkable rock contains very magnesian olivines (Mg_{94}), having similar Cr_2O_3 and NiO contents but lower CaO than those in olivine of the Archaean peridotitic komatiites. In this largely accumulate rock only the interstitial matrix and zoned phenocryst rims show evidence of rapid quenching. Extrusion temperatures for this rock are as yet not determined but also may be lowered by high water content. These examples of magnesium and silica-rich magmas appear to be characteristic of island-arc settings but the role of water in their petrogenesis and their tectonic significance remains uncertain. There is as yet no compelling evidence that this high-Mg, very high-SiO_2 and porphyritic (clinoenstatite and bronzite phenocrysts) magma type is represented among Archaean magmas. There are some compositional similarities but marked mineralogical and petrographic differences from the Archaean high-Mg, high-SiO_2 spinifex-textured basalt (cf. Table 19-I, column 5, and Table 19-II, column 4).

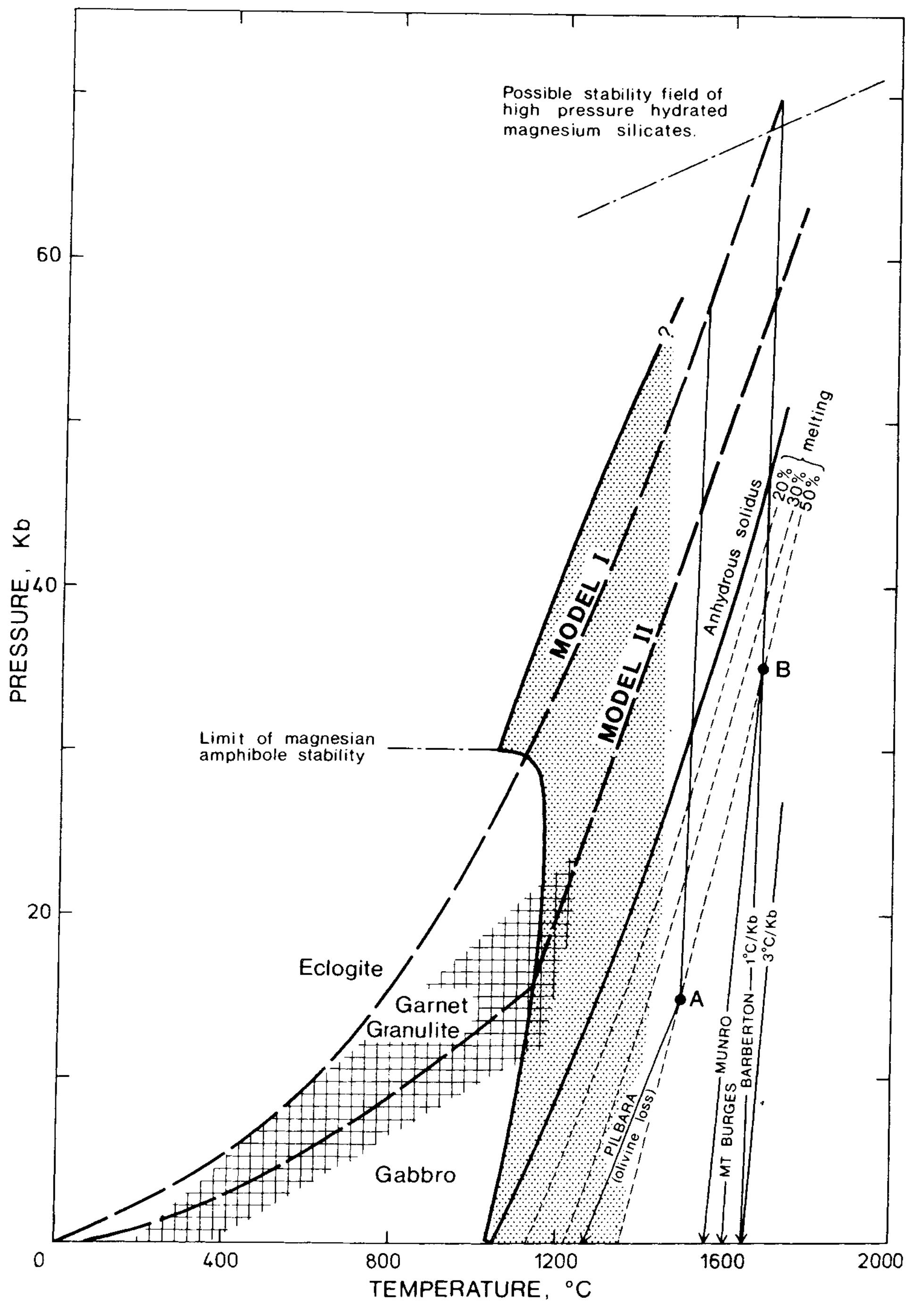
To summarize, the concepts of mantle diapirism, magma segregation at different pressures and temperatures leading to local mantle chemical heterogeneity, and continued ascent of peridotite diapirs after segregation of a melt fraction, are all encompassed within recent models of magma production in the modern and Phanerozoic earth. In particular these concepts are advanced in models of magma genesis at mid-ocean ridge or back-arc basin spreading centres. The principal difference between Archaean ultramafic and basaltic petrogenesis and Phanerozoic ultramafic and basaltic petrogenesis is in the *maximum* temperature of magma extrusion, i.e. ~ 1350°C on the Phanerozoic earth and ~ 1650°C on the Archaean earth.

THE IMPLICATIONS OF VERY HIGH EXTRUSION TEMPERATURES OF ARCHAEAN MAGMAS

The determination of the *P-T* conditions of segregation of a primary magma from mantle peridotite source rock also locates a point on a temperature/depth profile in the earth at the time of magma production. The concept of mantle diapirism, however, implies that this is a point on a perturbed geotherm. To establish a point on the geothermal gradient prior to diapiric upwelling of the upper mantle, it is necessary to extrapolate the *P-T* conditions reached by the peridotite diapir at the depth of magma segregation back to deeper levels — to the depth of origin of the ascending diapir itself. A large, ascending peridotite diapir can be considered, as a first approximation, to follow an adiabatic cooling path as it moves towards the earth's surface, and for a largely crystalline, olivine-dominated peridotite the adiabatic gradient is approximately 1°C/kb (Birch, 1952). If melting occurs, then the latent heat of melting and the higher compressibility of melt relative to crystals will increase the adiabatic gradient. Extrapolation of magma extrusion temperatures to deeper levels at an adiabatic gradient of 1°C/kb thus yields *minimum* values for the temperature distribution in the upper mantle.

The adiabatic ascent paths for peridotite diapirs and segregated magmas may be superimposed in *P-T* diagrams showing the melting behaviour of a suitable model mantle peridotite composition. Experimental studies of two peridotite compositions [pyrolite (Ringwood, 1966) and Tinaquillo lherzolite (Green, 1963)] have established the solidi, % melting contours and liquid and residual phase compositions over the pressure range 0—15 kb (Jaques and Green, 1979, 1980). The data for pyrolite, which is closer to Barberton 49J composition in Na and K contents than is Tinaquillo lherzolite, are summarized in Fig. 19-4. The solidus for pyrolite is strongly dependent on the presence or absence of water and on the water content. Accepting that the interior of the earth is still degassing and that the modern upper mantle contains small water contents ($<$ 0.3—0.4% H_2O), it may be argued that the Archaean upper mantle could not have been anhydrous. The solidus for pyrolite with H_2O ~ 0.1% H_2O is therefore considered most appropriate and this results in an important region of "incipient" or small degree of partial melting between this solidus and the position of the anhydrous solidus. With increasing temperature above the position of the anhydrous solidus, the effect of the water in controlling the melt fraction is progressively less important and the % melt contours approach those for anhydrous melting above the anhydrous solidus. The intersection of minimum adiabatic ascent paths for Barberton, or Mt. Burges, or Yackabindie peridotite magmas with solidi and % melting contours for pyrolite (Fig. 19-4) shows that only at

Fig. 19-4. Implication of very high extrusion temperature for models of Archaean geotherm. Heavy solid lines give solidi for pyrolite compositions under anhydrous conditions and with a water content $\leqslant$ 0.4% H_2O. The fine dashed lines indicate percent melting of anhydrous pyrolite and are approximately correct for $>$ 20% melting if the mantle contains ~ 0.1% H_2O. The heavy dashed lines represent two model geotherms (see text) and the hatched area illustrates conditions of the gabbro—garnet granulite—eclogite reactions. Superimposed on this diagram are the extrusion temperatures and minimum adiabatic ascent paths for peridotite diapirs yielding the magmas of Table 19-I and Fig. 19-1. Points *A* and *B* mark the deduced conditions of magma segregation of Pilbara and Munro compositions, respectively (see text). The shaded, above-solidus region denotes the *P-T* conditions of modern mantle diapirism and magma genesis.


Possible stability field of high pressure hydrated magnesium silicates.
60
40
20
0
PRESSURE, Kb
?
MODEL I
MODEL II
Anhydrous solidus
20%
30%
50%
melting
B
Limit of magnesian amphibole stability
Eclogite
Garnet
Granulite
Gabbro
A
PILBARA
(olivine loss)
MT BURGES
MUNRO
BARBERTON
1°C/Kb
3°C/Kb
400
800
1200
1600
2000
TEMPERATURE, °C

depths of > 200 km would the degree of melting be similar ($< 5\%$, probably 1–2%) to that inferred for the modern low velocity zone (Green and Lieberman, 1976) with its attendant lithosphere/asthenosphere tectonic and dynamic implications. Estimates of the modern geothermal gradient in oceanic regions (e.g. Clark and Ringwood, 1964) yield depths of ~ 400 km for intersection with peridotitic adiabatic ascent paths. The implications of steeper Archaean geotherms, to yield depths of origin of peridotite diapirs of ~ 200 km, were explored by Green (1975). Two possibilities with very different tectonic implications are illustrated in Fig. 19-4.

Model I suggests a geotherm similar to that estimated for ocean basin regions of the modern earth down to depths of ~ 100 km (30 kb) and implies a crystalline lithosphere of ~ 90 km thickness overlying a low velocity zone with 1–2% melt present. As illustrated in Fig. 19-4, this geotherm lies within the eclogite stability field for basaltic compositions. Basaltic rocks, cooling from igneous temperature to conditions on the regional geotherm, should recrystallize to eclogite mineralogy if they remain essentially anhydrous and reaction rates permit. At depths greater than 100 km, the model I geotherm departs from the Clark and Ringwood (1964) geotherm to higher temperatures to give an intersection with adiabatic peridotitic diapir ascent paths at depths of 200–250 km (> 60 kb). The plate tectonics of an Archaean earth with mean geothermal gradient similar to the model I geotherm would probably be much like that of the modern earth — the 90 km lithosphere and 1–2% melting in the low velocity zone being major controlling parameters to lithospheric plate dimensions and relative velocities between lithospheric plates and deeper mantle. In addition, the stability, along the mean geothermal gradient, of dense eclogite ($\rho = 3.4$–3.5) rather than basalt or gabbro ($\rho = 3.0$–3.1) mineralogy would ensure that formation of basaltic crust was only a transitional stage in mantle geochemical differentiation. With a model I geotherm, it would be expected that the Archaean earth should show evidence of subduction of basaltic oceanic crust, of linear zones containing glaucophase schist and eclogite mélanges and linear volcanic/intrusive belts in which an important role for water in magma genesis and evolution from peridotitic source rocks (as opposed to "wet" melting of deep crustal, basaltic or other rocks) could be demonstrated. An important implication of model I geotherm is that highly undersaturated basanitic and nephelinitic magmas, formed by $< 5\%$ melting in the presence of H_2O and $(CO_3)^{2-}$ in the garnet peridotite field, should occur among Archaean intrusives or volcanics, as they do in Phanerozoic and modern volcanics.

Model II suggests a steeper geotherm which intersects the pyrolite + 0.1% H_2O solidus at ~ 50 km, implying a lithosphere of only ~ 50 km thickness. The geotherm passes rapidly into conditions of 5–10% melting of pyrolite and is arbitrarily drawn to higher pressures following approximately the

5% melting content. The combination of thin lithosphere and greater melt fraction in the asthenosphere suggests that an Archaean earth with model II mean geotherm would be characterized by an absence of large, thick, slow moving, lithospheric plates, and by a relatively unstable set of thinner plates, particularly with respect to basaltic volcanism. Basaltic volcanism would be restricted to types derived by $> 5\%$ melting of pyrolite and thus olivine leucitites, olivine melilitites, olivine nephelinites and nepheline-rich basanites should not occur among Archaean extrusives or intrusives.

The second major consequence of a model II geotherm is that this geotherm remains within the stability field of garnet + clinopyroxene + plagioclase ± quartz mineral assemblages in basaltic rocks. Therefore, Archaean basaltic crust overlying peridotitic upper mantle would have no tendency to react to eclogite on cooling and would remain gravitationally stable ($\rho = 3.0$—3.2 overlying $\rho = 3.3$—3.4). The concepts that the gabbro $\rightleftharpoons$ eclogite reaction in thick oceanic crust may trigger subduction (Ringwood and Green, 1966) and that further geochemical processing of basaltic crust is responsible for Andean volcanism, would not be relevant to an Archaean earth if the basaltic crust could not cool sufficiently to transform to eclogite at deeper levels. It has been suggested (Green, 1975) that with a model II geotherm, thermally activated convective overturn of a thin Archaean lithosphere (~ 50 km) would not involve substantial return of basaltic crust to the upper mantle as eclogite but rather the basaltic crustal sequences would be scraped off the underlying convecting peridotite and folded against sialic continental nuclei in a manner analogous to the fate of ocean-floor sedimentary sequences during subduction of modern oceanic crust. If deeper levels of such infolded basaltic sequences (Archaean greenstone belts) attains depths of ~ 25 km then they may partially melt leading to granodioritic and granitic intrusives and associated volcanism. Green (1975) proposed that a model II geotherm, while permitting sialic crust and associated continental processes from sedimentation to deep crustal metamorphism and anatexis, would imply greater movement rates of smaller lithospheric plates to both the thinner lithosphere and the higher melt fraction in the mantle beneath the lithosphere.

In summary, the observations of Phanerozoic and modern volcanism are consistent with processes operating within the shaded area of Fig. 19-4, i.e. a *P-T* region bounded by the solidus for pyrolite + 0.1% H_2O, the extrusion temperatures of $\sim 1350°C$ for primitive, anhydrous magmas, and the adiabat for peridotite diapirs extrapolated from 20 kb, 1450°C (depth of magma segregation of tholeiitic picrite parental to MORB). The mean geothermal gradient for oceanic regions for the modern earth passes through ~ 30 kb, 1100°C, implying a lithospheric thickness of ~ 90 km and a low velocity zone or asthenosphere with 1—2% melt (composition of highly undersaturated olivine melilitite or olivine nephelinite). The voluminous tholeiitic volcanism of Phanerozoic and modern basaltic provinces results from perturbation of the mean geothermal gradient by diapiric upwelling of peridotite diapirs from depths of 100—180 km, such upwelling occurring beneath rifted plate margins in mid-oceanic ridges or back-arc basins. Although convective motion of deeper mantle levels is implied in the general plate-tectonic model, the character and rate of such movements is not such as to provide sufficient instability to produce *rapidly ascending* and thus adiabatically cooling peridotite diapirs from deeper ($>$ 180 km), higher temperature, levels of the upper mantle. This statement derives directly from the apparent absence among Phanerozoic or modern extrusives of anhydrous magmas extruded at temperatures above 1350°C and of anhydrous liquids of more olivine-rich composition than picrite.

In contrast, Archaean basaltic and ultramafic volcanism requires upwelling and peridotitic diapirism which tap deeper levels at higher temperatures, extending *at least* to the minimum (1°C/kb) Barberton adiabat of Fig. 19-4 and probably to higher temperatures and greater depths. It is possible that the Archaean thermal regime reflects an increase in the efficiency and magnitude of *convective* heat loss from the deep mantle, i.e. heat lost through mantle mass transport and consequent volcanism. If this is so then the mean Archaean geotherm, in non-volcanic areas, and consequent lithosphere thickness may not be very different from those of the modern earth, [model I geotherm of Fig. 19-4). It is more probable, however, that increased heat loss is both conductive, reflected in a higher mean geothermal gradient, and convective [model II geotherm of Fig. 19-4). In this case the Archaean lithosphere would be much thinner ($\sim$ 50 km), crustal and lithospheric temperatures higher and the nature of Archaean basaltic magmatism restricted to hypersthene normative or mildly undersaturated basaltic magmas (i.e. excluding olivine melilitites, olivine nephelinites and basanites). This model also implies that Archaean basaltic crust overlying residual mantle peridotite would not react to eclogite (with consequent gravitational instability), and thus would not return via a subduction-like process for further geochemical processing in the upper mantle.

SUMMARY AND CONCLUSIONS

Extrusion temperatures of South African and Western Australian Archaean peridotitic komatiite magmas were greater than 1600°C and the mineralogy of the residue from magma segregation is highly magnesian olivine (Mg_{93-94}) alone. Peridotitic komatiite from Canada has slightly lower extrusion temperature ($\sim$ 1540°C) and is inferred to have segregated from residual harzburgite at $\sim$ 35—40 kb, $\sim$ 1700°C. The immediate source compositions for these high-temperature liquids differed in minor and trace element compositions, and the combination of experimental petrological and geochemical data suggests that the Archaean upper mantle was as geochemically heterogeneous as the modern upper mantle, some regions having been enriched and others depleted in incompatible trace elements. Archaean magma genesis can be interpreted in terms of mantle diapirism, i.e. similar processes to those operating in modern mid-ocean ridge or back-arc basin spreading centres. The principal differences lie in the maximum temperatures of extrusion of Archaean magmas which are 300—350°C above those of modern or Phanerozoic magmas. The different temperature regime for magma extrusion implies differences between the Archaean geothermal gradient and that of the modern earth. By superimposing model Archaean geotherms on the known high-pressure melting behaviour of model upper mantle compositions, two different models for Archaean lithosphere/asthenosphere can be presented.

Model I implies a lithosphere of similar thickness (~ 90 km) to that of the modern earth and predicts similar processes of eclogitization of basaltic crust and of subduction of oceanic crust. This model predicts that intrusives, extrusives or differentiates of highly undersaturated magmas (olivine nephelinites, olivine melilitites etc.) should occur within the Archaean. The principal differences from the modern earth are in higher temperatures within the upper mantle and a more efficient heat loss by convective processes (i.e. more active mantle diapirism and consequent magmatic activity).

Model II implies a thin Archaean lithosphere (~ 50 km) and higher subcrustal and lithosphere temperatures such that basaltic rocks do not react to eclogite on cooling to the geotherm. This model predicts that Archaean basaltic crust will not be subducted as eclogite for further geochemical differentiation in the upper mantle but rather will be folded against and between sialic continental nuclei (Archaean granite/greenstone terranes). Model II further predicts that Archaean basaltic volcanism or intrusives and their differentiates will be restricted to hypersthene-normative or mildly undersaturated liquids, i.e. exclusive of olivine-rich basanites, olivine nephelinites etc. Model II predicts a different type of tectonics in the Archaean with much more convective heat loss by higher temperature diapirism from deeper levels of the upper mantle (200—250 km minimum for source depth of Archaean diapirs). The transition from a model II Archaean earth to the modern earth would be gradual and transitional except that the onset of eclogite stability along mean geothermal gradient might be expected to have an important signature within the geological record as a change from granite/greenstone terranes to appearance of eclogite-bearing mélanges, absence of thick, synclinal volcanic sequences dominated by basalt or basalt + peridotite and pronounced linearity of metamorphic/magmatic orogenic belts.

The interpretation of the distinctive Archaean magmatism in terms of mantle convection and heat loss has been the focus of this paper. However, it is possible that, in seeking understanding of processes within the earth prior to 3.5 Ga ago, the effects of major impacts on the earth's surface must be considered (Green, 1972). The energy impacted to enter regions of the earth's crust by impact phenomena in the 4.5 to 4.0 Ga period and the reaction of the earth to major impacts such as produced Mare Orientale on the moon must be considered in earth models which encompass the geology of the earliest Archaean.

With continuing programmes of field geology, geochronology and geochemistry of Archaean terranes it should be possible to test the relevance of model I or model II geotherms by the predictions on eclogite stability, geochemistry of basaltic magmas etc. which are intrinsic to each model. Similarly the applications of lunar impact chronology to the early history of the earth may lead to predictions which are testable in areas of the oldest Archaean shields.

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Chapter 20

ISOTOPIC EVIDENCE FOR CONTINENTAL GROWTH IN THE PRECAMBRIAN

S. MOORBATH and P. N. TAYLOR

ABSTRACT

Radiogenic Sr, Pb and Nd isotopic data demonstrate that continental growth by irreversible differentiation of the upper mantle throughout geological time predominates over recycling and reworking of older sialic crust. Together with much other geochemical and geophysical evidence, age and isotope data provide a simple, dynamic picture of the formation, metamorphic/geochemical differentiation and stabilization of continental crust from c. 3700 Ma onwards, at the same time contradicting those hypotheses which involve repeated recycling and regeneration of ancient continental protoliths, massive recirculation of continental crust throughout the mantle, and unrestricted mixing between continental crust and mantle.

Isotopic data are generally diagnostic of the source of a magmatic rock, but superimposed isotopic perturbations observed in many continental igneous rocks may result from complex crust—magma interaction processes. Of the various radiogenic isotopes uranogenic Pb is particularly versatile in its applications, and can be used to recognize and characterize primary mantle-derived sial, reworked or partially melted older sialic material, geochemically reactivated older sialic crust, or ancient sialic basement at depth.

Whilst initial isotopic compositions of Precambrian calc-alkaline orthogneisses generally indicate extraction of their magmatic precursors from mantle or associated oceanic lithosphere, followed by penecontemporaneous metamorphic/geochemical differentiation, true granites frequently have initial isotopic compositions suggesting derivation by partial melting of continental crust.

INTRODUCTION

The geochemistry of the radiogenic isotopes (^{87}Sr, ^{143}Nd, ^{206}Pb, ^{207}Pb and ^{208}Pb) is diagnostic of the magmatic source of a rock suite, but not directly of the tectonic environment of formation. Whether or not plate-tectonic regimes dominated in the Precambrian can be solved only by considering a wide range of evidence from diverse disciplines. However, geochronological and related isotopic data provide important constraints for deciding whether a given igneous rock suite or orthogneiss complex has been produced by melting of upper mantle material and/or basic lithosphere, or by melting of older continental crust. This is clearly directly relevant to the much-debated problem of whether or not the continental, sialic crust has increased in mass throughout geological time.

It is now widely recognized that calc-alkaline orthogneisses, which largely constitute the Precambrian shield areas, are the fundamental building blocks of the continental crust and are derived from the upper mantle or from crustal materials with short crustal residence times. This subject has been extensively reviewed recently (e.g. Windley and Smith, 1976; Moorbath, 1977, 1978a; Brown, 1977; Brown and Hennessy, 1978; O'Nions and Pankhurst, 1978; Barker, 1979; Glikson, 1979; Tarney et al., 1979). Furthermore, the thickness of Precambrian continental crust, the magnitude of Archaean continental geotherms and the characteristic continental distribution of the radioactive heat-producing elements (i.e. exponential decrease with depth) were all much the same in Archaean times as they are now (e.g. Bickle, 1978; Burke and Kidd, 1978; Brown and Hennessy, 1978; Tarney and Windley, 1977; Davies, 1979; England, 1979; Wells, 1979). Although radiogenic heat production in the mantle in late Archaean times, some 3000—2500 Ma ago, was about three times greater than that of today, the extra radiogenic heat was not removed by conductive transfer through stabilized continental crust. Thus, Burke and Kidd (1978) state that a "convective process seems necessary to dissipate this heat and plate tectonics is an efficient and familiar convective process capable of doing the job". Bickle (1978) considers that "the relation between heat loss, the rate of plate creation and the rate of heat transport to the base of the lithosphere suggest that a significant proportion of the heat loss in the Archaean must have taken place by the processes of plate creation and subduction. The Archaean plate processes may have involved much more rapid production of plates only slightly thinner than at present". Similarly, England (1979) concluded that "the lower geothermal gradients permitted by the *P-T* data would be consistent with the view of Archaean tectonics in which the continental thermal regime was similar to that of the present day and most of the additional heat was lost from the earth's interior by a faster rate of creation of oceanic lithosphere". All these considerations are in close accord with the constraints imposed by geochemical, geochronological and isotopic evidence.

Isotopic data on calc-alkaline igneous rocks of all ages, including Precambrian orthogneisses, suggest that irreversible chemical differentiation of the mantle commencing at least c. 3700 Ma ago has produced new continental sialic crust during several relatively short (*c.* 100—300 Ma) episodes which, in any given region, were usually widely separated in time and may have been broadly of global extent. During each of these so-called "continental accretion-differentiation superevents" (CADS; see Moorbath, 1975a, b, 1977, 1978a), juvenile sial derived from the mantle and/or basic lithosphere, underwent thorough nearly contemporaneous igneous, metamorphic and geochemical differentiation, to produce thick, stable, permanent, continental crust, which has a gradational geochemical stratification, and which exhibits close grouping of isotopic ages of rock formation, as well as mantle-type, initial Sr, Pb and Nd isotopic compositions for all major constituents. Isotopic data suggest that within most CADS, and especially during the earlier Precambrian ones, continental growth greatly predominated over reworking of older, sialic crust.

Nonetheless, reworking of much older sialic crust *does* occur in several types of tectonic environment and certainly becomes more common with the progressive growth and fragmentation of continental crust throughout geological time, and with the independent migration of the separated fragments. Reworking of older continental crust is usually distinguishable from the CADS regime in any given terrain from detailed age and isotope studies. On purely petrological grounds, true granites are more likely to be the magmatic expression of partial melting of sialic crust than are any other members of the calc-alkaline suite. Furthermore, true granites (of any geological age) tend to have higher and much more variable initial $^{87}Sr/^{86}Sr$ ratios than contemporaneous calc-alkaline rocks, an expression of the relatively higher proportion of continental crustal involvement in their genesis. The degree of isotopic divergence of a given granite from contemporaneous, mantle-derived, calc-alkaline or basaltic rocks is a function of the degree of crustal involvement, the geochemical nature of the crust and the magnitude of the age difference between the granite and its crustal protolith. Where the age difference is small, for example, initial isotopic compositions of crustally derived granites may approximate closely to values for contemporaneous mantle-derived rocks. In cases where granitic magma is formed by direct fractionation of mantle-derived basaltic or intermediate magmas, initial isotopic compositions may be identical within analytical error with contemporaneous mantle. There are many possibilities and it follows that the isotopic data on each granitic rock must be carefully interpreted on its own merits. True granites are petrogenetically much more diverse than most other igneous rocks because their parent magmas may be derived from mantle, oceanic lithosphere, continental sialic crust, or any combination of these. In contrast, petrogenetic (and isotopic) data show that the continent-building tonalites and their calc-alkaline plutonic and volcanic brethren have a more restricted range of parental material in the upper mantle or in oceanic lithosphere (see, for example, Ringwood, 1974; Stern et al., 1975; Thorpe et al., 1976; Anderson et al., 1978; Hanson, 1978).

Isotopic and geochemical evidence indicate that the volcanic components of the complex volcano-sedimentary ("supracrustal") Archaean greenstone belt assemblages were largely derived from the mantle (for review, see O'Nions and Pankhurst, 1978). The unifying term "greenstone belt" has hampered a closer understanding of these ubiquitous components of Precambrian continental crust. These ancient supracrustal belts were formed in a very wide range of tectonic environments, including deposition on: (a) older continental, sialic crust (*e.g. Henderson, this volume, Chapter 9, Nisbet et al., this volume, Chapter 7, ed.)*; (b) basic ("proto-oceanic"?)

crust away from the influence of continental crust (*e.g. Anhaeusser, this volume, Chapter 6, ed.*); and (c) basic ("proto-oceanic"?) crust near a continental margin or in a back-arc basin and thus accessible to continental detritus (*e.g. Barton and Key, this volume, Chapter 8, ed.*).

Thus it is evident that there are "greenstone belts and greenstone belts", and an appropriate multiplicity of tectonic hypotheses must be applicable. A marginal basin model seems particularly appropriate for some greenstone belts on or near continental crust (Tarney et al., 1976). Some of the oldest known supracrustal assemblages may have pre-dated continental crust altogether in their particular region. An example of this type may be provided by the c. 3800 Ma old Isua supracrustals of West Greenland (Moorbath et al., 1973; Allaart, 1976; Bridgwater et al., 1976; Michard-Vitrac et al., 1977; Hamilton et al., 1978), deposited in a shallow-water environment dominated by basaltic and rhyolitic lavas, near-shore clastic-volcanigenic sediments, as well as chemical sediments.

Ancient supracrustal assemblages range from virtually unmetamorphosed right up to medium/high metamorphic grade. In the latter case, they are often found as massive enclaves or dismembered inclusions in enveloping calc-alkaline orthogneisses. Few major gneiss terrains are free of such xenoliths of supracrustal parentage. The magmas from which the gneisses crystallized clearly invaded and buried the supracrustals during major continental accretion, most probably by the process of "overaccretion" envisaged by Wells (1979, 1980).

All these factors must be borne in mind when interpreting the complex geochronological and isotopic relationships which have been observed within and between greenstone belts and their surrounding orthogneiss terrains. In the simplest cases, close grouping of age and isotope data within a given Archaean "gneiss—greenstone" terrain demonstrates that both may result from a single CADS, extending over a period of up to c. 100—200 Ma (e.g. Arth and Hanson, 1975; Hawkesworth et al., 1975; Jahn and Murthy, 1975; Roddick et al., 1976).

Below we summarize some recent applications of isotopic studies, particularly to Precambrian continental evolution. Descriptions of the basic principles of the isotopic methods may be found in Faure and Powell (1972), Faure (1977), O'Nions et al., (1979) and Köppel and Grünenfelder (1979).

MAGMA—CRUST INTERACTION, SELECTIVE CONTAMINATION AND ISOTOPIC MIXING

There is a complicating factor in the interpretation of isotopic data which may have to be taken in account when considering mantle-derived magmatic rocks emplaced into, onto, or through older continental, sialic crust. Continental igneous rocks often exhibit Sr, Pb and Nd isotope systematics of far greater complexity than oceanic igneous rocks. Numerous conflicting interpretations have been given for this observation: different workers attribute the problem to isotopic heterogeneity within the upper mantle, within the continental crust, or within the sub-continental lithosphere ("tectosphere", Jordan, 1978), or a combination of these (e.g. Brooks et al., 1976a; Faure, 1977; Pankhurst, 1977; Carter et al., 1978a).

In our view, systematics in isotopically perturbed, continental igneous rocks are most frequently attributable to crust—magma interaction, resulting in the mixing of incompatible elements (including those with radiogenic isotopes) derived from mantle-derived magma and from surrounding continental crust. Bulk mixing of magma and crust (including bulk assimilation of crust by magma) is frequently ruled out from petrological major and trace-element data, so that selective contamination and mixing processes involving only incompatible trace elements are postulated (e.g. Moorbath and Welke, 1969a; Carter et al., 1978a, b; Briqueu and Lancelot, 1979; Moorbath and Thompson, 1980). The mechanism probably involves mixing of mantle-derived magmas with a fluid phase containing incompatible elements derived from hydrous minerals of the country rock in close proximity to magma conduits and chambers. Provided that sufficient heat is available, the process can go further and, in principle, lead to partial melting of country rock (Wells, 1979, 1980; Patchett, 1980). However, this process is not likely to happen, if the country rock surrounding a magma chamber is completely dry (and remains so), refractory and geochemically depleted, such as might be expected in the case of some granulite facies gneisses.

The magnitude of isotopic perturbation in any given continental igneous rock appears to be strongly dependent upon the thickness, age and geochemistry of the invaded or traversed crust. Regardless of the actual mechanism of crust—magma interaction, radiogenic Sr, Nd and *particularly* Pb isotopes can act as very sensitive tracers for this process, whilst Pb isotopes may also yield an approximate age for the rocks which supplied the continental Pb component, even where these are not exposed at the surface. The basic principles of this approach have been recognized for many years as applied to ore leads (see Russell and Farquhar, 1960). It is evident that Pb isotope systematics in common igneous silicate rocks can be similarly interpreted. Mostly, such radiogenic isotope studies have been carried out on Phanerozoic and Recent continental igneous rocks which have been emplaced into, or which have traversed, ancient continental basement (e.g. Moorbath and Welke, 1969a; Zartman, 1974; Armstrong et al., 1977; Leeman and Dasch, 1978; Lipman et al., 1978). The rather commonly observed combination of *radiogenic* Sr with *unradiogenic* Pb testifies to interaction of magma with ancient continental crust which has maintained correspondingly high Rb/Sr and low U/Pb ratios ever since its formation. Such crust can probably be broadly identified as amphibolite facies gneisses.

It follows that radiogenic isotope analyses of continental igneous rock of any type can sometimes be used to detect ancient continental basement at depth, even where this is not exposed at the surface (e.g. Moorbath and Welke, 1969b; Zartman, 1969; Blaxland et al., 1979). Conversely, the *absence* of continental crust from oceanic areas is clearly demonstrable from isotopic measurements, although the fine structure of radiogenic isotope data has proved a remarkably powerful tool for the spatial and temporal study of geochemical heterogeneity in the upper mantle (e.g. Sun and Hanson, 1975; Brooks et al., 1976b; Hofmann and Hart, 1978; O'Nions et al., 1978). We maintain that the isotopic effects of crust—magma interaction, resulting from selective contamination and trace-element mixing processes, are a powerful tool for the detection, characterisation and sub-surface "mapping" of ancient continental crust at depth, and for studying the extent and growth of such crust. Later on, we cite the application of Pb isotopes to determining the sub-surface extent of early Archaean (c. 3700 Ma) continental crust beneath late Archaean (c. 3000—2800 Ma) rocks in West Greenland.

Sr ISOTOPES

Many major Precambrian calc-alkaline orthogneiss terrains yield good Rb-Sr whole-rock regression lines over hundreds, even thousands, of square kilometres. Generally the smaller and more petrologically uniform the gneiss unit sampled, the closer the approximation to good isochron fit. Initial $^{87}Sr/^{86}Sr$ ratios [subsequently abbreviated to $(^{87}Sr/^{86}Sr)_i$] for most gneisses either coincide with, or lie very slightly above, contemporaneous upper-mantle values as deduced from plausible upper-mantle evolution models (e.g. Faure and Powell, 1972; Hart and Brooks, 1977; Hurst, 1978; Peterman, 1979). We cannot exclude a small degree of mantle heterogeneity in Precambrian times as the cause of some of this Sr isotopic variation. However, measured $(^{87}Sr/^{86}Sr)_i$ of calc-alkaline orthogneisses in high-grade Precambrian terrains are not necessarily identical within analytical error with their mantle source regions because there may be a significant time interval between magma formation and final closure of orthogneiss Rb-Sr whole-rock systems (Moorbath, 1978b).

Archaean and Proterozoic gneisses commonly yield $(^{87}Sr/^{86}Sr)_i$ in the range ~ 0.700 to 0.703 (e.g. Moorbath, 1975a, 1977; O'Nions and Pankhurst, 1978; Peterman, 1979), whilst there is good evidence from measurements on 2700 Ma-old unaltered, mafic volcanics that in at least one upper-mantle magma source the $^{87}Sr/^{86}Sr$ ratio at that time was close to 0.7011 (Hart and Brooks, 1977). It is widely agreed that low $(^{87}Sr/^{86}Sr)_i$ of Precambrian calc-alkaline orthogneisses preclude derivation by reworking of significantly older rocks with typical upper crustal Rb/Sr ratios, and that in such cases, whole-rock Rb-Sr isochrons broadly date crustal accretion events.

In our view, the main reason for the observed difference between penecontemporaneous juvenile crust and mantle $^{87}Sr/^{86}Sr$ ratios results from the geochronologically short, but finite, time interval occupied by the processes of genesis and crustal accretion of juvenile sial, with concomitant petrological, geochemical and metamorphic differentiation. The juvenile, calc-alkaline sial clearly has higher Rb/Sr ratios (typically by a factor of about 10) than its basic or ultrabasic source region, whilst any plausible multistage process of crustal accretion and internal differentiation, such as that proposed by Wells (1979, 1980), must take at least several tens, and more likely 100—200, Ma. The actual date recorded by a whole-rock Rb-Sr isochron represents the time at which the rocks become closed systems to migration of Rb and Sr on the scale of sampling of analyzed whole-rock specimens. Significant scatter of $(^{87}Sr/^{86}Sr)_i$ within different orthogneiss units belonging to one major CADS are possibly attributable to: (a) Sr isotope heterogeneity of the upper mantle; (b) redistribution of Sr within earlier intrusive units as a result of the thermal overprint of later intrusions; (c) selective contamination of later intrusive units by earlier ones, with resulting isotopic mixing; and (d) progressive metamorphism and accompanying geochemical differentiation within the newly accreted crustal segment as a whole, causing Rb-depletion in the lower crust (decrease of Rb/Sr ratios) and Rb-enrichment (increase of Rb/Sr ratios) in the upper crust, leading to the establishment and ultimate freezing-in of a Rb/Sr gradient in the "ripened" crust.

In this way, it is easy to see how small, but significant, variations in $(^{87}Sr/^{86}Sr)_i$ can be established within newly formed crustal components during a complex, multistage CADS extending typically over a period of up to c. 100—200 Ma. In detail, there is an almost infinite range of possibilities, but the overall trend is towards a more rapid and variable evolution of $^{87}Sr/^{86}Sr$ in the newly formed crust than in contemporaneous mantle, whilst the chance of any individual crustal unit preserving a contemporaneous mantle $^{87}Sr/^{86}Sr$ value is quite small. From the above discussion, the lowest $(^{87}Sr/^{86}Sr)_i$ value for a suite of more or less contemporaneous orthogneisses is probably closest to the $^{87}Sr/^{86}Sr$ in the contemporaneous mantle source region.

Thus, observed variations in $(^{87}Sr/^{86}Sr)_i$ of up to c. 0.001—0.002 within different crustal components formed in a single CADS can be accounted for by complex, short-term, multistage processes of the type outlined above, although with an enhancement in $(^{87}Sr/^{86}Sr)_i$ of c. 0.001—0.002 above the contemporaneous upper-mantle value the possibility of limited involvement of much older continental crust cannot be rigorously excluded.

Amphibolite facies calc-alkaline orthogneisses typically have Rb/Sr ratios in the range c. 0.2—0.4, on average some ten times higher than the upper mantle, with proportionally enhanced growth rates in $^{87}Sr/^{86}Sr$. In contrast, the average Rb/Sr ratio of deep-seated granulite-facies (Rb-depleted) continental crust is usually less than c. 0.04, reflecting the widely reported phenomenon of Rb depletion in granulite facies gneisses which many consider to make up the lower continental crust (e.g. Heier, 1973; Tarney and Windley, 1977). It follows that the average $^{87}Sr/^{86}Sr$ ratio of granulite facies, low Rb/Sr, continental crust is close to that of contemporaneous upper mantle, *if* the granulite facies assemblages date from the time of the CADS in which their protoliths were formed. Thus the present-day $^{87}Sr/^{86}Sr$ of many of the late Archaean (c. 2900—2700 Ma) Lewisian granulites of northwest Scotland are in the range 0.7021—0.7034 (Rb/Sr = 0.003—0.01, Chapman, 1978). If these gneisses were partially melted *today*, the resulting magmas would have $(^{87}Sr/^{86}Sr)_i$ which are in the same range as present-day mantle-derived rocks. In such a case, Sr isotopes cannot distinguish between an upper-mantle or deep-crustal origin for a given igneous rock, however improbable it may be on petrological and geochemical grounds to account for the production of vast amounts of juvenile, calc-alkaline magma from deep-seated, depleted, dry, geochemically barren continental crust. The use of Pb and Nd data is required to resolve this type of problem.

Notwithstanding the above interpretative complications, Fig. 20-1a and 1b present standard $^{87}Sr/^{86}Sr$ evolution diagrams for selected groups of Precambrian amphibolite facies calc-alkaline orthogneisses from West Greenland and Zimbabwe, respectively. The mantle growth line is that recommended by Peterman (1979), although its exact shape or position is not crucial to the applications discussed here. The crustal growth lines are based on average Rb/Sr ratios for individual gneiss units. Following customary methods of Sr isotope interpretation (e.g. Faure and Hurley, 1963; Faure and Powell, 1972; Moorbath, 1975a, 1975b) the gneisses plotted in Fig. 20-1a, b (as, indeed, many similar gneisses in the published literature) represent juvenile crustal additions at the times indicated, and cannot have been derived by reworking or partial melting of much older, high-Rb/Sr crust.

Figure 20-1a, b also shows growth lines for two Archaean granites (sensu stricto), namely the 2530 ± 30 Ma old Qôrqut granite of West Greenland (Moorbath et al., in press), and the 3340 ± 60 Ma old Mont d'Or granite of Zimbabwe (Moorbath et al., 1976). These granites have high $(^{87}Sr/^{86}Sr)_i$ of 0.7081 ± 0.0008 and 0.711 ± 0.001, respectively, and probably represent partial melts of c. 1100 Ma older and c. 200 Ma older continental basement rocks in their respective areas. In this connection, Pb isotopes in the Qôrqut granite are important and will be discussed later. For general comparison Fig. 20-1a also shows the average $(^{87}Sr/^{86}Sr)_i$ of 0.7040 for 86 samples of Recent calc-alkaline volcanics from Mexico, Costa Rica, Ecuador, Central Chile (Francis et al., 1977; Moorbath et al., 1978; Thorpe et al., 1979; Hawkesworth et al., 1979c; Déruelle and Moorbath, unpubl. data). In these areas of normal (c. 30 km) crustal thickness, where effects of crustal isotopic contamination are minimal, the rather

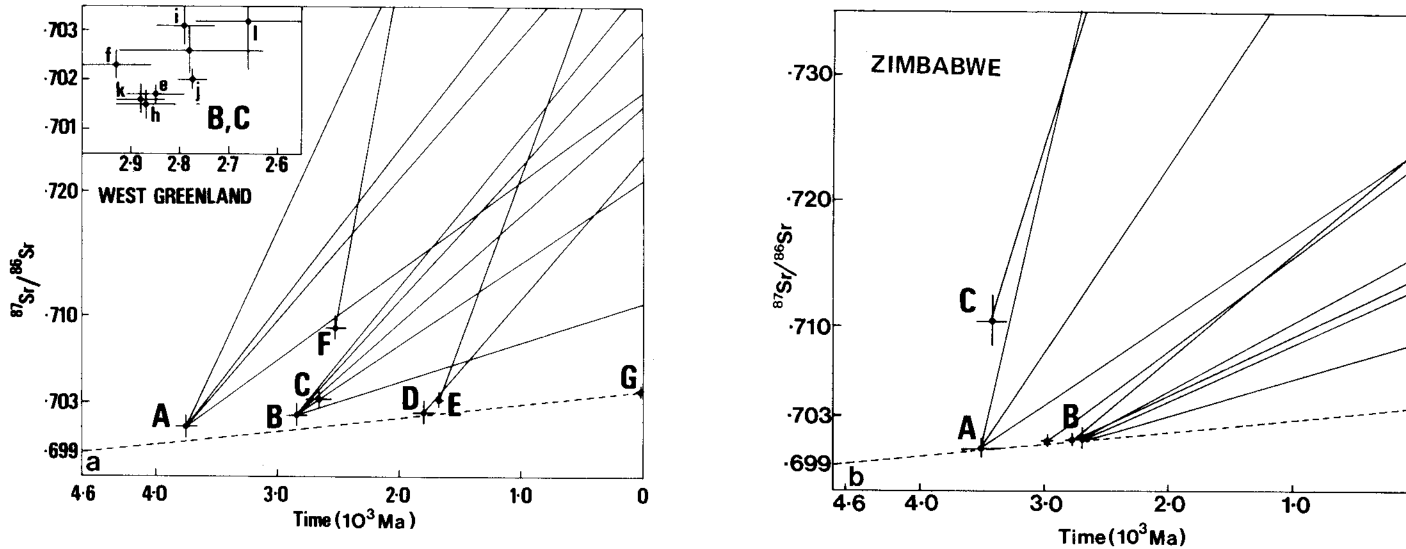

Fig. 20-1. a. Sr isotope evolution diagram for West Greenland: *A* = Amîtsoq gneisses (Moorbath et al., 1972); *BC* = Nûk gneisses and their equivalents (Moorbath and Pankhurst, 1976); *DE* = Ketilidian gneisses of South Greenland (Van Breemen et al., 1974); *F* = Qôrqut granite (Moorbath et al., in press). Inset at top left shows fine structure of initial $^{87}Sr/^{86}Sr$ ratios for Nûk gneisses and equivalents (Moorbath and Pankhurst, 1976). The mantle evolution line (0.6990—0.7039) is that recommended by Peterman (1979). Point *G* is the average (0.7040) of numerous isotopically uncontaminated Andean calc-alkaline volcanics from Central and South America (for references and further details, see text). Decay constant $\lambda_{87} = 1.42 \times 10^{-11}\ a^{-1}$.
b. Sr isotope evolution diagram for Zimbabwe: *A* = early Archaean gneisses (Hickman, 1974; Hawkesworth et al., 1975; Moorbath et al., 1977); *B* = late Archaean gneisses, granites and greenstone belt volcanics (Hawkesworth et al., 1975; Moorbath et al., 1977; Hawkesworth et al., 1979a); *C* = Mont d'Or granite (Moorbath et al., 1976).

narrow range of $(^{87}Sr/^{86}Sr)_i$ of 0.7036—0.7044 is virtually identical with that of calc-alkaline volcanics from oceanic regions and clearly close to the range of present-day mantle values. In agreement with Briqueu and Lancelot (1979), we regard the higher $(^{87}Sr/^{86}Sr)_i$ values mostly in the range of 0.705—0.708 of calc-alkaline volcanics from Southern Peru and Northern Chile (James et al., 1976; Francis et al., 1977; Klerkx et al., 1977) as due to selective crustal contamination of rising calc-alkaline magmas with incompatible elements derived from underlying thick (up to 70 km), ancient continental crust in these regions (Cobbing et al., 1977; Dalmayrac et al., 1977; Halpern, 1978; Shackleton et al., 1979; Thorpe and Francis, 1979). In this general connection, Windley and Smith (1976) regard the calc-alkaline batholiths of the Andes — presumably genetically connected in a broad sense with the overlying and surrounding calc-alkaline volcanics — as the geologically much younger high-level analogues of the ancient, high-grade gneiss complexes.

Pb ISOTOPES

Recognition of crustal accretion—differentiation superevents (CADS)

Many major Precambrian calc-alkaline orthogneiss terrains yield excellent whole-rock $^{207}Pb/^{206}Pb$ isochrons, which frequently agree within analytical error with corresponding Rb-Sr whole-rock isochron ages, although not nearly as many Pb/Pb data have been published. Simplicity of Pb/Pb isotope systematics in such cases suggests derivation of the magmatic precursors from a source region with a fairly uniform $^{238}U/^{204}Pb$ ratio, approximating to single-stage evolution from time of formation of the earth to the measured isochron age. Calculated $^{238}U/^{204}Pb$ ratios (μ_1 values) for the sources of parent magmas for many Archaean orthogneisses fall in the range from 7.5 to 8.0 (model parameters as used in Oversby, 1974). This range is considered characteristic of the upper mantle, so that gneisses with indicated μ_1 values in that range are assumed to have their origins in upper mantle or basic lithosphere (see Table 20-I).

In Fig. 20-2 Pb/Pb data for c. 3700 Ma old Amîtsoq gneisses and c. 2900 Ma old gneisses of West Greenland (outside the Godthaab area) are plotted in relation to a single-stage growth curve. Both $^{207}Pb/^{206}Pb$ isochrons indicate a required first-stage μ_1 value of 7.5 for their parent magma source region. Thus the $^{238}U/^{204}Pb$ ratio of the mantle source region of the magmatic precursors of the gneisses apparently remained constant throughout the Archaean in this region. Many other Archaean gneisses from different regions yield similar μ_1 values for their source region, as shown in Table 20-I. Each Pb/Pb isochron broadly dates a CADS, which comprises extraction of parent material from the upper mantle, right through to final stabilization of a new segment of continental crust by metamorphic and

TABLE 20-I

Pb-Pb isochron data for Archaean calc-alkaline orthogneisses[1]

Region	Age (Ma)	$^{238}U/^{204}Pb$ of source (μ_1)	No. of points	MSWD	Reference
West Greenland					
Amîtsoq gneisses, Godthaab area	3770 ± 160	7.62	27	21	Black et al., 1971; Oxford unpublished data
Amîtsoq gneisses, Isua area	3700 ± 80	7.58	7	4.8	Moorbath et al., 1975b
Nûk gneisses and equivalents:					
Nordland-Sukkertoppen granulites	3000 ± 80	7.60	19	2.3	Taylor et al., 1980
Sermilikfjord	3000 ± 80	7.48	11	1.5	Taylor et al., 1980
Fiskenaesset	2820 ± 70	7.48	9	1.2	Taylor et al., 1980
East Greenland					
Kangerdlussuaq area	2980 ± 60	8.08	7	5.6	Leeman et al., 1976
Kangerdlussuatsiaq area	2750 ± 90	7.77	7	1.5	Bridgwater et al., 1978
Labrador					
Uivak gneisses	3560 ± 80	7.48	11	6.9	Baadsgaard et al., 1979; Bridgwater and Taylor, unpubl. data
North Norway (Lofoten—Vesterålen)					
Northwest Hinnøy amphibolite facies gneisses	2690 ± 60	7.91	22	2.3	Griffin et al., 1978
Northwest Scotland					
Southern Outer Hebrides amphibolite facies gneisses	2600 ± 80	7.26	13	5.7	Moorbath et al., 1975a
Scourian granulites, Scourie area	2680 ± 60	7.66	20	3.0	Chapman and Moorbath, 1977
Karnataka craton, Southern India					
Chikmagalur area	3200 ± 30	7.95	9	2.5	Oxford unpublished data
Chitradurga area	3030 ± 30	7.62	8	2.4	Oxford unpublished data
Wyoming U.S.A.					
Granite Mountains	2750 ± 70	8.27	17	"good fit"	Rosholt et al., 1973

[1] Model parameters: $\lambda_{235} = 0.98485 \times 10^{-9}\ a^{-1}$; $\lambda_{238} = 0.155125 \times 10^{-9}\ a^{-1}$; $^{238}U/^{235}U = 137.88$; age of earth $= 4.57 \times 10^9$ a; primordial Pb in Canyon Diablo troilite, $^{206}Pb/^{204}Pb = 9.307$; $^{207}Pb/^{204}Pb = 10.294$.

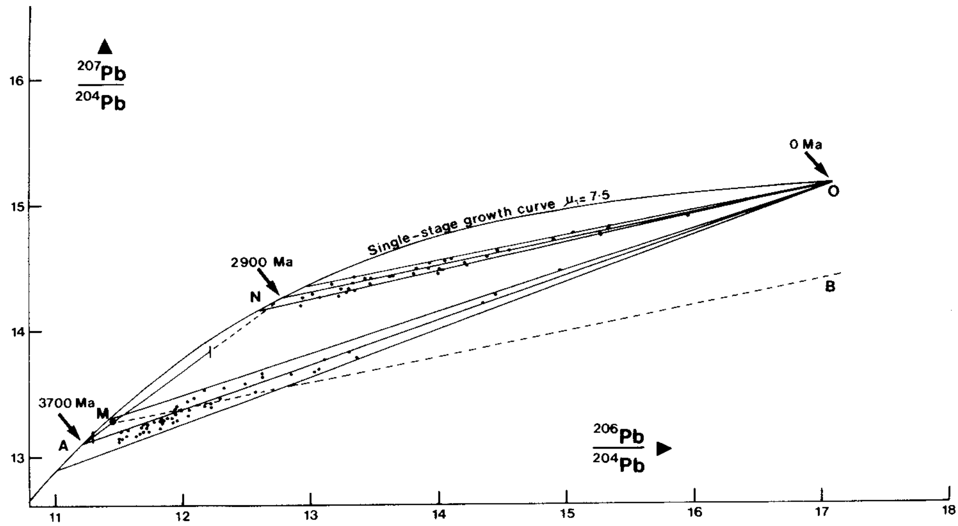

Fig. 20-2. Pb-Pb diagram for data from Amîtsoq gneisses (*AO*) and Nûk gneisses (*NO*) from West Greenland, plotted in relation to single-stage growth curve with $\mu_1 = 7.5$. Range of Amîtsoq gneiss isotopic compositions at 2900 Ma is between vertical bars on line *AN*. Average Amîtsoq gneiss isotopic composition at 2900 Ma ago is at point *M*. Line *MB* shows approximately where late Archaean gneisses would now lie if they had been produced by partial melting of Amîtsoq gneisses at 2900 Ma ago (Black et al., 1971; Moorbath et al., 1975b; Taylor et al., 1980; Oxford, unpubl. data).

geochemical differentiation over a time span not exceeding c. 100—200 Ma. Typically, a large proportion of granulite and upper amphibolite facies gneisses plot very close to the lower intercept of their isochron with the appropriate single-stage mantle growth curve (Fig. 20-2), demonstrating the characteristic phenomenon of U-depletion during high-grade metamorphism immediately following crystallization of the igneous protoliths.

In contrast to the situation for Rb and Sr discussed earlier, internal geochemical differentiation of sialic crust produces lower continental crust in which U/Pb ratios are *much lower* than in the upper mantle, so that radiogenic Pb isotopic evolution in the lower continental crust is severely retarded relative to the mantle. Pb isotopic analyses, in conjunction with U and Pb concentration data, show that U/Pb ratios in deep continental crust are typically 5—10 times lower than in the upper mantle. This very fundamental difference between lower continental crust and upper mantle has clearly persisted throughout the entire history of a given sector of continental crust. This makes it possible to use Pb isotopes to distinguish unambiguously between accretion of mantle-derived juvenile sial and reworking of much older sialic crust.

Severe depletion of U relative to Pb is by no means confined to pyroxene-granulites, but extends to many amphibolite-facies gneisses. U appears to be more mobile than Rb and Th in dehydrating crust, such that U is even more easily separated from Pb than Rb is from Sr. As an example, late Archaean (c. 2700 Ma) pyroxene-granulites of northwest Scotland, as well as the early Archaean (c. 3700 Ma) amphibolite facies Amîtsoq gneisses of West Greenland, both have extremely unradiogenic Pb isotopic compositions which plot close to the intersection of their respective isochrons with the single-stage mantle growth curve for Pb isotopic evolution, indicating that the crustal development of the Pb in these orthogneiss units began at or only shortly before c. 2700 Ma and 3700 Ma ago, respectively (Moorbath et al., 1969, 1975a; Black et al., 1971; Chapman and Moorbath, 1977; Gancarz and Wasserburg, 1977). If the Lewisian or Amîtsoq gneisses were partially melted *today*, the Pb isotopic ratios of the resultant granitoid magmas would both be far less radiogenic than that in any modern mantle-derived igneous rock. In contrast, a modern Lewisian-granulite-derived magma would have a low mantle-type $(^{87}Sr/^{86}Sr)_i$ of c. 0.703, whilst a modern Amîtsoq-gneiss-derived magma would have a very high $(^{87}Sr/^{86}Sr)_i$ in the range c. 0.72—0.74.

The same principle can be applied to determine the isotopic and genetic relationship of the late Archaean (c. 2900 Ma) gneisses of West Greenland outside the Godthaabsfjord region (Fig. 20-3, Nordland, Sukkertoppen, Fiskenaesset areas) to early Archaean crust (c. 3700 Ma) of Amîtsoq gneiss type. (Note that the so-called Nûk gneisses proper of the Godthaabsfjord region show a different type of Pb isotopic behaviour, to be described later.) Sr isotope data show that these gneisses are not reworked high-Rb/Sr Amîtsoq gneisses of the type exposed in the Godthaabsfjord region (see above; also Moorbath and Pankhurst, 1976). However, Sr isotope data cannot exclude the possibility that these late Archaean gneisses were produced by partial melting of deep-seated, low-Rb/Sr, early Archaean gneisses, of which the Rb/Sr and $^{87}Sr/^{86}Sr$ ratios were not significantly different from contemporaneous mantle values. In contrast, the known Amîtsoq gneisses became so severely depleted in U relative to Pb c. 3700 Ma ago that their Pb isotopic evolution virtually ceased at that time, with most of their $^{206}Pb/^{204}Pb$ ratios ranging from c. 11.5 up to only 13.1 as measured today (Fig. 20-2). Amîtsoq gneisses at greater depth could hardly be less depleted in U than any now exposed. It follows that any partial melt of the Amîtsoq gneisses would contain unradiogenic Pb whose crustal development began at c. 3700 Ma ago. Pb-Pb whole-rock measurements on numerous samples of late Archaean orthogneisses outside the Godthaabsfjord region demonstrate that observed Pb isotope systematics developed in a crustal environment only since c. 2900 Ma ago, and not since 3700 Ma ago (Black et al., 1973; Taylor et al., 1980). This is evident from Fig. 20-2 in which the Amîtsoq and late Archaean gneiss isochrons respectively intersect the single-stage growth curve at 3700 Ma and 2900 Ma ago. The continuous

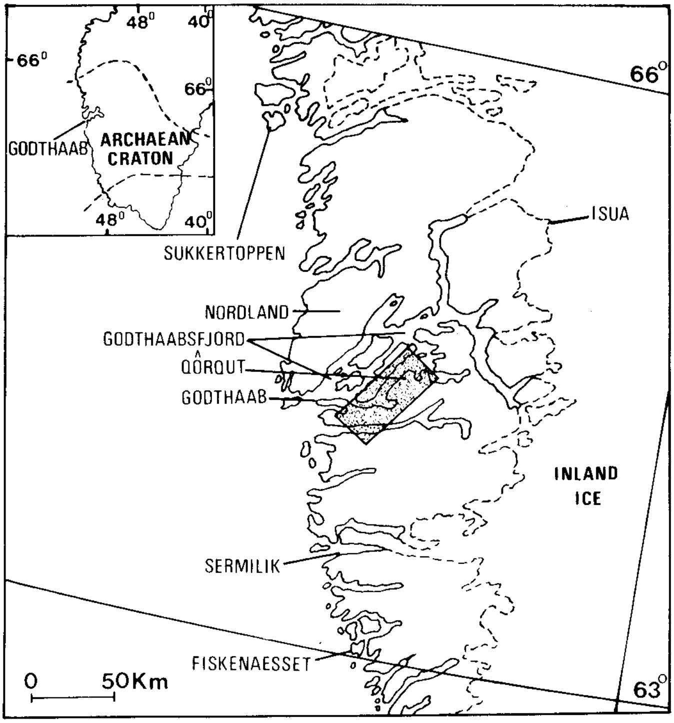

Fig. 20-3. Sketch map of southern West Greenland, showing localities referred to in the text. Most of the area covered by stippled rectangle is occupied by the Qôrqut granite.

part of line *AN* between the vertical bars shows the calculated range of Pb isotope compositions for the known Amîtsoq gneisses at 2900 Ma ago. The mean Pb isotopic composition of Amîtsoq gneisses 2900 Ma ago is at point *M*. The dashed line *MB* shows approximately where late Archaean gneisses would now lie if they had been produced by partial melting of the Amîtsoq gneisses at 2900 Ma ago.

Similarly, Pb isotope studies on late Archaean gneisses of northwest Scotland rule out any significant contribution from early Archaean Amîtsoq-type continental crust (Chapman and Moorbath, 1977), whilst the same holds for other gneisses from the North Atlantic craton (see Table 20-I). Likewise, there are no Pb isotopic indications that the 3700 Ma old Amîtsoq gneisses have continental sialic precursors of significantly greater age. Their initial Pb isotopic compositions provide very close temporal constraints indicating that the development of their Pb in the crust commenced not earlier than c. 3700 Ma ago. Thus we do not consider that the Amîtsoq gneisses can be derived by the partial melting or reworking of continental, sialic crust older than the c. 3800 Ma old Isua supracrustal sequence (e.g. Black et al., 1971; Baadsgaard et al., 1976; Gancarz and Wasserburg, 1977; Appel et al., 1978).

The principal diagnostic isotopic features for the recognition of continental growth in a CADS are: (a) age concordance between Rb-Sr and Pb-Pb whole-rock isochrons; (b) mantle-type initial Sr ratios; (c) primary μ_1 ($^{238}U/^{204}Pb$) values clustering closely around 7.5—8.0 for the source region; and (d) unradiogenic Pb isotopic compositions in U-depleted high-grade gneisses, which indicate an initial Pb isotopic composition close to the appropriate single-stage growth curve at the time of the CADS.

It is unfortunate that direct determination of initial Pb isotopic ratios from U-Pb whole-rock isochrons (i.e. plots of $^{238}U/^{204}Pb$ versus $^{206}Pb/^{204}Pb$, and $^{235}U/^{204}Pb$ versus $^{207}Pb/^{204}Pb$) does not appear to be of great promise. In principle, this would be better than deriving initial Pb isotopic ratios from the intersection of Pb-Pb isochrons with single-stage growth curves. However, there is much evidence that U can be lost easily from rocks in the zone of meteoric alteration. This effect has been reported by several workers (e.g. Black and Richards, 1972; Rosholt et al., 1973; Moorbath et al., 1975b). Pb-Pb systematics are unaffected within the precision of measurement since loss of U occurred geologically "recently". In the case of the Amîtsoq gneisses (Moorbath et al., 1975b) this leads to the remarkable conclusion that they suffered deep-seated, severe U depletion some 3700 Ma ago during the CADS which produced them, and then again in geologically "recent times" by a near-surface process, but *not* in between.

Thus, we conclude that, by analogy with the Rb-Sr method, Pb-Pb whole-rock isochrons can give the time of metamorphic and geochemical differentiation of a gneiss complex. The early and late Archaean gneisses of West Greenland, as well as many other calc-alkaline orthogneiss terrains from all continents, represent individual CADS of geochronologically short duration (Moorbath, 1977, 1978a). In many cases, the duration of a CADS may not exceed the measured uncertainties (c. ± 50 to 100 Ma) on the corresponding Pb-Pb whole-rock isochron age measurements.

Recognition of reworking of older continental crust

As a contrast to a CADS, we now consider a case of crustal reworking, namely, the Qôrqut granite, which represents the last major rock-forming event in the Archaean of the Godthaabsfjord region of West Greenland, and occupies an area of some 1000 km^2 (Bridgwater et al., 1976). A published Rb-Sr whole-rock isochron gave an age of 2460 ± 90 Ma, with corresponding $(^{87}Sr/^{86}Sr)_i$ of 0.709 ± 0.007, (Moorbath and Pankhurst, 1976). Additional work by Moorbath et al. (in press) has given a 23-point perfect whole-rock isochron with values of 2530 ± 30 Ma and 0.7083 ± 0.0004. This high $(^{87}Sr/^{86}Sr)_i$ clearly indicates a degree of crustal involvement (see Fig. 20-1a).

A 24-point Pb/Pb whole-rock isochron on the same samples yields an age of 2580 ± 80 Ma (Fig. 20-4). The isochron line lies far below a reference

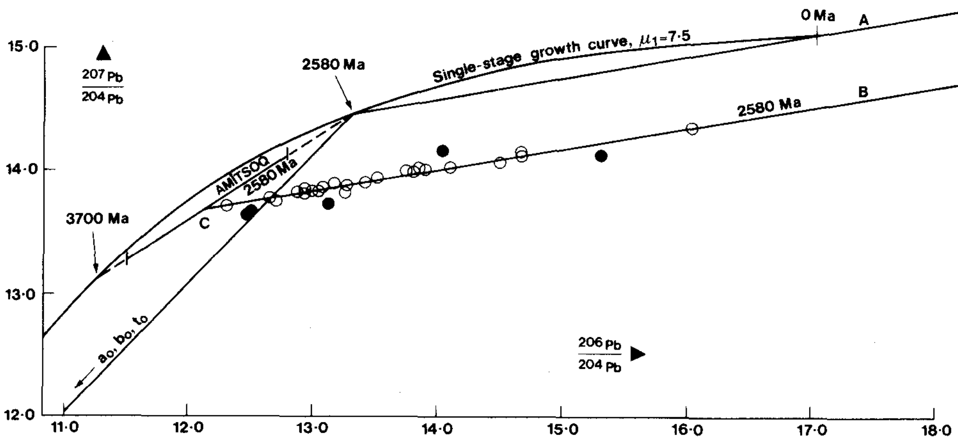

Fig. 20-4. Pb-Pb diagram for data from the Qôrqut granite of West Greenland in relation to a single-stage growth curve with $\mu_1 = 7.5$. The 2580 Ma line *BC* has been regressed through the 24 open-circle data points only. That part of the 3700—2580 Ma line between vertical bars represents the calculated range of Pb isotopic compositions of Amîtsoq gneisses at 2580 Ma ago. Line *A* shows where Qôrqut data points would lie if the granite had been produced from a mantle-like source region at 2580 Ma ago, similar to that which produced the Nûk gneisses at c. 2900 Ma ago as shown in Fig. 20-2 (Moorbath et al., in press).

isochron for rocks derived from an upper mantle source with a μ_1 value of 7.5 (characteristic of the source from which the Amîtsoq and Nûk gneisses were derived). Furthermore, several Qôrqut granite whole-rock and pegmatite feldspar Pb isotope data plot on the c. 2580 Ma isochron to the left of the first-stage isochron for 2580 Ma, a clear demonstration that isotopic evolution of Qôrqut granite Pb cannot be described by a simple two-stage model. On the (erroneous) assumption of a two-stage model the μ_1 value calculated for the source region would be 6.23, much lower than the range from 7.5 to 8.0 which typifies the source regions of many Archaean orthogneisses (Table 20-I).

The Qôrqut granite clearly had a very unradiogenic initial Pb isotopic composition, probably close to the least radiogenic Pb isotopic composition measured in a Qôrqut pegmatite feldspar, and only slightly more radiogenic than the average Pb isotopic composition of Amîtsoq gneisses in the Godthaabsfjord area at c. 2580 Ma. No other local source of very unradiogenic Pb is known than the Amîtsoq gneisses, and it is concluded that the Qôrqut granite magma was mainly derived by melting of Amîtsoq gneiss material, which is the predominant component of the country rock to the Qôrqut granite. However, more detailed consideration of initial Sr and Pb isotopic compositions of the Qôrqut granite (Moorbath et al., in press) indicates that a significant proportion of the Qôrqut granite magma was probably derived from younger source material, most probably the Nûk gneisses, although an upper mantle-derived contribution cannot be formally excluded.

Even if there were no other Pb-Pb age and isotope information from the Godthaabsfjord region, the anomalously low apparent μ_1 value of 6.23, in combination with the high $(^{87}Sr/^{86}Sr)_i$, would lead one to postulate partial melting of ancient U-depleted, sialic crust for production of the Qôrqut granite. We are currently studying several Proterozoic granites from East Greenland and Northwest Scotland with analogous Pb isotope systematics.

It will be obvious that partial melting of significantly older continental sialic crust can give rise to a wide range of permutations of initial Pb and Sr isotope characteristics in the resulting granitic magmas, depending upon the geochemical characteristics of the crustal level at which melting occurs. Deep-seated, low-Rb/Sr, low-U/Pb, crust will produce mantle-type $(^{87}Sr/^{86}Sr)_i$ and unradiogenic Pb; intermediate-level, high-Rb/Sr, low-U/Pb crust, broadly corresponding to amphibolite facies rocks will produce high $(^{87}Sr/^{86}Sr)_i$ and unradiogenic Pb; high-level, high-Rb/Sr, high-U/Pb crust will produce high $(^{87}Sr/^{86}Sr)_i$ and radiogenic Pb. These are, of course, only three arbitrarily chosen points in what amounts to a continuous spectrum of possibilities, but all of them are different to the combined isotope characteristics which would be produced from a mantle-type source region. However, some of these isotopic characteristics could also be produced in magmas of any kind undergoing selective contamination by surrounding continental crust, although diverse petrological, geochemical and other isotopic criteria can usually be applied for the recognition of this phenomenon (e.g. Carter et al., 1978b; Briqueu and Lancelot, 1979; Moorbath and Thompson, 1980).

The initial Pb and Sr isotopic compositions of a crust-derived granitic rock approach mantle-type values with diminishing age differences between production of the crustal source material for the granite from the mantle and the crustal reworking event in which the granite is formed. It is also obvious that basic crust with geochemical characteristics similar to upper mantle cannot evolve any significant isotopic contrast with the upper mantle. These inherent limitations of the isotopic approach must always be borne in mind and each individual case should be carefully interpreted on its own merits.

Recognition of ancient continental basement at depth

Earlier it was shown that Pb-Pb whole-rock systematics of late Archaean (c. 2900 Ma) gneisses of West Greenland, outside the Godthaabsfjord region, are straightforward and indicate derivation of their magmatic parents from a source region with typical mantle $^{238}U/^{204}Pb$ ratio ($\mu_1 = 7.5$) at, or shortly before c. 2900 Ma ago (see also Black et al., 1973). However, in the Godthaabsfjord region itself (Fig. 20-3), where late Archaean Nûk gneisses and

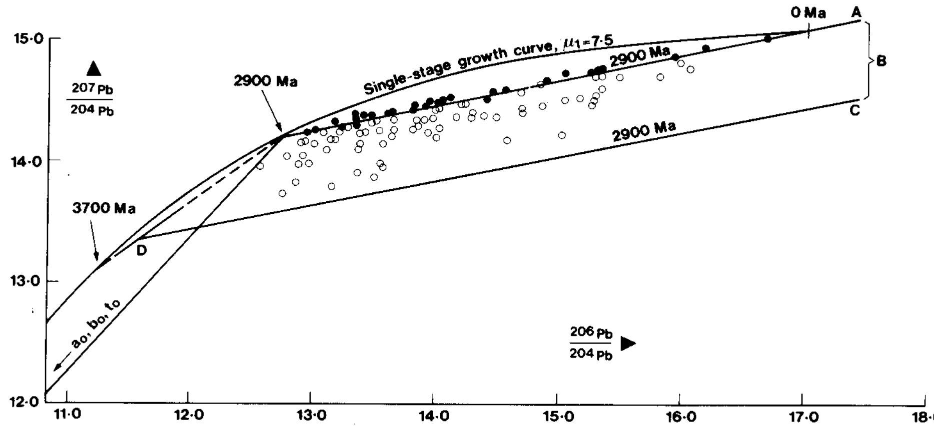

Fig. 20-5. Pb-Pb diagram for data from Nûk gneisses and associated anorthosites from the Godthaabsfjord region of West Greenland which have been contaminated with Amîtsoq-type Pb (open circles, field *B*), as explained in the text. Line *A* is regressed through uncontaminated equivalents to the Nûk gneisses from Sukkertoppen, Nordland and Fiskenaesset areas (Figs. 20-2 and 3). Line *DC* shows where late Archaean gneisses would now lie if they had been produced by partial melting of Amîtsoq gneisses at 2900 Ma ago (Taylor et al., 1980).

early Archaean Amîtsoq gneisses are both exposed (McGregor, 1973, 1979; Bridgwater et al., 1976), most Nûk gneisses contain variable proportions of two isotopically distinct types of Pb which started their respective crustal developments at c. 2900 and 3700 Ma ago. Both Pb and Sr isotopic constraints clearly demonstrate that Nûk gneisses are not derived by reworking or remelting of Amîtsoq gneisses. Instead they show that the mixing of early and late Archaean Pb results from selective contamination, involving extraction of very unradiogenic Pb derived from c. 3700 Ma old Amîtsoq-type continental crust invaded by Nûk magmas, and addition of this contaminant to Nûk magmas (Taylor et al., 1980).

This is clearly illustrated in Fig. 20-5, in which Nûk gneisses and several associated anorthosite bodies from the Godthaabsfjord region all scatter well below the line for contemporaneous late Archaean gneisses from outside the Godthaabsfjord region, but well above a hypothetical isochron for rocks derived by Amîtsoq gneisses. Each individual mixed-Pb point starts its radiogenic development at c. 2900 Ma ago somewhere along a mixing line joining the points for 3700 and 2900 Ma on the single-stage growth curve. Each individual radiogenic development line to the present-day clearly has a slope of c. 2900 Ma. Several relatively small, petrologically homogeneous rock suites nonetheless yield reasonable isochron fits, indicating that the initial mixed-Pb in the respective individual Nûk magma batches was more or less homogeneous (not shown in Fig. 20-5). There is furthermore a relationship between degree of isotopic contamination of a given sample (i.e. distance below the upper boundary line for uncontaminated gneisses) and Pb concentration. The isotopically most contaminated samples generally have the highest Pb contents, as well as certain other incompatible elements including K, Rb and U.

Variations in the extent of Pb isotopic contamination exist on a scale of tens to hundreds of metres, and probably within individual intrusive units. This suggests that the contamination process occurs close to the site of emplacement, and certainly rules out any deep-seated (mantle) heterogeneity as the primary source of initial Pb isotopic heterogeneity of Nûk magmas.

Using Pb isotopes in the manner outlined above, it has not been possible to detect any early Archaean Amîtsoq-type continental crust in areas more than 10—15 km outside the known outcrop of the Amîtsoq gneisses within the Godthaabsfjord region. We conclude that such crust does not exist at depth elsewhere in southern West Greenland. Pb isotope work on many late Archaean gneisses from elsewhere in the North Atlantic craton, for example Northwest Scotland (Chapman and Moorbath, 1977), East Greenland (Taylor et al., unpubl. data), North Norway (Griffin et al., 1978; Jacobsen and Wasserburg, 1978), similarly fails to provide evidence for subjacent early Archaean continental basement. All these gneisses (except the Vikan gneisses of North Norway, see next section) have simple Pb-Pb whole-rock systematics, analogous to the late Archaean gneisses of West Greenland outside the Godthaabsfjord region (Table 20-I). Furthermore, there is no "contamination-type" Pb isotopic evidence, such as that above, to suggest that the 3700 Ma old Amîtsoq gneisses were themselves emplaced through or into any continental, sialic basement of significantly greater age. The rapid evolution of radiogenic Pb in the upper mantle during the early Archaean would render the detection of contaminated, mixed leads particularly straightforward.

The approach outlined in this section should prove to be a very powerful and widely applicable tool for the sub-surface "mapping" of older continental sialic crust and thus for the study of the growth of continental crust with time.

Recognition of effects of high-grade metamorphism long after primary igneous crystallization

The Pb-Pb whole-rock isochron method can yield spurious age results when a high-grade metamorphism affects a rock unit a very long time after its primary formation, that is to say when crustal accretion and final metamorphic/geochemical differentiation do *not* form part of the same CADS. For example, Taylor (1975) published a Pb-Pb whole-rock isochron age of 3410 ± 70 Ma for the migmatitic Vikan gneisses of Langøy, Vesteralen, North Norway, and originally concluded that this age referred to the gneiss-forming event. A Rb-Sr whole-rock isochron on the same rock samples yielded an age of 2300 ± 150 Ma with an $(^{87}Sr/^{86}Sr)_i$ of 0.7126 ± 0.0011. Some anomalous features of the Pb isotope systematics and the discordance between Pb-Pb and Rb-Sr ages, as well as further Pb, Sr and Nd isotopic measurements (Griffin et al., 1978; Jacobsen and Wasserburg, 1978; Taylor et al., unpubl. data) have led to a drastic re-interpretation. The high $^{238}U/^{204}Pb$ (μ_1) value of 8.9 implied for the source region of the Vikan gneisses on the basis of a single-stage model (compared with typical values of c. 7.5—8.0), as well as the lack of any particularly unradiogenic Pb isotopic compositions, are highly unusual by comparison with other Precambrian high-grade gneiss terrains for which Pb-Pb data are currently available. It is now clear that the present-day range of Pb isotopic compositions cannot have resulted from a single stage of Pb isotopic evolution from a uniform unitial Pb isotopic composition. It has been convincingly shown that the linear array of Pb-Pb whole-rock data reported by Taylor (1975) resulted from the formation of the protolith of the Vikan gneisses from an upper-mantle source region at c.2600—2700 Ma ago, followed by a severe disturbance of whole-rock U-Pb and Rb-Sr systems (specifically U and Rb depletion) during granulite facies metamorphism c. 1800 Ma ago.

The linear Pb-Pb array for the Vikan gneisses is thus no longer regarded as a true secondary isochron, but as a "transposed palaeo-isochron" (Griffin et al., 1978) in which the present-day Pb isotope data fall on a slightly transposed line from that joining the points corresponding to 2680 Ma and 1760 Ma on a single-stage mantle growth curve (i.e. the isochron of c. 2680 Ma rocks at c. 1760 Ma). This is illustrated in Fig. 20-6. Clearly, interpretation of the line joining points 2680 Ma and 1760 Ma on the single-stage growth curve will give an anomalously high age if a simple two-stage model is assumed. The transposition of the "palaeo-isochron" developed between c. 2680 Ma and 1760 Ma results from radiogenic development of Pb in a low U-Pb environment from c. 1760 Ma to the present.

It seems likely that more cases of this type will turn up and form an important advance in understanding long-term crustal evolution. Combined Pb, Sr and Nd isotope systematics in such a case are very different indeed from those observed during crustal accretion and penecontemporaneous metamorphic/geochemical differentiation, not least by the apparent discordance between different isotopic methods. Provided, however, that several different isotopic methods are used, the interpretations should be unambiguous. The Vikan gneisses provide a clear example of geochemical reactivation of much older continental crust and, as such, are isotopically as clearly recognizable as the contrasted case of crustal accretion and differentiation during a single, major episode of continental growth. By analogy, we consider that the combined isotopic approach is capable, in principle, of resolving the geochemical and geochronological complexities of gneisses and other rock units which have undergone metasomatism and/or retrogression

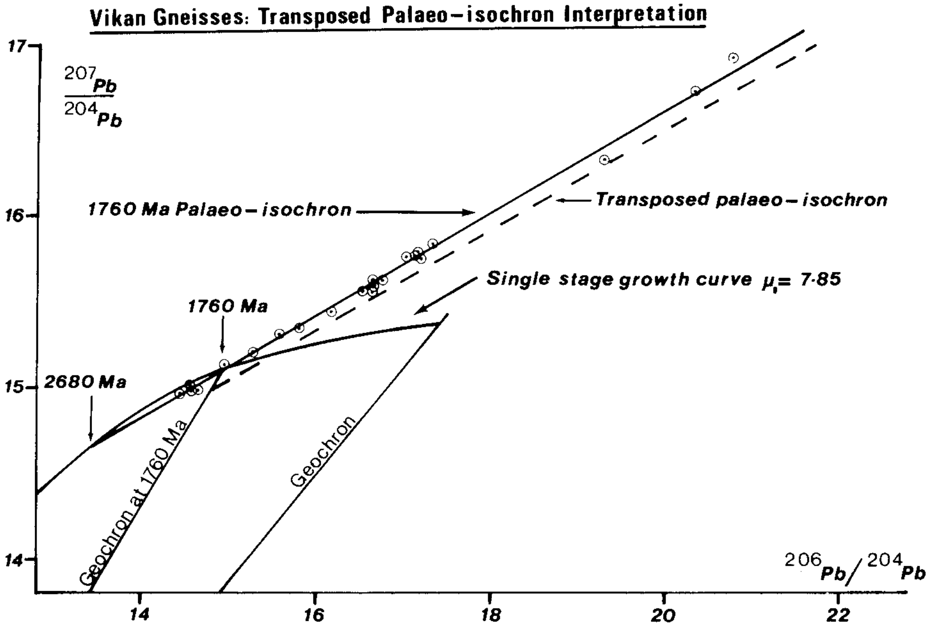

Fig. 20-6. Pb isotopic evolution in the Vikan migmatic gneisses, North Norway. The plotted data points have been corrected to 1760 Ma ago and thus define the 1760 Ma palaeo-isochron, the slope of which corresponds to an age of 2680 Ma. Assuming two stages for Pb isotopic evolution up to 1760 Ma, a μ_1 value of 7.85 is calculated for the first stage. The broken line represents the lower array of present-day isotopic compositions of the Vikan gneisses (Taylor, 1975). The slope of this line is indistinguishable from the slope of the 1760 Ma palaeo-isochron. The broken line is thus interpreted as a "transposed palaeo-isochron", resulting from small, fairly uniform increments in $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ ratios in all samples since the establishment of very low and fairly uniform $^{238}Pb/^{204}Pb$ ratios by severe U depletion during high-grade metamorphism at 1760 Ma ago (Griffin et al., 1978; Jacobsen and Wasserburg, 1978; Taylor, unpub. data).

at some time long after their primary formation. Little isotopic work has yet been reported for those parts of gneiss terrains in which these processes are clearly recognizable. However, the effects of metasomatism and retrogression on radiogenic isotope systematics can be predicted to be very different indeed from those postulated by Collerson and Fryer (1978) in their unconventional model of massive mantle—crust metasomatism and crustal transformation.

Nd ISOTOPES

In the following account all Nd isotopic ratios are normalized to equivalence with $^{150}Nd/^{142}Nd = 0.2096$.

Considerable pioneering advances have been made in recent years in the application of the $^{147}Sm/^{143}Nd$ method to the study of continental evolution

(De Paolo and Wasserburg, 1976a, 1976b; see also review by O'Nions et al., 1979). The published results are in good accord with conclusions regarding crustal evolution from Sr and Pb isotope systematics, but also provide additional important constraints on the age and origin of major crustal segments. The principal contrast between the Sm-Nd method and both the Rb-Sr and U-Pb methods lies in the comparative geochemical coherence of Sm and Nd. Unlike Rb and Sr, and U and Pb, these two rare earth elements (REE) are not fractionated on a large scale by crustal processes. This has been vividly demonstrated by McCulloch and Wasserburg (1978) who obtained Sm-Nd model ages of gneiss composites from the Canadian Shield, representing portions of the Superior, Slave and Churchill structural provinces, which indicated that these provinces were formed within the period c. 2700—2500 Ma, in general agreement with much published Rb-Sr work. In addition these workers determined the age of sediment provenances from many other areas, since it appears that the Sm-Nd isotopic system is not significantly disturbed during sedimentary or diagenetic processes.

However, Nd is enriched relative to Sm during magmatic processes leading to production of sialic crust from the upper mantle. The continental crust is a light-REE enriched reservoir compared to the upper mantle. Hamilton et al. (1979a) have suggested that mantle—crust differentiation may be precisely dated by the Sm-Nd method. They report a Sm-Nd whole-rock isochron age for Lewisian granulite gneisses of Northwest Scotland of 2920 ± 50 Ma with $(^{143}Nd/^{144}Nd)_i$ of 0.508959 ± 49. This age is significantly greater than reliable Lewisian Rb-Sr and Pb-Pb whole-rock ages, as well as U-Pb zircon ages, which all group fairly closely round c. 2650—2750 Ma. Initial Sr and Pb isotopic compositions of Lewisian gneisses have been interpreted as permitting a maximum time interval of c. 100—200 Ma between formation of the magmatic precursors of the gneisses and their metamorphic/geochemical differentiation (Moorbath et al., 1969, 1975a; Chapman and Moorbath, 1977). The Sm-Nd age of 2920 ± 50 Ma is interpreted by Hamilton et al. (1979a) as the time of mantle—crust differentiation to produce the calc-alkaline magmatic precursors of the Lewisian gneisses. The time interval between the Sm-Nd age and the Pb-Pb whole-rock age is 240 ± 90 Ma, interpreted as the time interval between crustal accretion and metamorphic/geochemical differentiation, when whole-rock samples became closed systems to Rb, Sr, U, Pb migration. The $(^{143}Nd/^{144}Nd)_i$ for the Lewisian magmatic precursors lies on the (chondritic) growth curve for the mantle (see below, and Fig. 20-7).

Since mantle and continental crust evolve with relatively high and low Sm/Nd, respectively, it follows that $(^{143}Nd/^{144}Nd)i$ ratios can provide a useful criterion for characterization of the source region of any given igneous rock, in analogy with the other decay schemes. Thus the rate of growth of $^{143}Nd/^{144}Nd$ in sialic crust is retarded relative to the mantle. No Sm-Nd work has yet been published on granites which might be expected on other

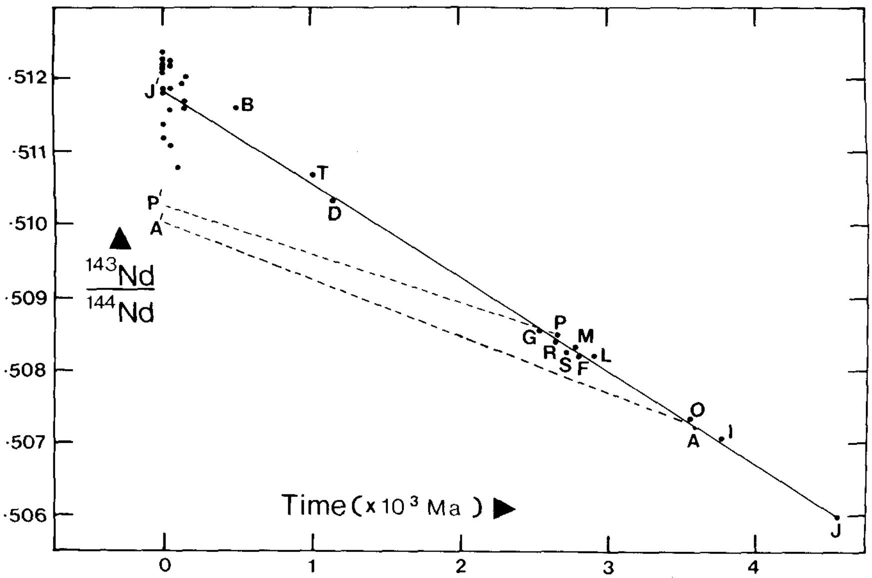

Fig. 20-7. Initial $^{143}Nd/^{144}Nd$ ratios of igneous rocks (including orthogneisses) plotted against age. Data for < 200 Ma old volcanic rocks (not coded in diagram) are from De Paolo and Wasserburg (1976a, b, 1977). B = Bay of Islands ophiolite, Newfoundland (Jacobsen and Wasserburg, 1979); T = Town Mountain granite, Llano, Texas (De Paolo and Wasserburg, 1976b); D = Duluth gabbro, Minnesota (De Paolo and Wasserburg, 1976b); G = Great Dyke of Zimbabwe (De Paolo and Wasserburg, 1976a); P = Preissac-Lacorne Batholith, Superior Province, Canada (De Paolo and Wasserburg, 1976a); R = Zimbabwean greenstone belts (Hamilton et al., 1977) and Louis Lake granodiorite, Wyoming (De Paolo and Wasserburg, 1976b); S = Stillwater Complex, Montana (De Paolo and Wasserburg, 1979); M = Munro Township tholeiites and komatiites, Abitibi belt, Canada (Zindler et al., 1978); F = Fiskenaesset anorthosite, West Greenland (De Paolo and Wasserburg, 1976b); L = Lewisian gneisses, northwest Scotland (Hamilton et al., 1979a); O = Onverwacht volcanics, southern Africa (Hamilton et al., 1979b); A = Amîtsoq gneisses, West Greenland (De Paolo and Wasserburg, 1976a); I = Isua supracrustals, West Greenland, (Hamilton et al., 1978).

J-J' is the Nd isotopic evolution line for Juvinas achondrite, with Sm/Nd = 0.308, very close to the chondritic average. A-A' and P-P' are Nd isotopic evolution lines for the Amîtsoq gneiss and for the Preissac-Lacorne Batholith, characterizing the retarded Nd isotopic evolution of continental crust. The Stillwater Complex is the only Archaean rock unit so far investigated with initial $^{143}Nd/^{144}Nd$ deviating significantly from a chondritic mantle evolution line, possibly as a result of crustal contamination. All data recalculated to conform to mass fractionation correction procedure of De Paolo and Wasserburg (1976a, b). Decay constant $^{147}Sm = 6.54 \times 10^{-12} a^{-1}$.

petrological, geochemical and isotopic grounds to be products of anatexis of ancient crust. Significant departures of initial $^{143}Nd/^{144}Nd$ ratios from mantle-type (chondritic) Nd isotopic evolution would be expected for such cases. In addition, it appears that magmatic rocks in a crustal environment may also be subject to Nd isotopic perturbations resulting from the type of magma—crust interaction which leads to selective contamination and isotopic mixing, as discussed earlier for other isotopic systems (e.g. Carter et al., 1978b).

The comparatively close geochemical coherence between Sm and Nd makes it possible to date reliably amphibolite sequences from Archaean greenstone belt supracrustals which frequently exhibit partial open-system behaviour to Rb and/or Sr (Hamilton et al., 1977, 1978, 1979b). Almost all $(^{143}Nd/^{144}Nd)_i$ ratios so far obtained from Sm-Nd isochrons or model calculations (De Paolo and Wasserburg, 1976a; O'Nions et al., 1979) on Archaean rocks lie on the chondritic growth curve for the bulk earth-mantle reservoir (Sm/Nd = 0.308, equivalent to $^{147}Sm/^{144}Nd = 0.1936$, Fig. 20-7). Each analyzed rock unit in Fig. 20-7 yields a Sm-Nd age for its respective mantle—crust differentiation episode, together with corresponding $(^{143}Nd/^{144}Nd)_i$ of the mantle source region at that time. Any significant crustal pre-history is precluded for these rock units.

Progressive development of continental crust must have increased the Sm/Nd ratio of the residual mantle, so that the latter should have a $^{143}Nd/^{144}Nd$ greater than the bulk earth value (estimated at 0.51262, see O'Nions et al., 1979). The actual range of measured $^{143}Nd/^{144}Nd$ ratios in modern oceanic basalts from c. 0.5125 to c. 0.5133 indicates that different segments of mantle have at different times suffered varying degrees of REE fractionation as a result of magma generation. In this connection, O'Nions et al. (1979) state that "the apparent uniformity of Sm/Nd in the mantle reservoir from which continental crust was extracted during the Archaean is perhaps surprising in view of the fact that the reservoir must change in response to continental growth. This apparent uniformity may reflect the combined effects of large reservoir size (e.g. whole mantle) and uniform extraction of REE in the Archaean such that the change in Sm/Nd in the whole mantle was insignificant. Alternatively the uniformity may reflect isolation of mantle source regions such that they escaped fractionation events. Differentiating between these two possibilities prevents an intriguing problem for the future".

It should be noted from Fig. 20-7 that the initial $^{143}Nd/^{144}Nd$ ratios of several post-Archaean igneous rocks also fall on the mantle evolution line, although an Archaean exception (the Stillwater Complex of Montana) has recently been reported (De Paolo and Wasserburg, 1979).

Nd isotopic data strengthen the isotopic case against models for terrestrial evolution which include creation of most of the continental crust very early in the history of the earth. Extraction of LREE-enriched continental crust from a bulk earth with a chondritic REE pattern necessarily leaves a LREE-depleted mantle residuum. If extraction of the continental crust occurred very early in the earth's history, Nd isotopic evolution in LREE-depleted mantle should have accelerated and diverged from the chondritic growth line from early Archaean times onwards (LREE depletion corresponds to increase of Sm/Nd ratio to a value greater than chondritic 0.308). In fact, departures of mantle source regions from chondritic Nd isotopic evolution are very small until at least well into Proterozoic times. This observation provides a strong case against early formation of most of the present mass of the continental crust.

Jacobsen and Wasserburg (1979) have deduced a mean age for continental crust—mantle separation of c. 1800 Ma from a detailed consideration of the divergence of Nd isotopic evolution of the upper mantle from a chondritic course during Phanerozoic times. In a uniformitarian model with more or less continuous crustal accretion the mean age of c. 1800 Ma would actually imply commencement of crustal accretion at c. 3600 Ma, in reasonable agreement with the ages of the oldest known segments of sialic crust in West Greenland, Labrador and Southern Africa.

RADIOGENIC ISOTOPES AND MANTLE EVOLUTION

This important and extensively documented topic almost certainly has an important bearing on evolution of continental crust through geological times.

It is well known that the immediate upper mantle source regions of oceanic basalt are heterogeneous with respect to $^{87}Sr/^{86}Sr$, $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, $^{208}Pb/^{204}Pb$ and $^{143}Nd/^{144}Nd$ ratios, and that this heterogeneity has developed over a period of time much greater than the development of present-day ocean basins. Indeed, the age calculated from the slope of Pb/Pb and Rb/Sr regression lines (often termed "mantle isochrons") for oceanic basalts suggests that this isotopic heterogeneity has existed and developed for at least c. 2000 Ma (see, for example, Sun and Hanson, 1975; Brooks et al., 1976b). Opinions differ whether this isotopic heterogeneity has developed continuously or episodically, and this depends largely on the interpretation of the Pb-Pb and Rb-Sr regression lines. In any case, the isotopic heterogeneities clearly demonstrate the existence of upper mantle heterogeneity with respect to Rb/Sr, U/Pb and Sm/Nd ratios, whilst there is much evidence that the source materials of oceanic basalts derived from progressively deeper levels in the mantle approach bulk earth (undepleted chondritic) elemental and associated isotopic ratios (corresponding to increase in Rb/Sr, and decrease in Sm/Nd). Recent combined Sr and Nd isotopic work demonstrates that most basalts erupted in the oceanic basins, including those in Iceland and Hawaii, are actually derived from source regions which have been depleted to varying extents in Rb and light REE (i.e. Nd relative to Sm). A few oceanic basalts, such as those from Tristan da Cunha or the Azores (Hawkesworth et al., 1979b), which have isotopic compositions similar to chondritic or calculated bulk earth values ($^{143}Nd/^{144}Nd = 0.51264$, $^{87}Sr/^{86}Sr = 0.705$, $Rb/Sr = 0.032$, see O'Nions et al., 1977, 1978), are volumetrically trivial in the ocean basins. O'Nions et al. (1978) infer that all recent basalts with $^{87}Sr/^{86}Sr < 0.705$ have been generated from source regions depleted relative to the bulk earth. The same surely holds for many mantle-derived continental igneous rocks. Indeed, it is likely that many continental igneous rock suites with $^{87}Sr/^{86}Sr >$ c. 0.705 have probably interacted with crust with concomitant enhancement in $^{87}Sr/^{86}Sr$ by selective contamination, with or without accompanying crustal assimilation or partial melting. We do not in general favour a mantle origin for such high $(^{87}Sr/^{86}Sr)_i$ values in continental igneous rocks. There is as yet little evidence for the existence of sub-continental mantle with enriched bulk-earth Sr isotope ratios (but see Carter et al., 1978a; Hawkesworth and Vollmer, 1978).

Sr and Nd isotopic and other geochemical data show that the upper-mantle sources of oceanic and continental igneous rocks have become progressively depleted in lithophile elements Rb, Sr, K, U, Th, Sm and Nd through much of geological time. Associated with these depletions, Rb/Sr has decreased and Sm/Nd has increased in upper-mantle magma sources relative to bulk earth. It thus seems reasonable that upper-mantle depletion and creation of (enriched) continental, sialic crust are complementary processes. Clearly, one very effective, well-tried method of producing continental crust from upper mantle is by plate-tectonic processes in island-arc and continental margin environments. The initial radiogenic isotope ratios of a mantle-derived, crustally uncontaminated, magmatic rock suite in such an environment will represent locally homogenized, regionally variable mixtures of isotopes derived from subducted, oceanic lithosphere (possibly contaminated with some seawater-derived Sr and arguably with minor amounts of subducted sediments) and from overlying upper-mantle wedge. This is quite adequate to account for small initial ratio variations between one region and another. Larger isotopic variations can then be superimposed by crustal contamination processes of one kind or another, wherever continental crust occurs.

Evidence for progressive geochemical depletion as well as increasing isotopic and geochemical heterogeneity of the source regions of mantle-derived oceanic and continental igneous rocks, together with the powerful isotopic evidence for continental growth obtained on crustal rocks themselves, argue strongly against the much-debated model of Armstrong (1968) and Armstrong and Hein (1973) in which continental crust is modelled as forever cycling through the mantle with accompanying geochemical and isotopic equilibration, so that the mantle source region of igneous rocks remains a permanently pristine, undepleted reservoir. In this "steady-state" model, all of the continental crust is regarded as having been produced early on in the earth's history, with only minimal subsequent change in total mass. In any uniformitarian model of continental evolution this would imply quantitative subduction of sediments, as well as highly efficient crust-mantle mixing. It is difficult to square this model with the great majority of geological, geophysical, geochemical and isotopic observations of the past ten years or so.

In contrast, we envisage that recycling of continental crust through the mantle has only occurred on a minor scale, if at all, ever since the initial formation of typical continental crust, probably at around c. 3700 Ma ago.

CONCLUSIONS

Geochronological and associated Sr, Pb and Nd isotopic constraints can be used reliably to distinguish between juvenile, mantle-derived sial and reworked continental crust. Both continental growth and reworking have occurred over much of geological time, usually in contrasted tectonic environments. Nowadays, continental growth predominates at continental margins where subducted oceanic lithosphere and the overlying mantle wedge are partially melted and differentiated to yield calc-alkaline magmas, whilst reworking of continental crust predominates in mobile belts in regions of continental collision, where magmatic rocks commonly exhibit a high degree of isotopic complexity. There is no compelling reason to suppose that this state of affairs has changed fundamentally since the time of formation of the earliest continental crust, although the proportion of growth-to-reworking may have decreased with time. Thus the continental growth rate has almost certainly decreased with time due to the decaying thermal output of the earth (Brown, 1977; Brown and Hennessy, 1978), whilst the frequency and scale of continental collision must have increased with time as more continental crust has been created. Both continental growth and some form of inter- or intracratonic mobility became fully established as major global tectonic processes by late Archaean times (c. 3000—2500 Ma ago) during which time perhaps as much as 50—60% of the total mass of existing continental crust came into being. Brown (1977) calculated a contemporary global growth rate of 0.5 km^3 per year, so that it would take c. 10,000 Ma to produce all 5×10^9 km^3 of continental crust which now exist, implying that the accretion rate must have been higher in the past than at present (see also Veizer and Jansen, 1979).

At present, the evolution of the earth's crust during the first 800 Ma of earth history is still largely conjectural. Mechanical and thermal turbulence resulting from impacting planetesimals and/or intense mantle-wide convection (Elsasser et al., 1979) may well have inhibited the production of continental sialic crust. By c. 3800 Ma ago, solid crust of probably simatic ("proto-oceanic"?) character was able to support sediments and lavas and a hydrosphere. The earliest known thick, stable, permanent, geochemically-metamorphically differentiated, calc-alkaline continental crust appeared c. 3700 Ma ago. During the period c. 3700—3500 Ma, many relatively small continental nuclei may have appeared for the first time over the surface of the earth. Due to high thermal output of the earth during its early history, convective mixing within its outer parts was probably much more efficient and "plates" may have been thinner, smaller and much more numerous than at present. Production of continental crust from the mantle before c. 3000–2800 Ma may have been less efficient than in later times because of higher rates of convective overturn. Nevertheless, the appearance of perhaps the first true continental crust at c. 3700 Ma ago probably marks the beginning of an ancestral form of modern plate tectonics.

There is little, if any, geological, geochemical, geochronological or isotopic evidence for early Precambrian globe-encircling granitic crust, for massive recycling of continental crust through the mantle throughout geological time, or for global mixing of continental crust and mantle on a vast scale (see, for example, Fyfe, 1976, 1978; Hargraves, 1976; Lowman, 1976; Shaw, 1976). Most modern evidence suggests that the history of the outermost parts of the earth is dominated by irreversible chemical differentiation of the mantle (and particularly the *upper* mantle) throughout geological time, and by the permanence of continental crust once it is formed. We cannot agree with Fyfe (1978) that "conservation of crustal mass is one of the basic tenets of plate tectonics". In principle, plate-tectonic processes can just as well lead to an increase in crustal mass (e.g. Ringwood, 1974).

How much continentally derived sediment is subducted and partially melted to form new igneous rocks is still much debated (e.g. Karig and Sharman, 1975; James, 1978; Molnar and Gray, 1979). Kay et al. (1978) suggest that Pb and Sr isotopic data from the Aleutian and Pribilof Islands of Alaska might be explained by a mixture of c. 2% of continentally derived sediment with melt derived from subducted oceanic crust and overlying mantle wedge. However, it is highly probable that significant amounts of melting of continentally derived sediments occur deep within sedimentary basins marginal to a continental region, to produce the so-called "S-type" granites of Australian workers (e.g. Chappel and White, 1974; Flood and Shaw, 1977). As noted earlier, true granitic rocks can be produced from a great variety of precursors in different types of tectonic environments. However, we regard the largely mantle-derived rocks of the predominantly intermediate calc-alkaline association of *any* geological age and crustal level as truly symptomatic of continental accretion and growth.

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Chapter 21

THE RARE EARTH ELEMENT EVIDENCE IN PRECAMBRIAN SEDIMENTARY ROCKS: IMPLICATIONS FOR CRUSTAL EVOLUTION

S. R. TAYLOR and S. M. McLENNAN

ABSTRACT

The chemical composition of Phanerozoic igneous rocks has been utilized in revealing plate-tectonic histories, but application of such procedures to Precambrian (particularly Archaean) rocks is equivocal. This is well demonstrated by the composition of the Archaean Marda Complex which mimics modern island-arc chemistry, but was clearly formed under a different tectonic environment.

Application of geochemistry to the evolution of the continental crust is more soundly based. The composition of the present-day upper continental crust can be characterized by the rare earth element (REE) patterns in sedimentary rocks, since these elements are not easily fractionated by most sedimentary or metamorphic processes. Sedimentary REE patterns indicate that the present upper continental crust approximates to granodiorite. Such a composition cannot be representative of the entire continental crust due to geochemical balance and heat flow arguments. Intracrustal melting of andesite (representing the total crust) at depths < 40 km would produce a granodioritic upper crust with a characteristic negative Eu anomaly, and a residual lower crust with a mild Eu enrichment. The meagre lower crustal nodule data are in agreement with such a model. The lack of significant negative Eu anomalies in mantle-derived rocks limits sediment recycling via subduction to trivial amounts ($< 10\%$).

Archaean sedimentary rocks have differing REE patterns, resembling island-arc volcanic rocks, and indicate a much less evolved upper crust at that time. The change in REE patterns, as suggested by the Huronian and Pine Creek Geosyncline data, was related to large-scale emplacement of K-rich granites, into the upper crust, near the Archaean—Proterozoic boundary. Modelling of the Archaean sedimentary data is most consistent with an upper crust dominated by mafic volcanics and Na-rich granitic rocks (tonalites, trondhjemites). The contribution of K-rich granitic rocks with negative Eu anomalies to the upper Archaean crust is restricted to $< 10\%$. On the other hand, significant amounts ($> 30\%$) of exposed sialic material, in the form of Na-rich granitic rocks, were present. This is in conflict with some popular models of the nature of the early crust. Archaean REE patterns are observable as far back as the sedimentary record is available (> 3.7 Ga) indicating the crust of the earth was chemically and lithologically complex as far back as the geologic record is available.

Firm evidence for modern subduction-like processes is only clear back to about 1000 Ma ago, although initiation of such phenomena may have been related to the massive growth of the continents near the Archaean—Proterozoic boundary. The chemical evidence from sedimentary rocks, consistent with tholeiitic-trondhjemitic bimodal igneous activity, implies that tectonic conditions in the Archaean did not resemble present-day plate-tectonic subduction regimes.

THE CHEMICAL EVIDENCE: CONSTRAINTS AND LIMITATIONS

An important contribution of geochemistry to the earth sciences has been the ability to distinguish among igneous associations and relate these to plate-tectonic processes (Carmichael et al., 1974). The relationship between plate tectonics and geochemical associations shed much light on the physical and chemical processes of the outer part of the earth. It also allows reasonable conclusions to be drawn about the origin of ancient rocks. If the uniformitarian concept truly holds for the Archaean, it would be relatively simple to reach conclusions about the nature of geological processes in that distant era from the geochemical evidence.

Geochemical studies of Archaean crustal volcanic terrains are not abundant (Arth and Hanson, 1975; Condie and Harrison, 1976; Hallberg et al., 1976; Taylor and Hallberg, 1977; Fryer and Jenner, 1978). Most of these studies have noted distinct differences in trace element composition between Archaean and Recent examples of comparable major element composition. Some workers (Condie and Harrison, 1976 for example) have freely applied classical plate-tectonic models on the basis of chemical composition while others (Fryer and Jenner, 1978) have refrained from such interpretations.

Careful examination of the Archaean Marda complex (~ 2700 Ma.) of Western Australia is instructive in examining the possible limitations of geochemical data in providing evidence for plate-tectonic models in the Archaean (and perhaps much of the Precambrian). The Marda complex consists of low grade metamorphic volcanic rocks, including andesite and dacite flows overlain by subaerial rhyolite ignimbrites. Detailed study of the major and trace element composition of these lavas (Hallberg et al., 1976; Taylor and Hallberg, 1977) show a remarkable correspondence to modern high-K calc-alkaline volcanic suites (Fig. 21-1). The diagnostic rare earth element (REE) patterns are equivalent (Fig. 21-2). Fryer and Jenner (1978) also found similar andesites in the Prince Albert Group of Canada. The conclusion could follow from uniformitarian principles that these rocks were erupted in an island-arc environment. However, field evidence clearly argues against such a source for these rocks (Hallberg et al., 1976). They were erupted in a small basin containing 5 km of siliceous clastic sedimentary rocks, including banded iron-formations and gabbros. Classical island-arc models seem untenable for the origin of these rocks. What other process may be invoked? Derivation of such magmas by thermal plumes through a thin crust (Goodwin, 1973), mechanisms involving crust—mantle mixing (Fyfe, 1973) or anatexis at the base of the Archaean crust (Hallberg et al., 1976) seem viable mechanisms.

Thus volcanic suites which mimic modern island-arc rocks were produced in the Archaean, but were formed in a different tectonic environment. Accordingly, some caution must be exercised in using the chemical evidence to construct tectonic models.

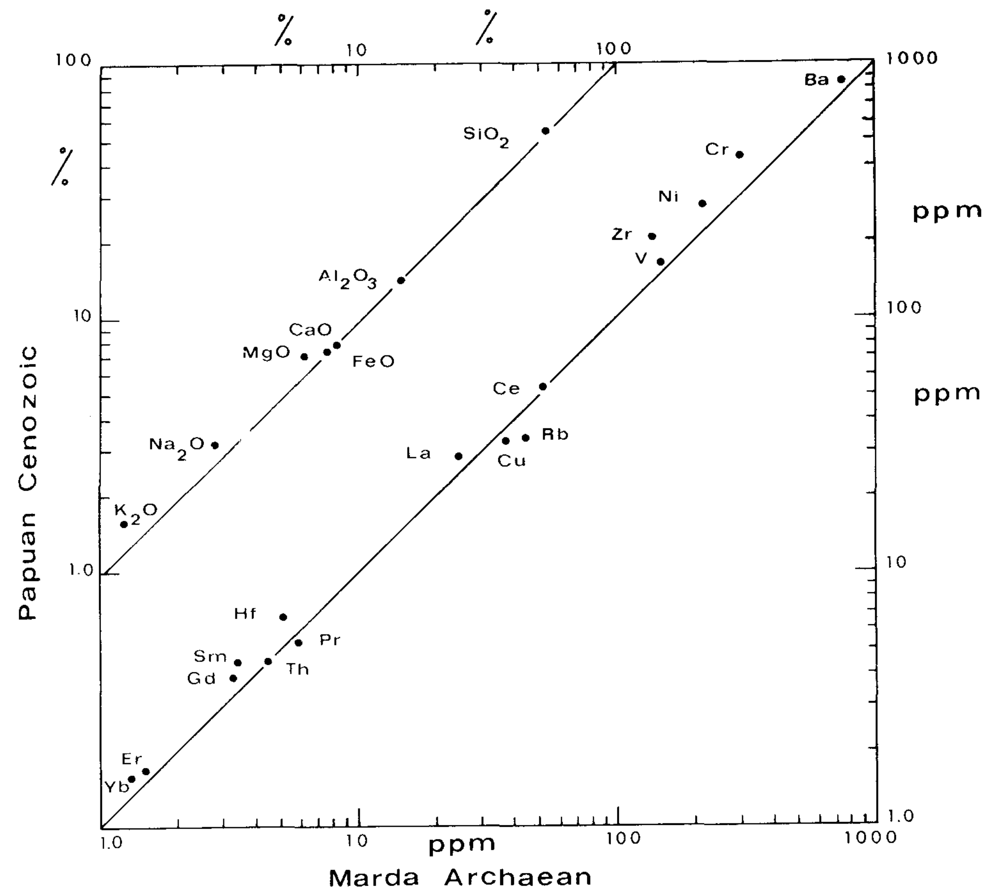

Fig. 21-1. Comparison of major- and trace-element data from a typical Marda complex andesite of Archaean age (Hallberg et al., 1976; Taylor and Hallberg, 1977) and a typical Cenozoic andesite from eastern Papua New Guinea (Smith, 1976). Points lying on the 45° diagonal lines indicate equality of composition for the two samples.

THE PRESENT COMPOSITION OF THE UPPER CONTINENTAL CRUST

Before commencing a study of Precambrian crustal evolution and the possible role of plate tectonics at that time, it is desirable to establish the present composition of the continental crust. Such information enables comparisons and extrapolations to be made with Proterozoic and Archaean crustal compositions which are necessarily less securely based. The present composition of the *upper continental crust* is reasonably well understood. Much detailed information has accumulated, based mainly on the use of extensive sampling programs (Shaw et al., 1967, 1976; Fahrig and Eade, 1968; Eade and Fahrig, 1971, 1973). Other workers (e.g. Goldschmidt, 1954; Taylor, 1964, 1977, 1979) have attempted to solve the compositional

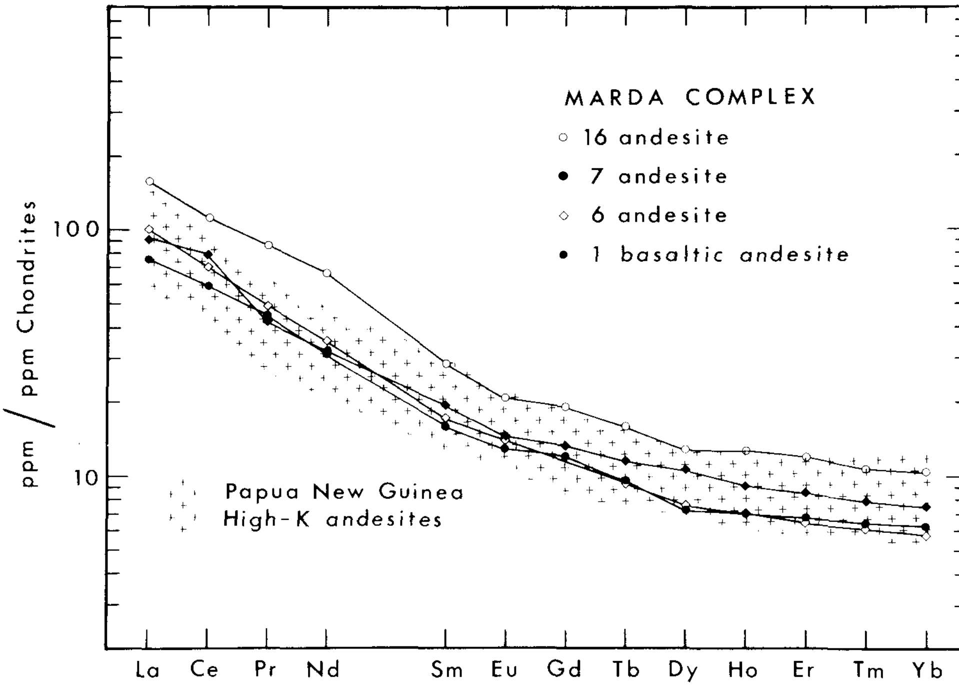

Fig. 21-2. REE patterns, normalized to chondrites, for andesites from the Archaean Marda complex (Taylor and Hallberg, 1977). Also shown is the field for Cenozoic high-K andesites from Papua New Guinea (Smith, 1976). The REE patterns for these two regions, widely separated in time and of differing tectonic environment, are indistinguishable.

problem of the exposed crust by employing the wide-scale natural sampling processes associated with sedimentation (e.g. glacial clays; rare earth elements (REE) in sedimentary rocks).

The observation that REE patterns in crustal sedimentary rocks were essentially uniform (Taylor, 1964, 1977, 1979; Haskin et al., 1966; Wildeman and Haskin, 1973; Jakes and Taylor, 1974; Nance and Taylor, 1976, 1977) coupled with the fact that these elements are not significantly affected by weathering, diagenesis and most forms of metamorphism (Herrman, 1970; Cullers et al., 1974; McLennan et al., 1979) enables the assumption to be made that processes of erosion and sedimentation are carrying out a wide-scale averaging of the upper, exposed crust for these elements. These patterns (Fig. 21-3) indicate that the average REE pattern (PAAS) is enriched in LREE relative to chondrites ($La_N/Yb_N = 9.2$ where N = chondrite normalized) and has a negative europium anomaly of constant magnitude ($Eu/Eu^* = 0.64 \pm 0.05$). These features are similar to the REE patterns observed in typical granodiorites. The widespread sampling and the averaging programs (e.g. Eade and Fahrig, 1971) indicate also that the major-element compositions of the upper crust as well as the trace element abundances

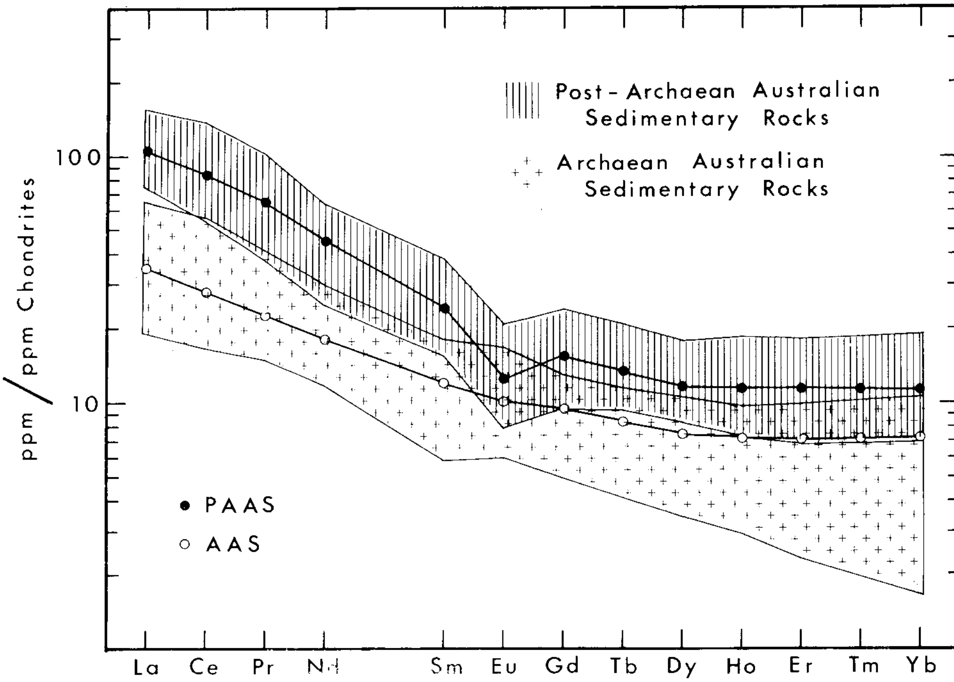

Fig. 21-3. REE patterns, normalized to chondrites, for the average Archaean sedimentary rock (AAS) and average post-Archaean Australian sedimentary rock (PAAS)(see Table 21-II). Also shown are the fields for individual samples of typical Archaean and post-Archaean Australian sedimentary rocks. Note the significant differences in the average patterns and the small area of overlap for the fields. AAS is similar to a typical andesite while PAAS resembles a granodiorite.

are approximated by granodiorite. These values, from Taylor (1979), with some amendments are given in Table 21-I.

TOTAL CRUSTAL COMPOSITIONS

Constraints on overall crustal composition come from various considerations. It must be capable of generating a granodioritic upper crust. It must be derived ultimately from the mantle by some common widespread and, hopefully, observable geological process. Such constraints make calc-alkaline, island-arc-type volcanism and associated igneous activity a viable candidate (e.g. Taylor, 1967, 1977) and this model is adopted here. Table 21-I lists a chemical composition which is compatible with such a model. The mechanism of the addition of new material to the present crust (required to offset the losses by erosion) may be volcanic, leading to volcanogenic sediments, or by intrusion of tonalites (andesite equivalents) and related lithologies. The production of this material by two or more stages from primitive mantle

TABLE 21-I

Composition of the post-Archaean continental crust

	Upper crust (%)	Total crust (%)	Lower crust (%)
SiO_2	66.0	58.0	54.0
TiO_2	0.6	0.8	0.9
Al_2O_3	16.0	18.0	19.0
$FeO^{\dagger}$	4.5	7.5	9.0
MgO	2.3	3.5	4.1
CaO	3.5	7.5	9.5
Na_2O	3.8	3.5	3.4
K_2O	3.3	1.5	0.6
Total	100.0	100.0	100.0

	Upper crust (ppm)	Total crust (ppm)	Lower crust (ppm)
Rb	110	50	20
Pb	15	7	3
Ba	700	350	175
Sr	350	400	425
La	38	19	9.5
Ce	80	38	17
Pr	8.9	4.3	2.0
Nd	32	16	8.0
Sm	5.6	3.7	2.8
Eu	1.1	1.1	1.1
Gd	4.7	3.6	3.1
Tb	0.77	0.64	0.58
Dy	4.4	3.7	3.4
Ho	1.0	0.82	0.73
Er	2.9	2.3	2.0
Tm	0.41	0.32	0.28
Yb	2.8	2.2	1.9
Lu	0.4	0.3	0.25
ΣREE	183	97	54
Eu/Eu*	0.64	1.0	1.1
Y	27	22	20
Th	10.5	4.5	1.5
U	3.0	1.25	0.4
Zr	240	100	30
Hf	7	3	1
Nb	25	11	4
Cr	35	55	65
V	60	175	230
Sc	10	30	40
Ni	20	30	35
Co	10	25	33
Cu	25	60	78
Zn	52	—	—

will not be discussed here (see review by Ringwood, 1974). The other candidates for eruption of voluminous material from the mantle (MORB, plateau basalts, oceanic island basalts) are too low in silica and other elements to be viable candidates for bulk continental compositions (which in this case would be about equivalent to bulk oceanic crust in composition).

THE PRESENT LOWER CRUST

Evidence for lower crustal compositions is inherently very difficult to obtain. The xenolith samples are random and the data base (particularly for REE) is very small. Examples of compositions close to those predicted by the above model are known (Lesotho kimberite xenoliths; Rogers, 1977; Taylor and McLennan, 1979). However, we still lack a large-scale process capable of providing an index of lower crustal composition with the same elegance with which we can determine upper crustal compositions. Well-exposed early Archaean granulites have been considered by a number of authors (Tarney and Windley, 1977, for example) as representative of the lower continental crust. This approach is probably not valid since much evidence (see below) indicates that the crust has not retained a uniform chemical character.

The island-arc model for continental growth firmly links plate tectonics with the addition of material to the continents. The production of the observable *upper* crust is due on this model to intracrustal melting to produce granodiorites leaving a depleted lower crust. For the model given above, the present lower crust has the composition given in Table 21-I. REE patterns for the upper, lower and total crust are given in Fig. 21-4. Also shown in this diagram is the average REE pattern for four lower crustal granulite xenoliths from the Lesotho kimberlite (Rogers, 1977). Although there is significant scatter among the individual samples, the average REE pattern is notable for its close resemblance to the lower crust pattern predicted from the "andesite model" for crustal growth (Taylor and McLennan, 1979).

Footnote to Table 21-I

† All Fe expressed as FeO.

Data for upper crustal composition adapted from the following sources: Major elements, Eade and Fahrig (1971) Canadian Shield data; REE from Nance and Taylor (1976) average post-Archaean Australian sedimentary rocks. Most other trace-element data from Shaw et al. (1976) values for Canadian Shield rocks. Data for overall continental crust composition is based on the assumption that the crust is derived from island-arc type volcanism (see Taylor, 1979). U values for the total crust are derived from $K/U = 10^4$ and Hf values from $Zr/Hf = 33$. Th values for the total crust are derived from $Th/U = 3.6$. The lower crustal composition is the residual remaining following extraction of the upper crustal values from the overall crustal composition.

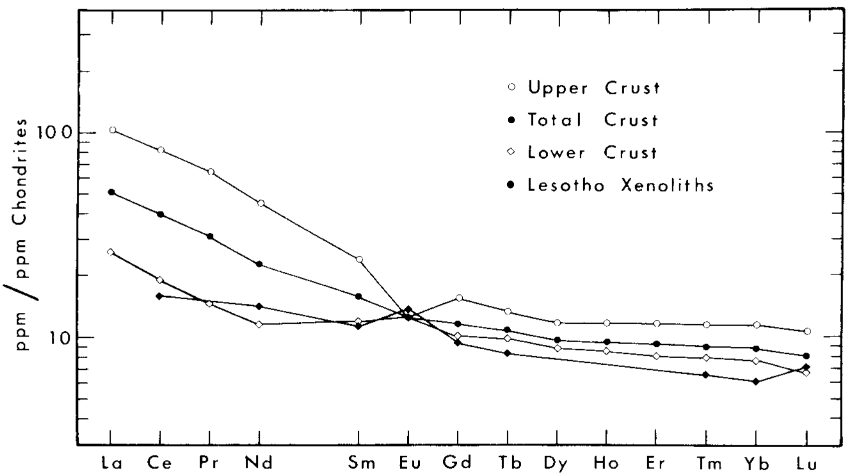

Fig. 21-4. REE patterns, normalized to chondrites, for the upper, lower and total continental crust (data from Table 21-I). The whole crust data are based on the model that the continental crust is derived from island-arc volcanic rocks. The upper crust is taken to be equal to the average sedimentary rock (i.e. PAAS). The lower crust is calculated from these data assuming that it comprises 2/3 of the entire crust. Also shown is the REE plot of the average composition of 4 lower crustal xenoliths from the Lesotho kimberlite (Rogers, 1977). Note the similarity of this pattern to the average lower crust and, in particular, the positive Eu anomaly.

MECHANISMS FOR PRODUCTION OF THE UPPER CRUST

What is the mechanism responsible for the present composition of the upper crust? The production of a "granodioritic" upper crust appears to be due to intracrustal processes. Several sources of evidence (heat flow, geochemical element balance calculations) combine to limit the thickness of the observable upper crust to 10—15 km. The thickness of the continental crust, down to the Mohorovičić discontinuity, is about 40 km without clearly defined seismic velocity breaks (except for the occasional presence of the Conrad discontinuity). The composition of the whole crust must be considerably more basic, containing less SiO_2, Th, U, K, etc. to account for the heat flow and geochemical balance arguments. Experimental evidence (Tuttle and Bowen, 1958) indicate that the production of granitic rocks occurs within the *P-T* range typical of crustal rather than mantle regions (< 10 kb, $1000°$C). Models for the production of the upper crustal granodiorites are not discussed in detail. Their production by partial melting processes seems established. The relative importance of removal of elements from the lower crust by metamorphic processes (Heier, 1973, 1978) or by fluid phases (Collerson and Fryer, 1978) is inherently more difficult to evaluate. The processes responsible for the generation of the upper crust must involve crystal—liquid equilibrium to produce the observed LREE enrichment and Eu depletion as well as large volumes of granodiorites.

The Eu depletion in upper crustal rocks

The Eu depletion in upper crustal rocks must be intracrustal in origin. No common volcanic rocks derived from the mantle exhibit negative Eu anomalies of the order (Eu/Eu* = 0.64) typical of the average upper crust. Mid-ocean ridge basalts, intraplate volcanics (e.g. Hawaii) and island-arc volcanics alike are characterized by the absence of positive or negative europium anomalies. The sporadic occurrences are associated with the presence, or absence, of cumulate plagioclase.

The basic cause of the difference in behaviour between Eu and the other rare earth elements lies in the difference in valency and crystal radius between Eu^{2+} (r = 1.39 Å for 8 fold CN) and Eu^{3+} (r = 1.21 Å for 8 fold CN, Shannon, 1976). The trivalent radius for Eu forms part of the monotonic decrease in radius exhibited by the trivalent REE. Eu, halfway through the REE sequence, is more readily reduced than the neighbouring REE. The increase in radius of the divalent ion causes it to enter different crystal sites to those available to the smaller trivalent ions. The Ca sites in feldspars readily accept Eu^{2+}, which closely mimics Sr^{2+} in radius. Thus the most likely mechanism to produce Eu depletion is partial melting where feldspar is a residual phase. Since plagioclase is not a stable phase below about 40 km (10 kb), Eu anomalies due to this cause are thus produced by shallow intracrustal processes. This explanation involves crystal—liquid equilibria rather than other processes (e.g. aqueous transfer).

Constraints on sediment subduction

As noted above, no Eu depletion is observed in common and voluminous volcanic rocks being erupted from mantle sources, except for some trivial exceptions. A current geological debate concerns the amount of continental material which is recycled through the mantle (Armstrong, 1969; Fyfe, 1978).

A variety of structural (e.g. Karig and Sharman, 1975) and isotopic constraints (e.g. Moorbath, 1977) argue against this interpretation. A further difficulty arises if upper crustal continental material is subducted. This will bear the characteristic Eu depletion signature (Eu/Eu* = 0.64). Removal of this deep anomaly will call for complex mixing and rehomogenization of the REE. Simple remelting will not remove the Eu anomaly and it could be expected to persist, regardless of oxidation conditions, as an inherited signature of the upper crust in lavas which involve some contribution from this material. As is shown later (p. 538, Fig. 21-6) even small contributions of material with a negative Eu anomaly have a significant effect on the resulting REE pattern. Accordingly, we wish to propose that the absence of a negative

Eu anomaly in mantle-derived rocks restricts the contribution of post-Archaean sediments in their source regions to trivial amounts.

UNIFORMITY OF CRUSTAL COMPOSITION WITH TIME

How far back in time can the present crustal composition be extrapolated? Nance and Taylor (1976) and McLennan et al. (1979) have established that there is no significant change in the upper crustal compositions, as represented by the sedimentary REE patterns, until the early Proterozoic. A major change is associated with the Archaean—Proterozoic boundary, conventionally set at 2500 Ma ago. The details of this change, best demonstrated from the Huronian and Pine Creek Geosyncline sequences, will be discussed later. The evidence from a major chemical change at the Archaean—Proterozoic boundary is reinforced by data from many sources (e.g. Sr isotopes, Veizer, 1976; Sm-Nd isotopes, McCulloch and Wasserburg, 1978).

On the basis of REE evidence, the island-arc model for the production of new continental material can thus be inferred to act uniformly back to the base of the Proterozoic. The role of crustal reworking resulting from intraplate tectonic activity (Wynne-Edwards, 1976; Kröner, 1977), which seems to be an important process in forming orogenic belts during much of the Proterozoic, cannot be addressed easily from the REE evidence. Firm geological evidence for modern subduction processes is only available back to about 1000 Ma ago (see Fyfe, 1976). It is clear from the palaeomagnetic evidence (McElhinny and McWilliams, 1977; Morris et al., 1979) that the configuration of the continents during most of the Proterozoic was much different from the present situation. It is also clear that the continents did move by large distances. Such movement would probably have been associated with some type of subduction processes. The REE evidence from the continental crust is consistent with the operation of presently observable processes of continental evolution back to the beginning of the Proterozoic Era. This uniformitarian approach does not appear to hold for the Archaean.

REE PATTERNS IN ARCHAEAN SEDIMENTARY ROCKS AND THE COMPOSITION OF THE ARCHAEAN CRUST

A significant change in REE patterns, matched by other parameters, occurs in sedimentary rocks at the Archaean—Proterozoic boundary. These patterns, also shown in Fig. 21-3, have significantly less LREE enrichment ($La_N/Yb_N = 4.8$) and they do not normally have any significant Eu anomaly. Several questions arise immediately about these patterns. Are they representative of Archaean sedimentary rocks? Do they provide us with an estimate of the Archaean crust, exposed to erosion, in the same manner that the REE patterns in post-Archaean rocks have been used? The sampling is inevitably more restricted. REE patterns also commonly show more variation in the Archaean sediments than appears to be the case for post-Archaean examples.

Sedimentary rocks used to estimate Archaean upper crustal abundances are derived exclusively from greenstone belts. This raises the spectre of biased sampling so that the Archaean REE crustal pattern so derived could be unrepresentative of exposed Archaean crust. This is a crucial consideration. Several pieces of evidence strongly indicate that Archaean sedimentary rocks do represent a wide-scale sampling of all major Archaean lithologies. A recent survey of Archaean mafic and ultramafic volcanic rocks (Sun and Nesbitt, 1978) indicate that relatively flat REE patterns are typical of greenstone lithologies. Felsic differentiates, which would be expected to be LREE enriched (Fryer and Jenner, 1978; for example), probably represent no more than about 10—11% of a typical greenstone pile. This is too small a volume to account for the shape of typical Archaean sedimentary REE patterns. Volcanic rocks which have REE patterns indistinguishable from modern calc-alkaline volcanics do exist but are quantitatively unimportant.

There are localized examples of Archaean sedimentary rocks which display very steep REE patterns (Nance and Taylor, 1977) and which are clearly derived from nearby granitic rocks with closely similar patterns. Archaean sedimentary REE patterns may also exhibit positive Eu anomalies (Nance and Taylor, 1977) with magnitudes far greater than those found in greenstone belt lithologies. Such a characteristic is best explained by plagioclase accumulation from typical Archaean Na-rich granitic rocks which may also display such Eu behaviour (O'Nions and Pankhurst, 1978). Convincing evidence for widespread sampling is also available from detailed studies in the Archaean greenstone belt from the Kambalda district, Western Australia. REE patterns from this area, where sedimentary rocks are sandwiched between basaltic and ultramafic lavas, are not influenced by the local environment (Bavinton and Taylor, 1980). The patterns are very typical of Archaean sedimentary rocks and must reflect a more distant provenance. Detailed petrological and geochemical studies display a significant granitic component (Bavinton, 1979). A major granitic component (commonly > 25%) for many Archaean sedimentary rocks is well established (Pettijohn, 1943).

Another piece of evidence relating to the provenance of Archaean sedimentary rocks comes from Lower Proterozoic sedimentary REE patterns. Sedimentary rocks from the lower part of the Huronian succession (McKim, Pecors Formations) display REE patterns which are similar to those of Archaean sedimentary rocks (the significance of this will be discussed in following sections). These rocks were clearly derived from a complex granitic and greenstone terrain (see McLennan et al., 1979). Similar characteristics are found in Lower Proterozoic sedimentary rocks from the Pine Creek Geosyncline, Australia (McLennan and Taylor, 1980). It is thus well established that Archaean-like REE patterns can be derived from such a complex provenance (see p. 538).

Accordingly, we can safely conclude that the Archaean sediments do indeed represent a widespread and, by analogy with post-Archaean sediments, an average sampling of the exposed Archaean crust, and that we may with suitable caution proceed to an overall estimate of the upper Archaean crust. The resulting REE patterns (Fig. 21-3) most closely resemble those of present-day island-arc volcanic rocks. The absence of an Eu anomaly indicates the lack of demonstrable intracrustal melting or of crystal—liquid differentiation at depths where plagioclase is a stable phase. The lower LREE enrichment likewise indicates a more primitive or less evolved composition. This composition appears to be similar to the bulk composition of the present crust (Table 21-I).

The evidence from the Akilia metasediments of West Greenland indicates that this situation extends back to 3.7 aeons. We adopt the model here that a cataclysmic bombardment, similar to that which afflicted the Moon, Mars, Mercury and probably Venus (as well as Ganymede and Callisto), struck the Earth at or before 3.9 aeons, obliterating the earlier record.

THE ARCHAEAN—PROTEROZOIC TRANSITION AND THE ABUNDANCE OF K-RICH GRANITES IN THE ARCHAEAN CRUST

An underlying theme in this discussion has been the distinct difference between Archaean and post-Archaean sedimentary REE patterns and, by inference, the difference in upper crustal compositions. Some understanding of the reasons for this difference is now available. Detailed studies of Lower Proterozoic sequences in Canada and Australia (McLennan et al., 1979; McLennan and Taylor, 1980) reveal a rapid evolution of sedimentary REE patterns associated with the Archaean—Proterozoic boundary (Fig. 21-5). It has been suggested (McLennan et al., 1979) that the large volumes of K-rich granitic rocks intruded into most shield areas (Goodwin, 1972, for example) were responsible for the change. The rapidly changing REE patterns documented in the Huronian (Fig. 21-5) represented increasing unroofing of these rocks at the end of the Archaean.

Can quantitative limits be placed on such a model? A compilation of 36 late Archaean ($<$ 3000 Ma) K-rich granitic rocks has been made (Table 21-II). Mixing calculations between this composition and the Average Archaean Sedimentary rock (AAS) indicate that about 1/3 K-rich granite must be added to the upper Archaean crust to produce typical post-Archaean sedimentary REE patterns (Fig. 21-6). Such a calculation is approximate since complexities arise from other lithologies derived from late Archaean plutonism as well as the presence of felsic differentiates in the greenstone belts.

K-rich granites and granodiorites appear sporadically throughout the Archaean (e.g. Greenland; Mason, 1975) but it is unclear how much of this evolved granite was present in the early Archaean crust. Evidence from

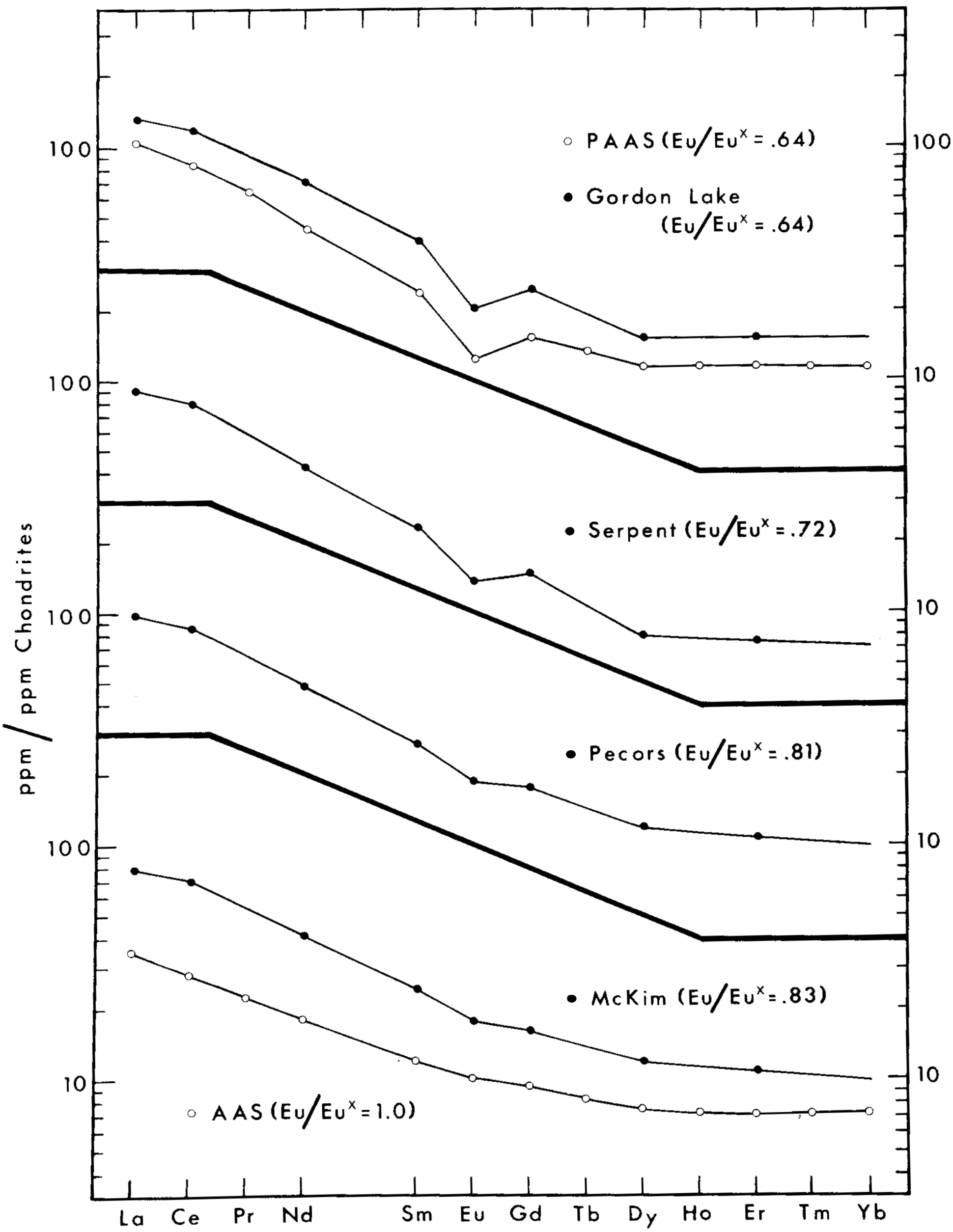

Fig. 21-5. Average REE patterns, normalized to chondrites, for fine-grained sedimentary rocks from selected Lower Proterozoic Huronian formations. The formation averages are presented in stratigraphic order (data from McLennan et al., 1979). Also presented, for comparison, are AAS and PAAS. Note the similarity of the lower Huronian units (McKim and Pecors Formation) to AAS except for slightly higher La_N/Yb_N and a slight negative Eu anomaly. These patterns are considered to be slightly evolved from AAS towards PAAS. REE patterns from the top of the Huronian (Gordon Lake Formation) are indistinguishable from PAAS. Data from Lower Proterozoic sedimentary rocks from the Pine Creek Geosyncline, Australia, show similar trends (McLennan and Taylor, 1980).

TABLE 21-II

REE abundances in AAS, PAAS, average late Archaean K-rich granite and chondrites

Element	AAS[1]	PAAS[2]	K-Granite[3]	Chondrites[4]
La	12.6	38	96	0.367
Ce	26.8	80	179	0.957
Pr	3.13	8.9	—	0.137
Nd	13.0	32	—	0.711
Sm	2.78	5.6	11.7	0.231
Eu	0.92	1.1	1.3	0.087
Gd	2.85	4.7	—	0.306
Tb	0.48	0.77	1.4	0.058
Dy	2.93	4.4	—	0.381
Ho	0.63	1.0	—	0.0851
Er	1.81	2.9	—	0.249
Tm	—	—	—	0.0356
Yb	1.79	2.8	4.0	0.248
Lu	—	—	—	0.0381

[1] AAS = Average Archaean Sedimentary rock. Values derived from average of typical fine-grained Archaean sedimentary rocks from Western Australia (Nance and Taylor, 1976; Bavinton and Taylor, 1980).
[2] PAAS = post-Archaean Average Australian Sedimentary rock. Values from Nance and Taylor (1976).
[3] Average of 36 late Archaean K-rich granitic rocks. Data from Condie and Hunter (1976); Chou et al. (1977); Nance and Taylor (1977); McLennan et al. (1979) and McLennan and Taylor (1980).
[4] Chondrite normalizing factors. These values are derived from the Type I carbonaceous chondrite abundances (Evensen et al., 1978; Mason, 1979) multiplied by 1.5 to allow for removal of volatiles (mainly carbon compounds and H_2O). These values are systematically 18% higher than those given in Taylor and Gorton (1977).

mixing calculations, similar to those given above, reveals that the contribution from such rocks must have been less than 10%.

Figure 21-6 illustrates the effect on the Archaean REE patterns of the addition of typical K-rich granitic material. Apart from the LREE enrichment the most diagnostic feature is the Eu depletion which the granitic debris imposes on the Archaean sedimentary pattern. As is evident from Fig. 21-6 the presence of more than about 10% "granite" (s.l.) will produce a discernible depletion in Eu in the patterns.

THICKNESS OF THE ARCHAEAN CRUST

The strong negative Eu anomalies seen in late Archaean K-rich granites, and which are characteristic of the post-Archaean upper crust, are explained by partial melting in the lower crust with feldspar as a residual phase (see p. 535). The thickness of the late Archaean crust is thus critical for such melting to be intracrustal.

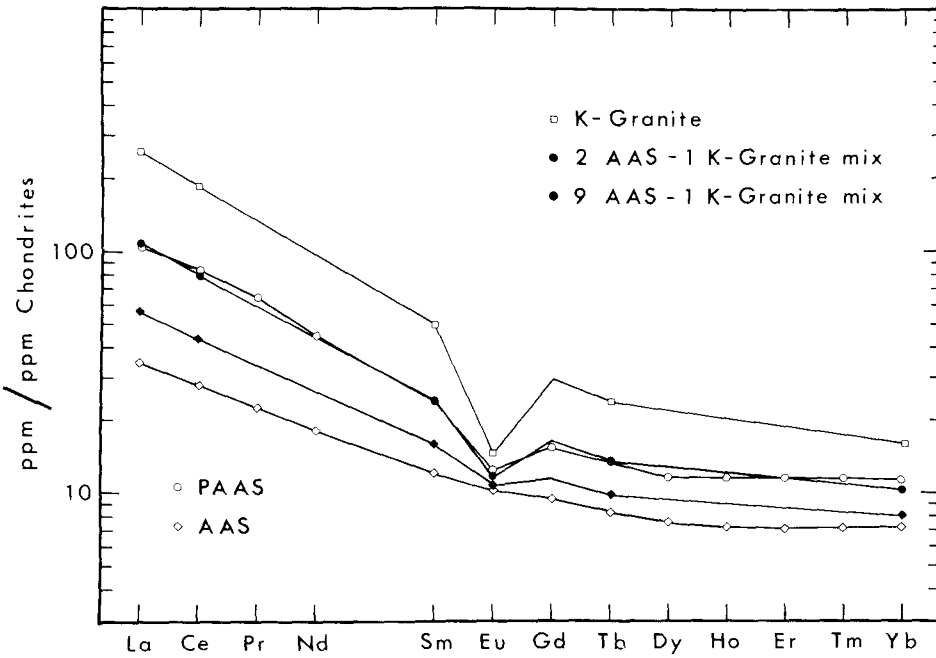

Fig. 21-6. REE diagram, normalized to chondrites, for AAS, PAAS and the average of 36 late Archaean K-rich granites (see Table 21-II). Also shown are REE patterns for various mixes of AAS and average K-rich granite. Note that even a small amount of K-rich granite (10%) mixed with AAS results in a significantly Eu anomaly. This indicates that K-rich granites were not a significant source for Archaean sedimentary rocks. A mixture of 2/3 AAS and 1/3 K-rich granite results in a REE pattern virtually indistinguishable from PAAS. The intrusion of K-rich granites near the end of the Archaean appears to be responsible for the change in the composition of the upper continental crust.

Present concepts are highly polarized with estimates ranging from very thin (< 15 km; Fyfe, 1973, 1978; Hargraves, 1976) to thick crusts (25—30 km; Tarney and Windley, 1977) with some estimates as high as 80 km (Condie, 1976). Collerson and Fryer (1977) have noted that estimates for thin crust are derived from examining early Archaean rocks (> 3000 Ma). The evidence from metamorphic and experimental studies (Collerson and Fryer, 1977; Newton, 1977; Tarney and Windley, 1977) strongly suggests that by about 2900 Ma geothermal gradients had decreased sufficiently to allow crustal thickness of up to 40 km or more. In summary, the geological and geochemical evidence are consistent with a thick crust in the late Archaean. This permits the formation of K-rich granites by intracrustal melting, in accordance with the isotopic (McCulloch and Wasserburg, 1978) and experimental evidence (Tuttle and Bowen, 1958).

MODELS FOR THE ORIGIN OF THE ARCHAEAN UPPER CRUST

The average Archaean REE patterns which mimic those of calc-alkaline rocks could be derived from the mantle, unlike the post-Archaean upper crustal compositions. We thus arrive at the conclusion that the Archaean upper crust was much less "evolved" than that of Proterozoic and later times. This conclusion agrees for example with the Sm-Nd models of McCulloch and Wasserburg (1978).

By following the same line of reasoning as was used to determine the composition of the post-Archaean upper crust, it may be deduced (Taylor, 1979) that the Archaean upper crustal composition is close to that of average island-arc volcanic rocks (that is, it is very similar to the overall bulk composition of the present crust). Does this constitute evidence for uniformitarianism? The answer is no, or at least equivocal. The evidence for large-scale "calc-alkaline" volcanism of conventional type is not persuasive. A number of examples occur (see pp. 536—538) but these do not appear to comprise a large fraction of Archaean rocks exposed at present. Although such volcanics are rapidly eroded to form volcanogenic sandstones, the evidence both from the greenstone belts and the "granitic" terrains is that of two principal rock types. These are tholeiitic basalts and tonalitic or trondhjemitic granitic rocks. The former have mainly flat REE patterns while the latter are typified by steep LREE enriched-HREE depleted patterns (O'Nions and Pankhurst, 1978; Arth and Hanson, 1975; Drury, 1978). Localized sedimentary REE patterns (Nance and Taylor, 1976) occasionally show this distinctive pattern. Mixing of the tholeiitic patterns with these steep REE patterns can reproduce the Archaean sedimentary rock pattern by this mechanism (Nance and Taylor, 1976; Taylor, 1977), so that the REE evidence cannot be used to distinguish this model from an overall calc-alkaline crustal model. Attempts to use Ni and Cr abundances in the sediments (which could be higher than in more recent sediments due to the input from the tholeiitic component) do not clearly distinguish Archaean from post-Archaean sediments (Taylor, 1977). We assume that the dominant rock types produced in the Archaean were tholeiitic basalts and tonalitic-trondhjemitic intrusives with minor production of calc-alkaline rocks indistinguishable from their modern analogues (Taylor and Hallberg, 1977).

Quantitative estimates of the various lithologies are inherently difficult to make but some significant constraints are available. Let us consider the two component system, consisting of greenstone and granitic material (which would be dominated by the tonalite-trondhjemite series). The average REE pattern of greenstones is best approximated by the average Archaean mafic volcanic. This pattern is flat (Sun and Nesbitt, 1978)(minor fluctuations in the LREE do not affect this argument) at about 8—11 times chondritic values. A value of 10 × chondrite is adopted for this discussion. Using the values and the REE pattern for the average Archaean sedimentary rock (AAS) we can calculate the probable shape of the average REE pattern in the tonalite-trondhjemite component for varying mixing proportions.

Although Archaean tonalites and trondhjemites possess a wide range of

REE concentrations they are characterized by steep patterns. Glikson (1979) has recently summarized much of the available data. The range of *most* of these rocks can be represented by a field ranging from 30—100 x chondrite at La to about 0.5—5.0 x chondrite at Yb (Fig. 21-7). Any Eu anomalies must average out or they would be seen consistently in the sedimentary data. If we compare the known range of REE patterns in the tonalite-trondhjemite suite with the values obtained by the mixing calculations, we can constrain the proportion of this material exposed in the Archaean crust. The proportion must have been more than 30% to explain the HREE abundances in AAS

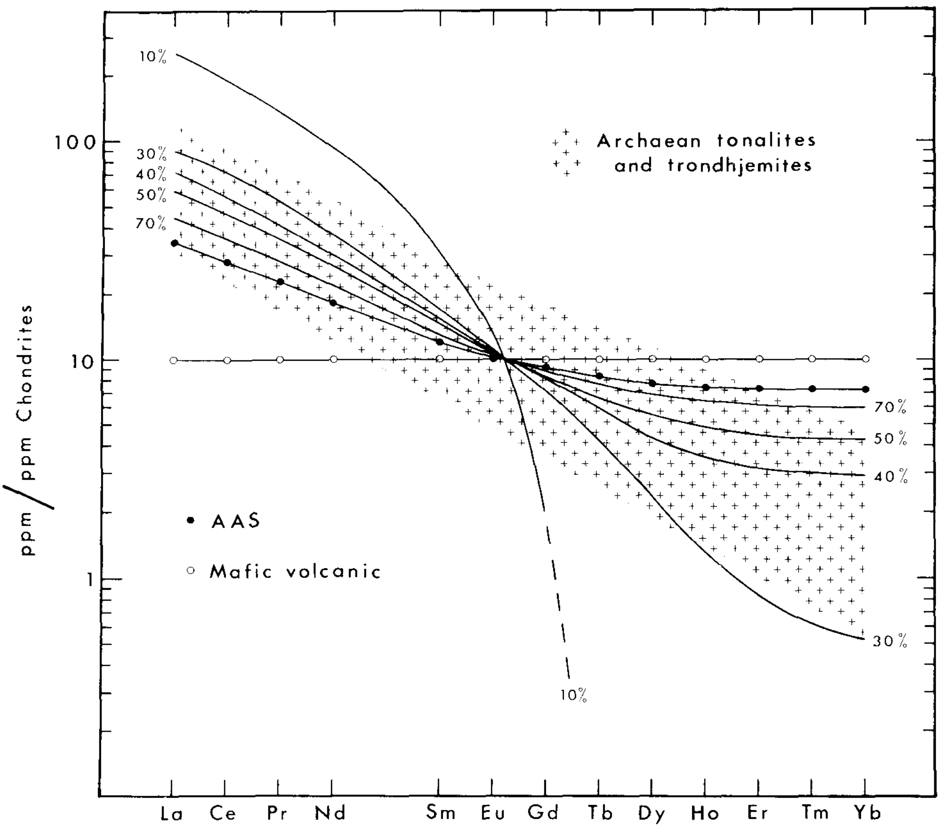

Fig. 21-7. REE diagram, normalized to chondrites, for AAS, average Archaean mafic volcanic (see text) and a field representing the majority of Archaean Na-rich granitic rocks (tonalite—trondhjemite suite). Also shown are the expected average REE patterns of Na-rich granitic rocks for various mixing calculations (assuming that AAS equals the average Archaean upper crust REE pattern); for example, the 30% REE pattern is the expected average for Na-rich granitics if they represent 30% of an exposed crust composed of mafic volcanics and Na-rich granitic rocks (see text for discussion). It is clear that reasonable mixes of Archaean basalts and Na-rich granites can produce the average Archaean sedimentary rock pattern.

(note that lowering REE levels in the mafic volcanic would increase this value) and the La_N/Yb_N ratios seen in the ancient granitic rocks. An upper limit of about 60% is also constrained by the La_N/Yb_N of the granitic rocks.

Given reasonable values for the average La_N/Yb_N, we would suggest that the range of ratios of tonalite-trondhjemite series rocks to mafic volcanics is from 1:2 to 1:1. Such an estimate is only a first order approximation since calculations would be complicated by the presence of lithologies such as ultramafic rocks, felsic differentiates and less fractionated granitic rocks such as the K-rich variety. The proportion of these rocks is low and their REE patterns tend to "average out" to AAS. Thus, the suggested ratios are not changed by these factors.

These calculations enable some tests to be made on current models of crustal evolution and growth. Hargraves (1976) has proposed that by about 3.5 Ga the earth had a globe-encircling sialic crust which was completely submerged beneath the oceans. During the period 3.5—2.5 Ga greenstone piles dominated the land masses above sea level. This model has been supported by Fyfe (1978). Such a scenario is in conflict with the sedimentary REE data. It is clear that significant amounts of greenstone and granitic debris were incorporated into sedimentary rocks throughout the Archaean. This indicates that the upper exposed crust was chemically and lithologically complex for as far back as the sedimentary data are available (Akilia association, > 3.7 Ga; McGregor and Mason, 1977; Taylor, 1979).

SPECULATIONS ON TECTONIC CONDITIONS IN THE ARCHAEAN

What were the geological conditions which gave rise to the Archaean crust. It appears that extrusion of basalts (some indistinguishable from MORB; Sun and Nesbitt, 1978) took place under conditions probably not much different from today, except for higher heat flow and more rapid mantle convection. Such conditions imply that sea-floor spreading occurred, a view reinforced by the lateral extent of greenstone terrains. Possibly large volcanic domes were also constructed. Sinking of dense basaltic material could lead to the formation of mafic amphibolites, mafic granulites or eclogites depending on *P-T* conditions. Partial melting of such materials at depth produces the tonalitic and trondhjemitic intrusives. Their steep REE patterns are a consequence of equilibration with garnet as a residual phase. Weathering and erosion of both these terrains (which must have been above sea level, pace Hargraves, 1976) produces the typical Archaean sedimentary REE patterns. This model is similar to that of Tarney et al. (1976). It involves a distinctive tectonic regime which commences with extrusion of tholeiitic basalt, possibly followed by sea-floor spreading, piling up of tholeiitic masses and sinking of these which, in turn, produce the more acidic intrusives by partial melting. The intrusive masses coalesce, or are pushed together, to form the Archaean cratonic nuclei.

At the end of the Archaean, massive intracrustal melting produces a granodioritic upper crust and the continental crust assumes its present character. This event is conventionally dated at 2500 Ma, although the actual age varies on each continent (e.g. Cloud, 1976). The continental regions now form massive barriers to sea-floor spreading and basalt sinking of Archaean times, largely as a consequence of increasing size. Initiation of modern subduction-type tectonics occurs when the sea-floor basalts spreading from ridges, encounter the buoyant continental masses.

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Chapter 22

HOW DO WE RECOGNIZE PLATE TECTONICS IN VERY OLD ROCKS?

W. S. FYFE

ABSTRACT

It is considered that Archaean volcanic and metamorphic patterns indicate more closely spaced, mantle magma penetration sites than for the present earth. If this is valid, lateral lithosphere motions may also have been smaller. It is considered that "hot-spots" may have been quantitatively more significant in the past. As there is mounting evidence that light materials can be subducted, models involving steady increase in the mass of continental crust must be questioned. In particular, the lack of aluminous blueschists may indicate massive recycling of pelagic sediments. It is suggested that the scale of mixing interactions between hot and heavy mantle magmas underplating crust and anatectic high-level melts may have been underestimated and may have played a major role in the generation of the ancient grey gneiss complexes.

INTRODUCTION

The purpose of my brief contribution to this volume is to consider the question involved in the above title. Put another way we ask the question as to whether or not the presently operating convective style of the earth is adequate to explain the ancient geologic record. I think that when one approaches the problem of the nature of processes in the early earth, we must first remind ourselves of how recently we discovered plate tectonics, and of the types of observations that were vital to the acceptance and quantification of this model. We must remember too, that when we go back in time, the sample of unmodified early surface is very small.

Let us imagine for a moment that we are presented with a sample of only a few percent of the earth's surface; that this surface has no topography, its rocks carry no reliable palaeomagnetic record and heat flow cannot be measured. Accept too that we do not have the tools of seismology, that we do not know the dimension of either continents or oceans, that there is no fossil record and that we can measure ages only with a precision of say ~ 100 Ma. Given such limitations of observation for the modern earth, would we have discovered plate tectonics? These are the problems we face when we consider ancient tectonics and we are forced to concentrate on only those parts of the record which are likely to be preserved and those parts of the record with unambiguous significance.

It is clear that ancient tectonics must involve primarily the mineralogist, petrologist, geochemist and structural geologist; the full power of modern geophysics will play a smaller role, and quantification of ancient dynamic processes must be very difficult.

I think there are certain fundamental starting points for our approach to the ancient earth. For me, one such point is John Elder's (1976) concept of "roll-over time", residence time or the time it takes to modify old surface or form new surface. At the present, this time for oceanic crust is in the order of 100 Ma. As stated by Uyeda (1978, p. 120) "since the Mesozoic era, nearly 10 000 km of oceanic plate along with ocean ridges seems to have descended beneath Japan". Given such a modern observation, it is possible that two fragments of Archaean surface now welded together, could have been "around the world" in the time contained in the errors of absolute dating.

If so much new oceanic crust can form in something like 100 Ma at present rates, what of the past? The early earth must have been much more energetic. Exactly how much depends on the model of accumulation and dissipation of energy (Runcorn et al., 1977; Runcorn, 1978). But prior to 3 Ga ago, a roll-over time in the order of 20 Ma might have been possible. Thus, given the uncertainty of absolute age dating of old rocks, correlation of events across fragments of ancient surface may be misleading.

I think there is another very vital consideration in our approach to the ancient earth. When objects cool at the surface by internal convective processes, these processes are more turbulent and temperature gradients are more erratic or noisy, if the body is hotter and is losing more heat (see Elder, 1976). This implies that events in the early earth would be less regular than for the present, and tectonic styles should also be more varied (see Young, 1978).

The present earth operates in a rather smooth manner. Basalt produced in the hot mantle invades the crust at widely separated ocean ridges. Basalt-crust production leading to ocean-floor spreading proceeds at a rather constant rate. The new crust cools largely by seawater convection which may penetrate almost to the oceanic Moho (Lewis and Snydsman, 1977): As a result, thermal gradients and heat flow are drastically perturbed. Oceanic lithosphere thickens by cooling as it moves away from the ridges, eventually becoming heavier than the underlying asthenosphere and sinks at subduction zones. As the lithosphere bends at the trench, it cracks and these cracks may trap sediments and carry them into the mantle (Jones et al., 1978). Degassing of the descending slab triggers melting in or above the slab; the melts rise to form the andesite volcanoes. If the mantle melts rise under continental crust this crust may melt and produce granodiorites and the like. There is good evidence that mantle and crustal melts mix on a grand scale (Eichelberger, 1978) and produce hybrid melts. Above the subduction zone, the combination of cool descending plate and hot rising magma produce the characteristic paired metamorphic belts of Miyashiro (1961).

Our models of this process are quantified by geophysical observations and obvious topographic surface features. Petrological-geochemical arguments on the evolution and changes from basalt → spilite → blueschist → ecologite → andesite → granite correlate with the tectonic model. All these phenomena ultimately root in concepts of heat production and transfer and the resulting heat-flow patterns. Modern plate tectonics involves vast areas of crust undergoing lateral motion only and small but very long zones where vertical motions dominate. The lateral motions in oceanic crust are quantified by measurements of age, magnetism and heat flow. The vertical motions are defined by heat flow, topography and seismic events.

If we are to quantify ancient regimes similarly, then at present all we have to work with are approximate ages, petrological and structural patterns. If such patterns for the ancient surface are not similar to those of the present, then presumably the convective style must also differ.

MAJOR QUESTIONS

When I consider the topic of this book certain questions come to my mind which must eventually be answered:

(1) Were there vast areas of ocean crust of the modern type?

(2) Was this oceanic crust conserved (more or less) as it is at present, i.e. did subduction balance formation at ocean ridges?

(3) How were ocean ridges spaced globally if they existed?

(4) If subduction occurred was the process similar and how were subduction zones spaced?

(5) What was the rate of mantle magma production compared to the present?

(6) Was there great periodicity in magma production and continental growth as some writers have suggested?

(7) Were the oceans similar in extent and volume?

(8) Were there regions of oceanic and continental crust of similar type to the present?

(9) If there was continental crust and oceanic crust, how much of each?

It is easy to ask the questions but I doubt if we can adequately answer any at the present time.

If we contemplate geological maps of regions of ancient crust (Rhodesia, Canada, Greenland, etc.) and if we assume that apparent age differences of a few Ma may mean synchronous formation, then it is clear that in relation to the modern situation the ancient crust-forming processes were quite different to those of the present. Does Goodwin's (1978a) geotraverse map in the Superior Province of Canada showing a succession of greenstone and granite belts represent a single fragment of crust or the welded remnants of many mini-plates? If there were miniplates what was between the pieces and how did it move out of the way?

Perhaps there are a few features of Archaean geology from which some reasonable conclusions can be formed. If we look at typical volcanic successions (see Price and Douglas, 1972; Windley, 1977; Goodwin, 1978a, b, Fraser and Heywood, 1978) in the Archaean greenstone belts we seem to find something that is not modern. First, the successions often contain very high temperature, ultramafic lavas. Mantle which could produce both more or less normal basalts and komatiites must have been convectively turbulent and thermally erratic; large volumes of upper mantle could have been extensively molten. The problems of high-degree melting has been recently discussed by Arndt (1977). In the Archaean volcanic successions, there appears to be a sequential development (sometimes repeated) of volcanic rocks across the entire gamut of compositions from periodotite to rhyolite. All the types we now associate with an ocean ridge and its remote subduction zone are present and appear in the same stratigraphic sequence. And just as now, basalt dominates in volcanism, "granite" in plutonism (Goodwin, 1978b).

Today it is common to use the strontium isotope systematics to decide on a crust or mantle origin for the magmas. But recent observations (Armstrong et al., 1977; Eichelberger, 1978; Taylor and Silver, 1978; Elston, 1978) clearly show that many modern andesitic to rhyolitic rock types are a product of varying degrees of mixing of crust and mantle. From these studies it becomes increasingly clear that large volumes of mantle melt products do accumulate beneath and in lighter crust, causing crustal melting, and that mixing of the anatectic melt with the mantle fusion product can lead to hybrid magma types. Eichelberger (1979) has discussed possible mixing mechanisms (see also Fyfe, 1980). It should be noted that, given the densities of some komatiitic liquids (ca 3,0), it is surprising that they appear at the surface at all. They may have been much more voluminous than indicated by surface manifestations. Magma underplating by basalt or komatiite liquids may well have contributed to the Moho complexity under continents discussed by Oliver (1978). Convective and diffusional mixing of light anatectic melt and dense mantle melt seems probable.

Finally, I think it intriguing that, to my knowledge, a modern type ocean-floor ophiolite has never been reported from an Archaean greenstone belt. I know many who have searched for such complexes. Further, time after time, late granites appear to invade the greenstone belts. I think that until we find good examples of Archaean ophiolite complexes we cannot discuss modern type ocean-floor spreading or even the existence of modern type oceanic crust. If they exist, there are enough low-grade greenstone belts including rocks formed at the marine interface that we should be able to find them.

The examination of Archaean lithologies suggests to me that if ocean-floor spreading occurred, the scale was small. In Archaean terrains typical ridge and subduction zone products appear almost in the same place at the same time and granitic basement never seems far away.

What was the Archaean recycling process? I think that this is one of the greatest questions in Archaean geology. Petrologically we recognize modern subduction zones by features like andesite-granodiorite belts, blueschist-eclogite belts and the like. I see nothing in an Archaean geological map to match the modern scale or assembly of rock types. There are no glaucophane schists, no eclogites in the Archaean. In general, low-T/high-P metamorphic facies are not characteristic of the Archaean (see Lambert, 1976).

Perhaps there are some clues as to the nature of the Archaean recycling process. First, the eruption of dense, hot magma types like the komatiites may indicate that the lithosphere was much thinner or even that parts of the upper mantle were largely molten. Such liquids would tend to underplate the crust. Second, the appearance of complex magma series from basalt to rhyolite all in one place may suggest that basaltic magma edifices were poured out on thin crust essentially uniquely Archaean, being a mixture of present oceanic and continental crusts. The edifices may have collapsed back into their own plumes. Goodwin (1978c) uses the term sag-subduction. During this process, more siliceous elements in the base of the crust would melt out to form the less voluminous andesites, rhyolites and granitic types at the same time homogenizing with voluminous basaltic components. The picture might look more like a modern Hawaii (not on a moving plate), poured out on a granite-basalt sandwich (dominated by basalt), foundering as eclogite forms at the base. Perhaps the trigger for such vertical motions could be the settling of large volumes of deeply crystallized komatiite plus eclogite into hotter peridotite mantle. But fundamental to any such argument is the probably transient nature of mantle plumes rising through thin lithosphere on a short wavelength pattern.

PAN-AFRICAN SUTURES

Many modern studies of late Proterozoic regimes discuss possible evidence for plate collisions and the formation of ophiolite-decorated sutures during this period (Shackleton, 1979; Gass, 1977; Black et al., 1979) (*see also Caby et al., this volume, Chapter 16; Gass et al., this volume, Chapter 15; Leblanc, this volume, Chapter 17, ed.*). Certainly, the magmatic patterns in regions such as the Arabian Peninsula seem quite analogous to those above modern subduction zones. But in such Precambrian regions there is presently no evidence for the high-P/low-T facies of metamorphism as seen in, say, the Alps or Urals. Most rocks in the vicinity of "sutures" seem to be quite normal greenschist-facies rocks (see Nawab, 1978). Why are the blueschists missing in these older zones where there is otherwise normal magmatic evidence for subduction?

To form a massive zone of blueschist-eclogite metamorphic rocks, a wedge of sediments and basaltic rocks must be dragged down to depths in the order of 30 km. By the time the lower members achieve a mineralogy dominated by phases like garnet-omphacite-amphibole-lawsonite-kyanite, they are much denser than normal continental crust with most of the minerals having densities of three or more. Why should they rapidly rebound into structural positions such as seen in the Coast Ranges of California? And they must rebound fast to preserve minerals such as metamorphic aragonite (Brown et al., 1962).

Recently Moore et al. (1979) have shown that trench sediments tend to underthrust the earlier members of an accretionary wedge. The potential for uplift of early members deeply depressed thus increases the length of time over which the subduction process operates. Thus failure to produce observable blueschist zones may simply reflect the size of the oceans which closed. A highly controversial subject involves possible sediment subduction. Recent writings of Gilluly (1971), Sibley and Vogel (1976), Garrels and Lerman (1977) suffice to indicate that many features of sedimentary balances seem to point to the possibility of massive subduction. I would stress that many crustal components, when moved into their blueschist and eclogite facies state, are dense materials and if we consider spilitic, andesitic (greywacke) or pelagic sediment starting materials, the final products may all have densities well above 3 gcm^{-3} which would aid in their subduction. Pelagic sediments (Sibley and Vogel, 1976) are abundant and are Al-rich, but kyanite blueschists are rare, implying that they must be subducted. Uyeda (1979) shows clearly the evidence for pelagic sediment removal and even erosion of the continental plate.

CONTINENTAL GROWTH

One of the present great debates is concerned with the growth of continental crust. Perhaps the two extremes of the argument are to be found in Hargraves (1976) who favours complete early segregation and Moorbath (1978) or McCulloch and Wasserburg (1978) who prefer a steady growth with views being more moderate in Shaw (1976). One strong line of evidence for continental growth involves the analysis of strontium and samarium-neodymium isotope systematics (*see for example Moorbath and Taylor, this volume, Chapter 20, ed.*). But more and more evidence from modern orogenic belts shows that the interpretation of these data is complex and that, at present, crust mantle mixing is more pervasive than previously thought (Eichelberger, 1978; Taylor and Silver, 1978). As I have discussed previously (Fyfe, 1978) we do know that some "dissolved" continental material delivered to the ocean (at a rate of about 4×10^{15} ga^{-1}) is fixed in spilites and recycled. If this material is reprocessed through the mantle it will show mantle isotopes. Further, the subduction of light crust is mechanically possible (Molnar and Gray, 1979; Hargraves, 1978; *see also*

Dimroth, this volume, Chapter 13; Kröner, this volume, Chapter 3, ed.).

As many writers have stressed (see Windley, 1977) the high-grade basements of many ancient regions are dominated by grey tonalitic gneisses with varying contributions from more basic amphibolites. On the basis of their Sr isotope systematics many writers insist that these represent new additions to the crust from the mantle. But if some modern views on the origin of andesites have merit (Eichelberger, 1978; Fyfe, 1980) then these Archaean rocks may indicate vastly more pervasive mixing at the base of the crust. This might be no surprise if, on average, mantle melts were more basic and more dense. Underplating of scum crust and magmatic layering could produce the complex continental Moho's described by Oliver (1978).

I recall that in the 1960's my former colleague John Verhoogen (see Verhoogen et al., 1970) was reflecting on continental growth and found that he could come to no firm conclusion on the matter. In the modern plate-tectonic cycle we can roughly quantify continental growth by the addition of andesites and the like, we know something about Phanerozoic sedimentation rates, but we cannot yet quantify sediment subduction loss. I think that until we do, arguments about continental growth are likely to be to some extent futile. Present data indicate that continental erosion rates (Garrels and Lerman, 1977) far exceed magmatic additions above subduction zones (Brown and Hennessy, 1978). What is urgently needed is more refined seismic studies of the structure of descending slabs, particularly at their upper levels. Here I am most impressed by the recent work of Jones et al. (1978) on the Japan Trench. They report little evidence for massive sediment build-up which should occur if the sediment is scrapped off, and they also indicate cracking mechanisms that may have caused sediment removal. It would take very little cracking of the bending lithosphere at global subduction zones for significant removal of sediments; the cracks must fill with water or sediment (see also Fyfe, 1980). Cook et al. (1980) have shown clearly how thrust tectonics can build continental crust.

RAYLEIGH NUMBERS

The convective state of a cooling object is often discussed in relation to the Rayleigh number of the system defined as $R = \alpha\beta g h^4/K\eta$ (h the thickness, α coefficient of thermal expansion, β the thermal gradient, g the gravitational acceleration, K the thermal diffusivity and η the viscosity). As discussed by Elder (1976, p. 53) it is possible to classify the type of convection by the amount by which R exceeds a critical value needed for convection. If we look at estimates of R for the present earth (Elder, 1976; Uyeda, 1978) the values are so large that we might expect "chaotic eddying motions" (Elder) or turbulence. This does not appear to describe the present, rather steady, convective patterns. It does appear more appropriate for the Archaean patterns. If we reflect on the form of R in the above equation, it is clear that certain factors exert a major influence on changes. In particular larger values of the thermal gradient, and even more h, can change R by orders of magnitude. Thus, in the Archaean thicker and hotter (more molten) asthenosphere could greatly increase turbulence and create a high frequency of hot-spots.

CONCLUDING STATEMENT

From a knowledge of heat production through time we can approximately quantify the rates of mantle magma production. At present our knowledge of crust destruction, particularly continental crust, is not quantified. We can only be sure that if crust mantle recycling occurs as the earth cools, we must tend to gradually lose continental crust and hydrosphere (Fyfe, 1976, 1978).

There is good evidence from Pan-African terrains that the magmatic events of the late Precambrian were similar to those of modern style. But in the Archaean there is little evidence for the modern, large separation of constructive and destructive plate margins. The style of magmatic activity is more like that of the frequent small plumes as depicted by Shaw (1976, Fig. 2). Even given a greater magmatic intensity in the past (3—4 × present) if mantle penetration sites were numerous and closely spaced, lateral motions of the surface crust need not have been extensive. The Archaean recycling process is a mystery but some form of crustal foundering into cooling mantle plumes could be appropriate. Thus, in a sense, ancient tectonics could have been dominated by hot spot phenomena (*see for example Lambert, this volume, Chapter 18, ed.*) on an essentially stationary thin lithosphere (Fyfe, 1978). Obviously, as Ramberg's (1967) studies have shown, the thickness of the lithosphere will exert a controlling influence on the frequency of mantle eruptive sites. Komatiites may be telling us that the degree of melting in the asthenosphere was much greater than at present and the viscosity of the mantle much lower; essentially solid mantle of today may not have been present till the Proterozoic (Schwarz and Fujiwara, 1977).

Over the past five years or so we have just begun to appreciate the geophysical and geochemical consequence of "crustal water cooling" (Wolery and Sleep, 1976; Lewis and Snydsman, 1977; Burke and Kidd, 1978; Fyfe, 1978). It is only now that petrologists are becoming aware that basalt erupted is not spilite subducted. Geochemists do not seem to like massive return flow of "continental" elements (pace Heier, 1978; Moorbath, 1978) but there is no doubt that some does occur. If Aumento et al. (1976) and Sibley and Vogel (1976) are correct, there is massive feedback of uranium and potassium to the mantle. Such things must be quantified.

Thus, while geophysicists and sedimentologists seem to accept the possibility of sediment subduction, geochemists don't like the added complexity. Again we are all happy about sinking basaltic eclogite, but a quartz eclogite from basaltic andesite will sink just as nicely.

The Archaean was dominated by submarine igneous activity and thus all we are learning about modern hydrosphere—crust interaction must be applied to the Archaean on an even greater scale. We know the oceans were there 4 Ga ago, and water cooling must have limited the outer thermal gradients (Burke and Kidd, 1978). If at present 30% of the entire earth's surface contains presently active geothermal convection (Anderson et al., 1979) then 3 Ga ago, before massive continental emergence (as perhaps indicated by the date of Veizer and Jansen, 1979), this figure would easily have approached 100%. The crust of the earth would be modified by mantle underplating and hydrosphere convection over almost its entire surface.

Finally, the problem of sialic crustal growth rests on the question of crust—mantle recycling. Here I think we must wait for more refined geophysical studies and drilling in the trench environments. It is with good reason that Uyeda (1978) lists subduction among the "remaining questions". But if sediments and even andesites are subducted, then we must keep an open mind on the meaning of "age peaks". These age peaks may correspond to quantum jumps in convective style of the mantle as suggested by Runcorn (1965). But do we observe quantized creation or quantized preservation (Fyfe, 1978)? Perhaps in the next decades as Ion Microprobe Mass Analyzer techniques for dating reach their full potential (Hinthorne et al., 1979), we may see a great increase in our appreciation of ancient events.

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6. Precambrian palaeomagnetism

Chapter 23

ON THE COHERENCE, ROTATION AND PALAEOLATITUDE OF LAURENTIA IN THE PROTEROZOIC*

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ABSTRACT

Palaeomagnetic results from Laurentia are grouped mainly in the intervals from about 2300 to ~1650 Ma (early Proterozoic) and from about 1450 to ~930 Ma (middle Proterozoic). The data from the first part of the older interval (~2300 to ~1850 Ma) are in disarray. It is therefore not possible to determine whether or not the Slave and Superior Provinces had separate or common paths of apparent polar wander (APW) and hence whether or not they moved relative to one another, prior to and during the Hudsonian Orogeny, as plate tectonics would require. The early Proterozoic results are, however, sufficient to conclude that large but as yet very poorly described motions relative to the pole have occurred. Palaeomagnetic results are consistent with the Laurentian Shield, except for the Grenville Province, being a single entity since ~1800 Ma ago, and with the entire shield being together since about 980 Ma ago. For intervals of time of several hundred million years in the early and middle Proterozoic average rates of palaeolatitudinal motion of 5—6 cm, and occasionally as high as 10 cm per year, are recorded. The occurrence of such high average rates of motion means that in order to make satisfactory comparisons with results from other continents time correlations are needed to accuracies of 10 to 20 Ma. Such accuracies are rarely achieved. We argue that this is the reason why contradictory interpretations regarding Proterozoic continental drift have been made when such comparisons have been attempted. In middle and early Proterozoic rocks the palaeomagnetic signature of local rotations about vertical axes (declinations in contemporaneous rocks disagree, inclinations agree) is observed, and it is argued that such rotations were common in Proterozoic terrains as they are in Phanerozoic foldbelts. None of the palaeomagnetic evidence is inconsistent with plate tectonics but the record is fragmentary.

INTRODUCTION

The purpose of this article is to present our reasons for believing that the palaeomagnetic evidence from Laurentia is consistent with the supposition that during the Proterozoic plate-tectonic processes comparable to those of the Phanerozoic occurred. We shall also seek to explain why arguments to the contrary, based on palaeomagnetism, have sometimes been made. We have attempted to critically appraise the progress made in the application of palaeomagnetism to tectonic problems in the Precambrian. We shall focus

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* Earth Physics Branch contribution no. 844.

our attention on certain problems that arise from the use of multicomponent analysis, from the choice of reference plane, from the possibility of geologically unrecognized local rotations, and from the uncertainties of determining ages of magnetization at a time when the palaeogeography of the earth may have been evolving rapidly.

A decade ago palaeopoles from Precambrian rocks of the Laurentian Shield were few and reasonably well-spaced in time, although there were often very long intervals between ages assigned to successive poles. There was some bunching, for example, of palaeopoles from Keweenawan rocks upon which early efforts had been concentrated (Du Bois, 1962; Palmer, 1970), but up to about 1970 the spacing of palaeopoles was not *grossly* non-uniform in time. Consequently workers connected them by continuous paths of apparent polar wandering (APW) (Spall, 1971; Irving and Park, 1972). As more results appeared the palaeopoles tended to become grouped in time and in position. Within each group the palaeopoles can usually be linked into an APW path forming curvilinear tracks or loops. But as the groupings become more evident and ages better defined, it is becoming more difficult, and in some instances no longer reasonable, to connect the groups together into a continuous APW path. The absence of dated intergroup palaeopoles arises either because palaeomagnetic observations have not been obtained from rocks of intermediate age or because suitable rocks have not been found.

Palaeomagnetic results are commonly displayed as palaeopole positions (Creer et al., 1954). The observations are reduced thereby to a common frame, allowing comparisons to be made. Necessary as palaeopoles are for analysis, they themselves have no tectonic significance. It is only the values of palaeolatitude and rotation that can be calculated from palaeopoles that have tectonic significance.

Prior to the mid-1970's, the objective in palaeomagnetic work was generally to isolate only the most stable magnetization. It was usual to regard this magnetization as being a record of the geomagnetic field at the time the rock was formed, this assumption being supported with varying degrees of confidence by field and laboratory evidence. Magnetizations of low stability removed during demagnetization were not used. However, it has long been evident that stable secondary magnetizations could partially or completely overprint an earlier magnetic record (see for example, Briden, 1965), and there is the possibility that secondary magnetizations could be more stable than primary ones (Zijderveld, 1975, p. 35). The complexity, brought about by multicomponent analysis, has been further extended by the increased sensitivity of magnetometers which has brought with it the possibility of examining weak magnetizations that could not previously be observed. This expansion of information has resolved some questions but has raised others that are unanswerable at present. We refer in particular to the difficulty of dating all the magnetizations present, especially overprints. In the early work on Phanerozoic rocks it was possible to relate overprints to recognizable geological events of known age. In Precambrian studies the dating of overprints is much more difficult because the timing of processes and events is more obscure.

The correct reference plane is the palaeohorizontal when the magnetization was acquired. The magnetization directions can then be resolved into the geomagnetic coordinates D, I and the palaeopoles calculated (see for instance, McElhinny, 1973). The palaeohorizontal can be readily obtained for the primary magnetization of sedimentary rocks, vertically sided intrusions and layered volcanic sequences, but in massive intrusions and metamorphic complexes it is less readily estimated. In all tilted rocks the problem of determining the correct reference plane for overprints is very difficult, because little can generally be determined from internal evidence alone about the tilt angle of rocks at the time overprints were acquired. The uncertainty in determining the palaeohorizontal for secondary magnetizations and in dating them means they are frequently difficult or impossible to interpret.

Therefore we would characterize much of the data from overprints as "soft" data. The "hard" data are those derived from stable magnetizations that can, from field tests and consistency arguments, be related to the time of formation.

The procedure of assembling palaeopoles from what is now a continuous continental block is complicated by the possibility that sub-blocks may have undergone rotations about local vertical axes or displacements in a palaeolatitudinal sense relative to the whole. Such effects would cause palaeopoles to differ from the APW path. Because of chronological and structural difficulties in Precambrian work, such departures might not be easily recognized. In Fig. 23-1 palaeopoles obtained from the Mesozoic and Cenozoic rocks of the western Cordillera are compared with the reference APW path obtained from cratonic North America which remained a rigid entity during this time. The majority of Cordilleran palaeopoles falls to the right of the path. These are referred to as right-handed palaeopoles and their angular displacement from the reference path provides a measure of the clockwise rotation about local vertical axes that the various small blocks have undergone relative to cratonic North America (Irving, 1964, p. 250; Simpson and Cox, 1977). A smaller number falls on the other side of the reference path. These are infrared to as far-sided palaeopoles and indicate the northward movement of small blocks relative to North America by distances varying from 1000 to 5000 km (Irving and Yole, 1972; Beck and Noson, 1972). In detail these two broad categories merge into one another, — all far-sided palaeopoles show rotations, and the right-handed poles are usually far-sided to varying degrees. Evidently rotations about vertical axes and translation in a palaeolatitudinal sense relative to North America are common in the western Cordillera. These aberrant palaeopoles reflect the interaction between the North American craton and oceanic plates to the west, the

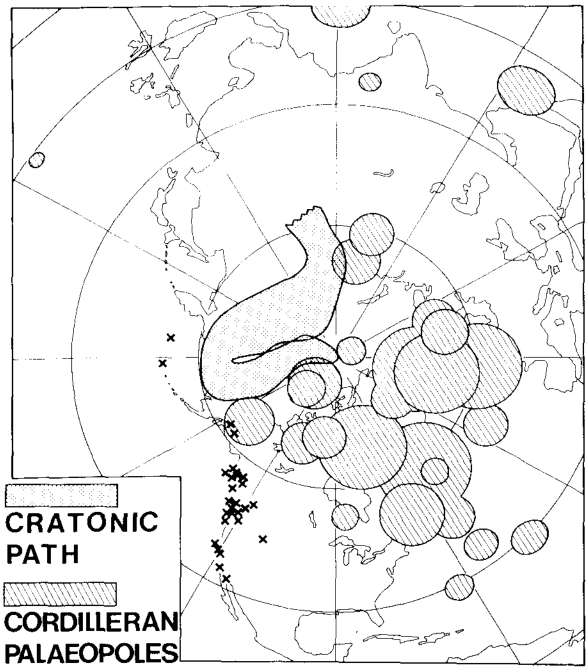

Fig. 23-1. Aberrant palaeopoles from the western Cordillera of North America (crosses are localities) compared with the APW path for the stable craton of North America. Compiles from Irving (1979).

northward displacements and clockwise rotations being produced by the generally northward movement of the oceanic plates relative to North America. One thing is clear from Fig. 23-1, that it would be pointless to attempt to draw an APW path through the aberrant palaeopoles. Such a path would be meaningless because the various sub-blocks in the western Cordillera have been moved relative to the craton and to each other by varying amounts. Given a group of Precambrian palaeopoles for which the ages are very much less well-defined one could draw a serpentine APW path through the cluster but, on the basis of Cordilleran experience, such a path would be meaningless tectonically. It might therefore be expected that results from Precambrian foldbelts would be uninterpretable. There are several reasons why this is not necessarily so.

The first reason is that a substantial part of the Precambrian record is from rocks formed on older already stabilized cratons, so that segments of the reference path should be obtainable. The second reason is that the predominant effects in the western Cordillera are rotations, and large palaeolatitudinal displacements are less common. This would lead one to expect that, although an attempt to draw APW paths may be fruitless, the analysis of palaeolatitudes should reveal generalized trends. This is born out in

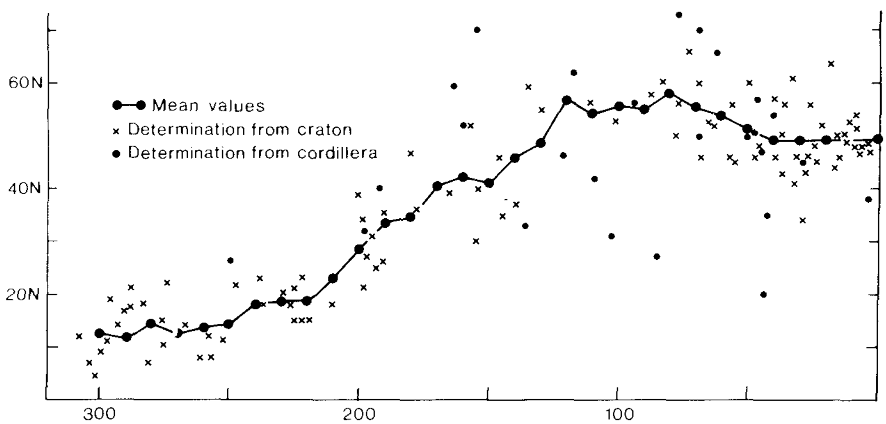

Fig. 23-2. Palaeolatitudes for post mid-Carboniferous times determined for 49°N, 123°W (Vancouver) by extrapolation from the North American craton and from displaced terrains of the western Cordillera. The mean values are calculated from the mean poles given by Irving (1979, table 1).

Fig. 23-2. Palaeolatitudes calculated from most but not all aberrant palaeopoles now fall within the distribution of results from cratonic North America. Rotational effects are suppressed and the main palaeolatitude trend, although fuzzier, is still clear. Finally, the record of the potential complexities of palaeomagnetic results from within older foldbelts are likely to be partially obliterated by overprinting, only later events being recorded. The palaeomagnetic record of the early part of an orogenic cycle may be difficult or impossible to retrieve. Thus, although overprinting and the absence of suitable unaltered rock types may diminish the palaeomagnetic record of the earlier parts of orogenic cycles, the rotational and latitudinal motions of the later parts should be observable.

GEOLOGICAL SETTING

Laurentia comprises the Precambrian shields of North America (except for the Avalon Platform of Newfoundland), Greenland, and the Lewisian Platform of northwest Scotland (Fig. 23-3). Laurentia as we know it today was formed during four stages and possibly part of a fifth. Each stage comprised a series of events by which recognizable terrains were formed and stabilized. The general scheme is set out in Table 23-1.

The first stage, which culminated in the Kenoran Orogeny (about 2700 Ma ago), produced the Slave, the Superior, and the North Atlantic Structural Provinces, the Beartooth uplift in Wyoming and Montana, the Archaean terrains of the Minnesota River valley, and several Archaean blocks in the Churchill Province. The second stage was initiated by rifting and depositional

TABLE 23-1

Framework for Precambrian events

	Late Proterozoic, 570 to ~ 950 Ma ago
5	Pan-African Orogeny, ~ 600 Ma
5	Franklin intrusions, 700 Ma
	(gap of about 200 Ma in palaeomagnetic record)
	Middle Proterozoic, ~ 950 to ~ 1700 Ma ago
4	Grenvillian Orogeny, ~ 1050 Ma
4	Keweenawan, ~ 1100 Ma
4	Mackenzie intrusions, ~ 1200 Ma
4	Seal Lake Group, ~ 1300 Ma
4	Sibley, ~ 1300 Ma
4	Beltian, ~ 1350 Ma
3	Elsonian Intrusives, ~ 1450 Ma (gap of about 200 Ma in palaeomagnetic record)
	(gap of about 200 Ma in palaeomagnetic record)
	Early Proterozoic, ~ 1700 to ~ 2500 Ma ago
2	Hudsonian Orogeny, ~ 1800 Ma
2	Animikie, ~ 2000 Ma
2	Pre-Coronation intrusions, ~ 2100 Ma
2	Nipissing diabase, ~ 2100 Ma
2	Huronian, ~
	(below this tł (below this the palaeomagnetic record becomes very poor)
	Archaean, older than ~ 2500 Ma
	Matachewan dykes, ~ 2600 Ma
1	Kenoran Orogeny, ~ 2650 Ma
	Archaean greenstone belts, ~ 2700 Ma

Approximate ages in millions of years before Present given. The numbers on the left denote stages 1 to 5 as described in the text.

events and terminated by phases of deformation, metamorphism and granitic intrusion, generally referred to as the Hudsonian Orogeny. The Churchill, Bear and Southern Provinces were created in approximately their present form during this second stage. The succeeding Elsonian event seems to have been dominantly intrusive in nature and was followed by rifting and deposition of the Belt and Sibley Groups. The fourth stage began with rifting events, and was terminated by phases of deformation and metamorphism, referred to as the Grenvillian Orogeny. Finally, towards the end of the Precambrian, widespread rifts and associated intrusions were formed in Laurentia which could represent the commencement of a fifth stage that was terminated elsewhere by the Pan-African Orogeny.

LATE PROTEROZOIC (~ 570 to ~ 970 Ma)

There are few palaeomagnetic results from the late Proterozoic. Stratigraphic sequences are few and generally contain rocks that have not, up to now, been considered suitable for palaeomagnetic work. However, a very widespread set of gabbroic intrusions dated between ~ 650 and ~ 700 Ma have been studied following the work of Fahrig et al. (1971). They extend

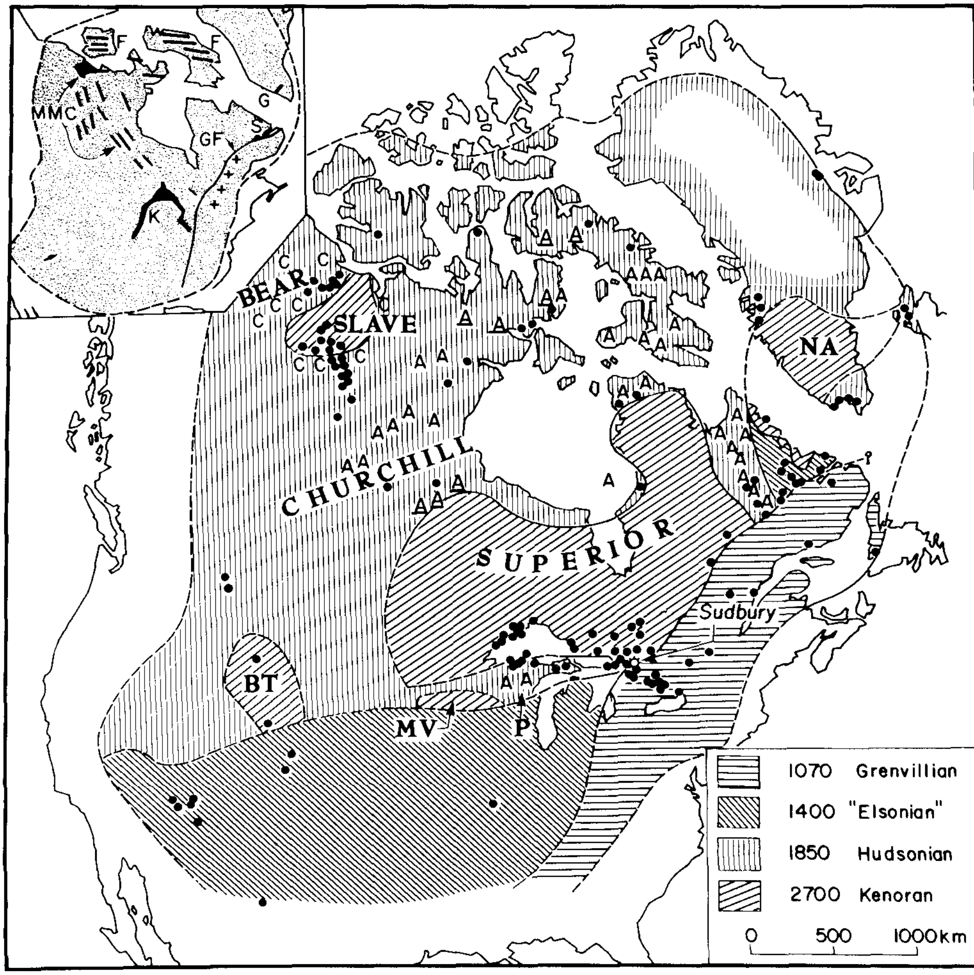

Fig. 23-3. Structural provinces of Laurentia. The more important palaeomagnetic sampling localities are marked by dots. The approximate times at which the four major subdivisions of Precambrian terrain were stabilized are from Stockwell (1973). *P* is the Penokean fold belt. *BT* is the Beartooth uplift, *MV* the Archaean terrains of the Minnesota Valley and *NA* the North Atlantic craton. Early Proterozoic sedimentary and volcanic basins are denoted by the letter *A*. The *A*'s are underlined if extensive ultrabasic rocks are present. Greenland and the Lewisian basement of northwest Scotland are rotated back to their supposed positions at the end of the Precambrian. Piper (1979) has questioned the correctness of this reconstruction of the position of the Lewisian platform. In the inset the pre-Grenville rifts are shown: *F* = Franklin intrusions; *K* = Keweenawan; *G* = Gardar intrusives; *MMC* = Muskox intrusion, Mackenzie diabase, and Coppermine lavas; *S* = Seal Group volcanics; crosses are alkaline complexes near to or just south of the Grenville Front. Modified after Irving and McGlynn (1976b).

across northern Canada from Victoria Island to Baffin Island, and may extend southward sporadically into the Superior and Grenville Structural Provinces (Murthy, 1971; Park, 1974). These bodies yield consistent palaeopoles which are situated in the east-central Pacific. The identification of these bodies over a wide area of the Canadian Shield has been made possible largely on the basis of the palaeomagnetic evidence, and they indicate the occurrence at this time of a widespread extensional (rifting) event.

MIDDLE PROTEROZOIC (~ 950 to ~ 1450 Ma)

Older results

Palaeopoles for the middle Proterozoic are plotted in Fig. 23-4. The APW path is sketched, annotated with some of the more important results, and calibrated in Fig. 23-6.

The Elsonian plutons of Labrador, Ontario, Missouri and Colorado (*CRC*, *MIP*, *HL*, *MK*, *FA*, *NC*, *SHG*) are well-grouped and their probable ages are in the range ~ 1400 to ~ 1500 Ma. There are small differences that could have arisen from small local rotations which may or may not be associated with movements along the Grenville Front (Irving et al., 1977). The somewhat younger Beltian rocks from Alberta and Montana (*BE1*, *BE2*) and the Sibley Group of Ontario (*SIB*) have yielded palaeopoles situated to the north. There is therefore no evidence of gross dismemberment since about 1400 Ma of those parts of the Laurentian Shield spanned by these sampling localities. This conclusion does not apply to the Grenville Structural Province because there are no results of this age from there. It is possible also that it does not apply to the northwestern Canadian Shield, because the palaeopole for the Western Channel diabase (*WC*) from the Bear Province falls about 30° east of the palaeopoles in the range ~ 1300 to ~ 1450 Ma (Fig. 23-4). This diabase has yielded an Rb-Sr* isochron age of 1392 ± 48 Ma (Wanless and Loveridge, 1978), and a clockwise rotation of 30° would bring the pole into agreement with the main cluster for this age range from the eastern shield. Relative rotation, in post ~ 1400 Ma time, of part of the northwestern shield relative to the eastern shield is not precluded by the palaeomagnetic data.

Passing forward in time, results are available from the Seal Lake Group, a rift sequence of sedimentary red beds (*SER*) and igneous rocks (*SEV*) dated by Rb-Sr isochron at 1321 ± 92 Ma and situated just north of the Grenville Front (Roy and Fahrig, 1973; Park and Roy, 1979). The red beds are generally younger than the igneous rocks, but a few samples are from within the igneous sequence. The inclinations of the two groups are not significantly

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* Rb-Sr ages quoted in this paper have been calculated using decay constant of $1.42 \times 10^{-11} a^{-1}$.

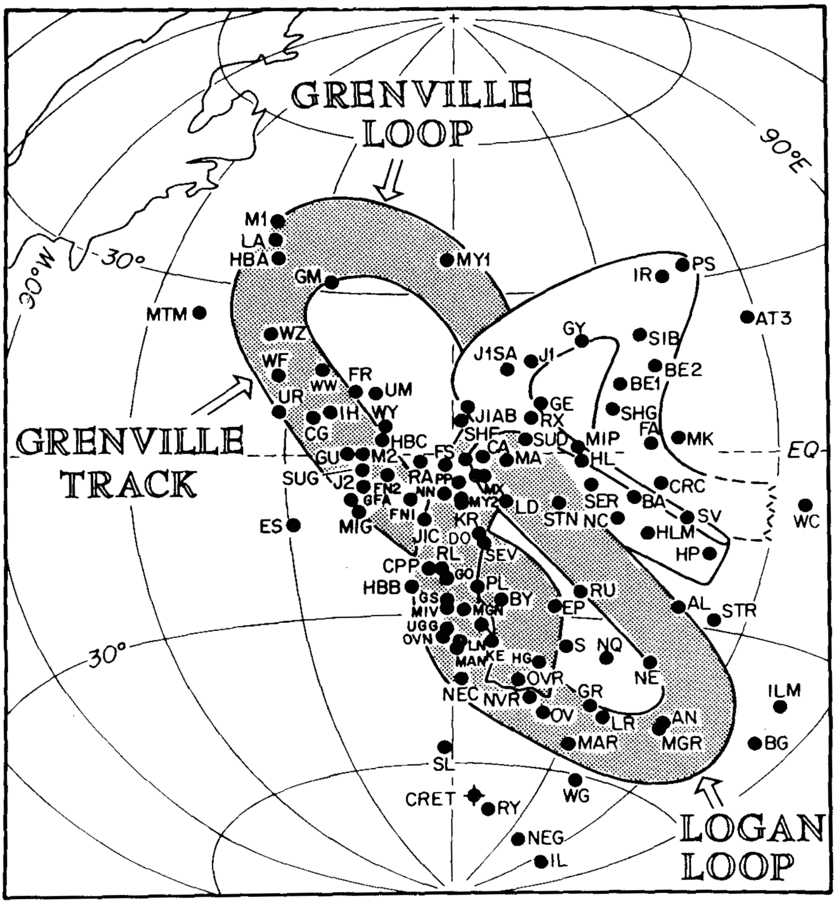

Fig. 23-4. Suggested APW path for middle Proterozoic times. The path is calibrated in Fig. 23-6. Modified from Irving (1979). Individual poles are keyed by letters to a list in the appendix. Poles *CP* (between *FS* and *CA*), *TO3* (near *MX*) and *NVN* (near *MGN*) are unlabelled because of space restrictions. The unshaded portion of the path is less secure. *CRET* is the position of the Cretaceous pole (~ 100 Ma) for North America.

different but the declinations differ (*SEV* 291 ± 7°; *SER* 275 ± 6°), indicating that relative rotations about local vertical axes could have occurred. Such rotations are more likely to be caused by syn-rifting processes than by the formation of the much younger Grenville Front (about 1000 Ma ago), because the igneous and red bed sampling sites are spatially intermingled. However, the igneous and red bed sampling sites are also to a limited extent stratigraphically intermingled (Roy and Fahrig, 1973; Fig. 23-1) so that such a model is obviously too simple. Nevertheless, the data do indicate relative rotation about local vertical axes.

The APW path for the early–middle Proterozoic of Fig. 23-4 is left unshaded, because it is speculative. The first workers to attempt to construct APW paths for this interval of time drew a simple loop (Spall, 1971) but complications are now apparent for reasons that are not clear (Murthy, 1978). The APW path for Laurentia as a whole could be truly more complicated, and relative rotations about local vertical axes could be in part responsible.

The next well-established results are from the Mackenzie igneous units dated at about 1200 to 1250 Ma (*CP*, *KR*, *LD*, *MA*, *MX*, *SUD* of Fig. 23-4). Intrusive and extrusive volcanic and sedimentary rocks are represented which, with one local exception, have very similar palaeopoles. The exception is from a section of lavas within the Coppermine sequence. These lavas have the low inclinations typical of Mackenzie-age rocks but have aberrant declinations that indicate a 20° counter-clockwise rotation of a restricted fault-defined block. The rift faulting, which was presumably responsible for the rotation, occurred soon after the extrusion of the lavas and deposition of overlying red beds (Baragar and Robertson, 1973).

Logan Loop

Following the Mackenzie palaeopoles there is an arcuate polar sequence. This is the Logan Loop, first outlined by Du Bois (1962), and since discussed by many workers, notably Robertson and Fahrig (1971). The descending limb of the loop is defined, not very firmly, by palaeopoles from the Lower Keweenawan (*STN*, *STR*), but its ascending limb is well-established from the stratigraphic sequence of Upper Keweenawan sediments and lavas and associated intrusions (*LN*, *LR*, *KE*, *PL*, *CPP*, *FN1*, *FN2*). The Logan Loop, plotted on the opposite hemisphere, is shown in Fig. 23-5 together with an important set of data, as yet incompletely published, from the Grand Canyon Supergroup (Elston and Grommé, 1974). The Grand Canyon rocks have yielded 14 poles in stratigraphic sequence that appear to generally confirm the existence of the loop. They do not extend as far north so that the depth of the loop remains uncertain. Some authors (Morris et al., 1979; Baer, 1979) have recently suggested that the younger poles should be connected backwards to the antipole of the Mackenzie rocks, replacing the tight Logan Loop with a broad open polar swath. We see little evidence for this alternative.

A feature of the polar sequence from the Grand Canyon Supergroup is that the palaeopole from the stratigraphically youngest rocks is placed southward beyond the end of the Logan Loop (15, Fig. 23-5) and indicates the extension of the loop southward (or northward in the antipolar configuration of Fig. 23-4). Palaeopoles (*J1C*, *J1AB*, *J1SA*) from the Jacobsville Formation of the Lake Superior region are considered to form a time sequence overlapping with the younger poles from the Upper Keweenawan

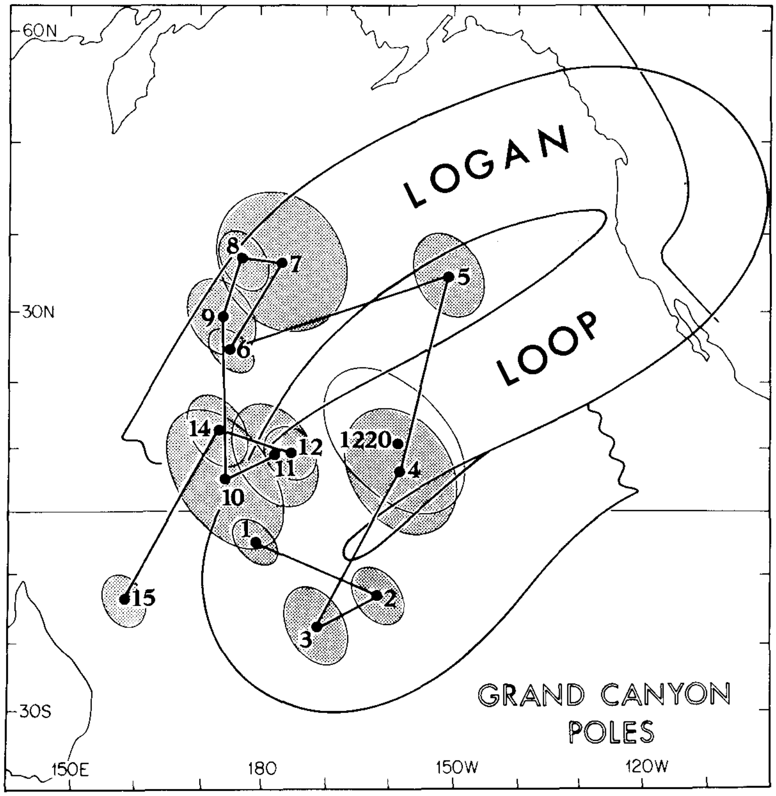

Fig. 23-5. Palaeopoles for the Grand Canyon Supergroup (Elston and Grommé, 1974) compared with the Logan Loop, plotted as antipoles of Fig. 23-4. The poles are numbered, oldest first, as follows: *1* = Bass limestone (lower); *2* = Bass limestone (middle and upper); *3* = Hakatai Shale (lower and middle); *4* = Hakatai Shale (middle and upper), *5* = Shinumo Quartzite (middle); *6* = Dox Sandstone (upper lower); *7* = Dox Sandstone (lower middle); *8* = Dox Sandstone (middle); *9* = Dox Sandstone (upper middle); *10* = Dox Sandstone (upper); *11* = Cardenas lavas (flows); *12* = Cardenas lavas (intercalated sandstones); *14* = Nankoweap Formation (ferruginous member); *15* = Nankoweap Formation (upper member). The Shinumo quartzite result is based on only 9 samples and therefore not firmly based. Modified from Irving and McGlynn (1976b).

(Roy and Robertson, 1978). The Jacobsville palaeopoles also provide evidence that the Logan Loop does indeed continue northward using the configuration of Fig. 23-4. It is important to note however, that this argument depends only on the apparent continuity of the palaeomagnetic results. Neither relative age relationships of the sections studied within the Jacobsville nor the relationship of the Jacobsville to the Keweenawan are known, because of the absence of stratigraphic continuity.

Grenville Loop

Grenville palaeopoles, dated at about 980 Ma, together with younger overprints define the Grenville track. These results are from high-grade metamorphic rocks. They may be connected backwards with the Logan Loop to form a second loop, the Grenville Loop, which can be continued south as far as the pole for the Stoer Group (S) which is dated at 970 ± 24 Ma. Piper (1979) suggests that the Lewisian block upon which the Stoer Group rests has been rotated relative to Laurentia and this would invalidate this continuation. A consensus is now emerging favouring the construction of a Grenville Loop continuous with the Logan Loop. This has not been achieved without controversy centred on the very difficult problem of dating the magnetization of high-grade metamorphic rocks.

When Grenvillian palaeopoles were first obtained they were considered to be somewhat younger than palaeopoles from mid-Keweenawan rocks to which they were linked sequentially (Du Bois, 1962). This implied that the Grenville Province has been part of Laurentia since mid-Keweenawan time (about 1150 Ma ago). Palaeopoles *GM*, *HBA*, *M1*, *MTM*, *WF* and *WW*, situated in intermediate latitudes, are typical examples. Arguments were advanced, based on erroneous ideas about the relationship between radiometric ages and magnetizations observed in metamorphic rocks, to suggest that these palaeopoles were roughly the same age (about 1150 Ma) as those from mid-Keweenawan rocks (Palmer and Carmichael, 1973; Irving et al., 1972). The latter palaeopoles are situated some 60° to the southeast. Such arguments lead to the proposal that the southern two-thirds of the Grenville Structural Province has been displaced by several thousand kilometres relative to interior Laurentia at this time. The idea was not popular, but it had the merit of being amenable to disproof by dating Grenville-derived palaeopoles. The magnetizations of high-grade metamorphic rocks can be dated by $^{39}Ar/^{40}Ar$ release spectra, and the age of Grenvillian poles is very probably between ~ 1000 and ~ 950 Ma (Berger et al., 1978; Dallmeyer and Sutter, 1980), some 100 Ma younger than anticipated. The intermediate latitude Grenville palaeopoles therefore are not evidence for a displaced Grenville Province because they post-date all major tectonic activity.

Following the intermediate latitude Grenvillian palaeopoles there is a sequence extending into low latitudes. These have been widely observed as overprints in the Grenville Province, in older rocks up to a distance of 10 to 20 km north of the Grenville Front, and as the sole magnetization in rocks just south of the Front (Palmer and Carmichael, 1973; Ueno and Irving, 1976; Palmer et al., 1977). They have been produced from magnetizations acquired during late-stage cooling and uplift of the Grenville Province. It is noteworthy that the change in direction of the APW path (or hairpin) at ~ 1000 Ma ago corresponds in time to the culmination of the Grenville Orogeny.

The age of the end of the Grenville Loop is probably about 950 Ma and in the interval between this and the Franklin intrusions (about 650 Ma ago) there is one dated pole from the Torridon Group (779 ± 24 Ma). Morris and Roy (1977) and Roy and Robertson (1978) have constructed a polar track based on overprints (*RY*, etc.), but we shall not adopt their reconstruction because the ages of the overprints are unknown. For example one of those overprints, *RY*, is very close in the Cretaceous palaeopole from North America (CRET of Fig. 23-4) and since it has been obtained from rocks deformed during the Laramide Orogeny the overprint could be of that age.

Summary

The palaeopoles for the later half of the middle Proterozoic (approximately 950 to 1200 Ma ago) can be formed into two loops, the Grenville and Logan Loops. These loops reflect large and rapid changes in palaeolatitude of Laurentia during the later half of the middle Proterozoic. The distribution of palaeopoles for the earlier half of the middle Proterozoic is complex and the paths drawn in Figs. 23-4 and 23-6 are schematic only.

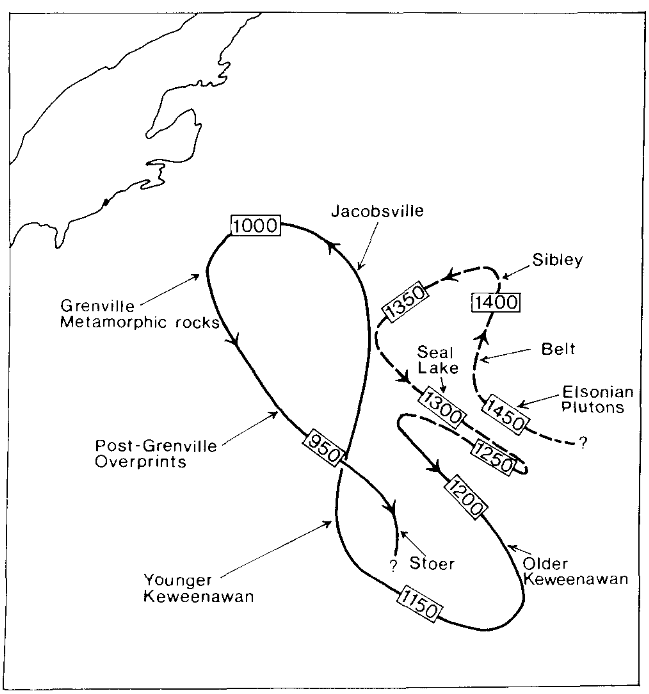

Fig. 23-6. Summary and suggested calibration of the APW path for the middle Proterozoic compiled from Fig. 23-4. The less secure part is dashed. Important palaeopoles are marked.

Some of the complexity could be unreal and caused by local rotations. There is no reliable forward continuation of the path of Fig. 23-4 to the late Proterozoic, so its polarity is not known. Nor can a backward connection to older poles be made (see below).

EARLY PROTEROZOIC AND LATEST ARCHAEAN (~ 1700 to ~ 2600 Ma)

Numerous palaeopoles are available from early Proterozoic rocks in the age range ~ 2200 to ~ 1700 Ma. There is a younger group, comprising a sequence known as the Coronation Loop, derived mainly from the Coronation Geosyncline, and an older group derived mainly from the Slave and Superior Structural Provinces. A few results are also available from the oldest Proterozoic.

Coronation Loop

The Coronation Loop begins at about 1850 Ma with palaeopoles obtained from the Dubawnt and Kahochella Groups (*DU*, *UG*, *SA*), the latter being the oldest rocks studied from the Great Slave Supergroup (Fig. 23-7). Poles from the upper parts of the Great Slave Supergroup fall farther to the south (*TA*, *DP*, *PA*, *PPS*).

The ascending limb of the Coronation Loop is characterized by palaeopoles from the Richmond Gulf (*RG*, *RGM*), from post-tectonic molasse of the Coronation Geosyncline (*ET*), from some post-Hudsonian dykes (*SY*) and by a group of palaeopoles focussed on middle America and extending northwards to Hudson Bay (*KD*, *CH*, *FFA*, *IT*, *KA*, *KC*, *KD1*, *KD2*, *KL*, *KM1*, *KM2*, *LG*, *MDB*, *NO*, *PC*, *SG*, *SS*). The latter group, which are mostly overprints, probably reflects the general uplift of Laurentia at about 1700 Ma ago, following the Hudsonian Orogeny and recorded by K-Ar ages. *SG* could be the youngest of this group and is estimated to be about 1640 Ma old (Beckman and Mitchell, 1976). Afterwards there is a hiatus of about 200 Ma, so that the forward continuation cannot be made, and the polarity of the APW path of Fig. 23-7 is unknown.

The palaeopoles from medium- to high-grade metamorphic rocks from West Greenland all fall on the ascending limb of the Coronation Loop and appear to be caused by magnetizations acquired by cooling during post-orogenic uplift. They therefore tell us nothing about the relative motions between the North Atlantic Craton (Greenland and coastal Labrador) and the Superior Structural Province during the Hudsonian Orogeny; they do not tell us whether the Labrador Trough was a wide ocean or a narrow rift during early Proterozoic time.

That the Coronation Loop exists has been recognized for some time. There is a trend first away from and then towards Laurentia representing a variation in palaeolatitude. The results, however, allow considerable sideways

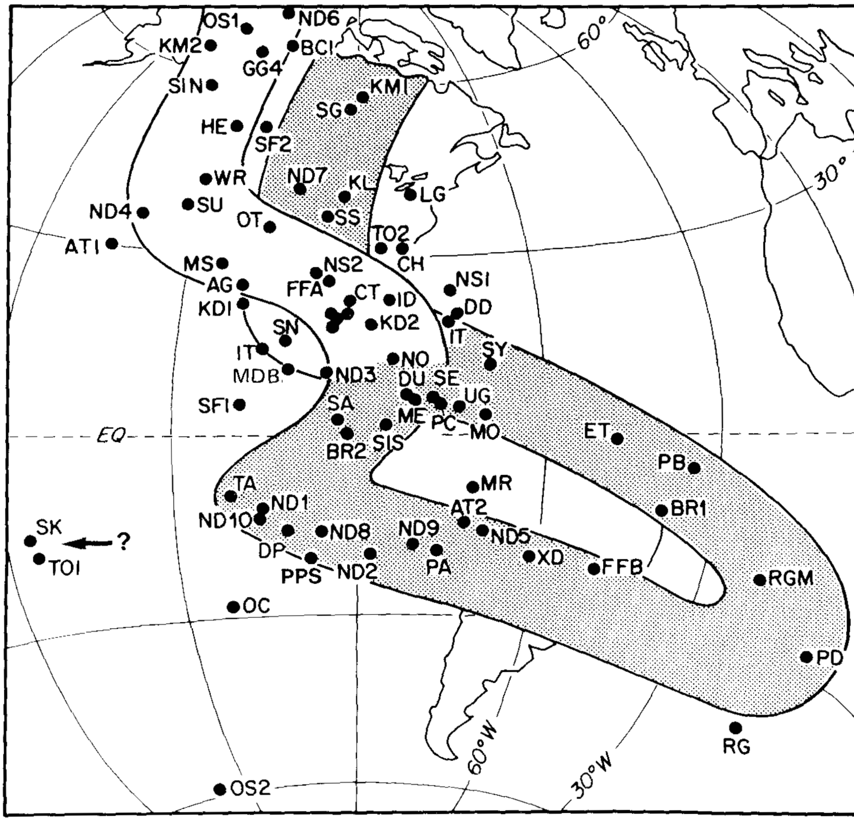

Fig. 23-7. Suggested APW path for the middle Proterozoic. Modified from Irving (1979). Three poles, *GT*, *KA* and *KC* near *CT* are unlabelled because of space limitations. Unshaded path (old Track 5) is no longer considered satisfactory (see text). The arrow denotes the suggested tectonic rotation of the Stark and Tochatwi palaeopoles.

motion of the APW path (caused by declination variations) superimposed on the general to-and-fro motion (caused by variations in inclination), and it is uncertain to what extent local rotations about vertical axes have been responsible for the former. In this way the Coronation Loop differs from the Logan and Grenville Loops which mainly reflect changes in palaeolatitude. The possibility that local rotations may have contributed to the sideways spread of palaeopoles in the Coronation Loop is now considered.

The first example is from the Dubawnt Group (*DU*) and the lower part of the Great Slave Supergroup (*SA*, *UG*). These three sequences are probably not much separated in time and yield statistically identical inclinations but differing declinations so that the palaeopoles have a longitudinal spread of about 25°. Such is the signature of local rotations. One of the results (*UG*) is from a nappe in the Great Slave Lake region (Hoffman et al., 1977) and, although this nappe is not internally deformed and does not appear to have travelled very far, it could have been rotated by the 25° relative to the autochthonous rocks from which palaeopole *SA* was obtained.

A group of palaeopoles from the upper part of the Great Slave Supergroup may provide evidence of even larger local rotations. These are, in stratigraphic order from oldest to youngest, *DP*, *SK*, *TO1* and *PA* (Fig. 23-7). *SK* and *TO1* from the Stark and Tochatwi formations are displaced over 30° to the west of the main group. Both have been obtained from the same limited area in which the beds are tightly folded, with near vertical dips locally overturned, but they are not metamorphosed. Evans and Bingham (1976), authors of the *SK* and *TO1* palaeopoles, were uncertain about the meaning of their result and suggested that they "should not be incorporated into polar wander paths until verified by results from tectonically 'normal' areas" because of the uncertainty in structural restorations. Many authors, including ourselves (Irving and McGlynn, 1976b), chose not to accept this good advice. Subsequently, further evidence has become available adding weight to the opinion of Evans and Bingham. Palaeopoles from two rock units (*DP* and *PA*), stratigraphically beneath and above the Stark and Tochatwi, are located 45° and 65° to the east. Furthermore, palaeopole *TA* from the Takyiuk Formation is also situated well to the east. The Takyiuk is considered by Hoffman (1973) to be the stratigraphic equivalent of the Tochatwi (*TO1*) and it is autochthonous on Archaean basement 300 km to the north. Bingham and Evans (1976) noted that the strikes in the sampling area of the Stark and Tochatwi are systematically 60° clockwise of those in rocks of the Great Slave Supergroup as a whole. If the *SK* and *TO1* poles are rotated 60° anticlockwise they fall between *DP* and *PA* in the correct position required stratigraphically. We regard this as good evidence favouring the suggestion of Bingham and Evans (1976) that the westerly displacement of the *ST* and *TO1* palaeopoles has been caused by local rotation of the sampling area. Gough et al. (1977) have argued against such a rotation from studies of the anisotropy of susceptibility of the sedimentary rocks. Irving and McGlynn (1979) have criticized their argument, firstly on the grounds that uniformity of directions of maximum susceptibility is observed at localities in which the remanent magnetizations are demonstrably secondary, and secondly the evidence from remanent magnetization for rotation occurs in different sequences from those in which the uniformity in directions of anisotropy of susceptibility is observed. The situation, however, is clearly very complicated, because although the rotation of *ST* and *TO1* invoked by Bingham and Evans (1970) and supported by our own work (McGlynn and Irving, 1978; Irving and McGlynn, 1979) brings these palaeopoles into conformity with other results from the Great Slave Lake, it does not bring them into agreement with results from the Takyuak Formation (*TA*) which rests allochthonously on Archaean basement to the north. Assuming that Hoffman's correlation of the Takyuak and Tochatwi rocks is correct, then the disagreement can be explained by supposing that the Great Slave

Supergroup of the Great Slave Lake region has been rotated about 20° anticlockwise relative to the Slave Province. Such rotations could be caused by the emplacement of nappes that occur within the Great Slave basin (Hoffman et al. 1977), or by movement along strike-slip faults which cut the Great Slave Supergroup.

We conclude that because of the possibility of unresolved rotations of local blocks the actual form of the Coronation Loop is uncertain although its existence seems reasonably well-established. It is sketched and calibrated in Fig. 23-8.

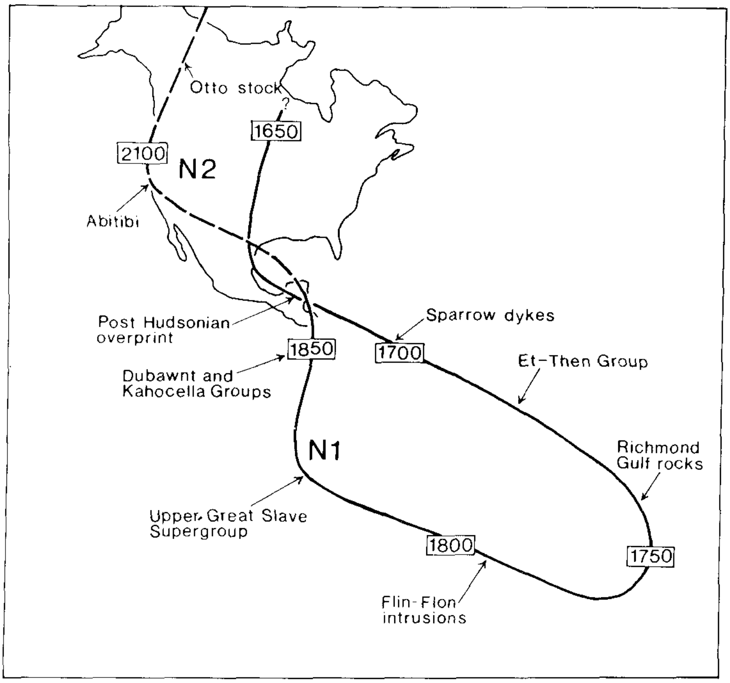

Fig. 23-8. Summary and suggested calibration of the APW path for the middle Proterozoic compiled from Fig. 23-7. Old Track 5 is dashed. Important palaeopoles are marked. *N1* and *N2* are the positions for the two main types of Nipissing polar groups.

Slave—Superior comparison

The unshaded part of the APW path of Fig. 23-7 has often been referred to as Track 5. It is characterized by palaeopoles from intrusions in both eastern and western Laurentia that are dated at about 2200 to 2000 Ma (*AT1*, *AT2*, *BC1*, *ID*, *MS*, *ND1* to *10*, *OT*), by palaeopoles from Animikee rocks of the Superior region (*HE*, *GT*) and the Labrador Trough (*CT*, *ME*, *SF1*, *SF2*, *SN*). When early and middle Proterozoic palaeopoles first became available from the Slave Structural Province they raised the possibility of motion relative to the Superior Structural Province (McGlynn and Irving, 1975) but the results were not definitive, and as more were obtained, it became possible to reconcile all of them to a simple single path, Track 5 (Pullaiah and Irving, 1975; Roy and Lapointe 1976), which could be connected southward with the Coronation Loop. Since Track 5 contained results from both the Slave and Superior Province it has been taken as evidence that there was little relative movement in a latitudinal sense (longitudinal motions were not excluded) between these Archaean blocks during the interval ~ 2100 to ~ 1900 Ma ago, just prior to the Hudsonian Orogeny (Irving and McGlynn, 1976b). Large movements would be expected at this time if that orogeny was indeed caused by processes comparable to modern plate tectonics. Burke et al. (1976) and Cavanaugh and Seyfert (1977) have argued, as McGlynn and Irving (1975) originally did, that separate APW paths do indeed exist for the Slave and Superior Structural Provinces, and hence that such motions in fact occurred (Fig. 22-9)

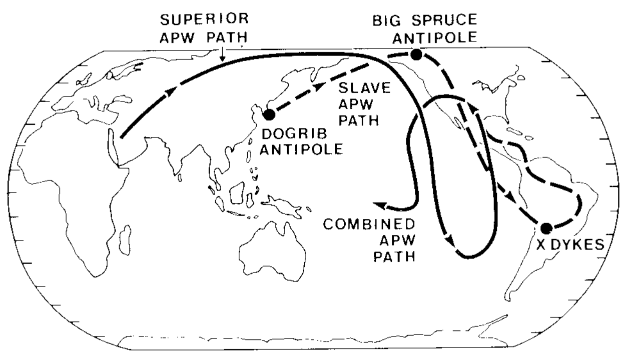

Fig. 23-9. Separate APW paths for Superior and Slave Structural Provinces constructed by Seyfert and Cavanaugh (1978).

We now consider these contradictory interpretations. In Fig. 23-10 the palaeopoles from the Slave and Superior Provinces are plotted. The palaeopoles from the Otto Stock and Abitibi dykes both from Superior differ by 35° even though their ages (Rb-Sr isochron ages of 2114 ± 80 Ma; Bell and Blenkinsop, 1976, and 2100 ± 68 Ma; Gates and Hurley, 1973) do not differ significantly. Both results are supported by contact tests and both are from the same general area in which there is no geological evidence of local rotations. The stock and the dykes have vertical contacts indicating that subsequent tilting has been small or negligible.

Numerous studies of the quasi-contemporaneous Nipissing diabase (Rb-Sr isochron age of 2100 ± 50 Ma; Van Schmus, 1965) and have yielded a wide distribution of poles (*ND1* to *N11*, Fig. 23-10). The metamorphic grade of the Nipissing diabase varies from sub-greenschist to amphibolite. It would

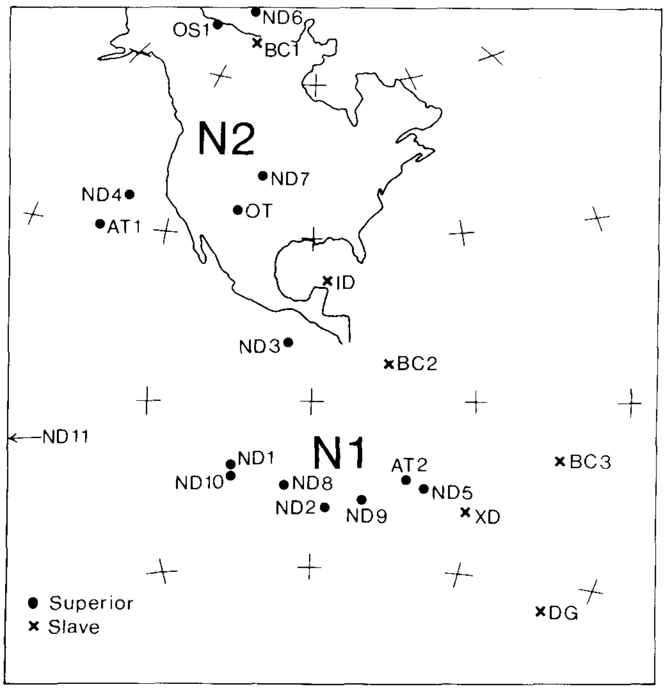

Fig. 23-10. Palaeopoles from rocks in the age range ~ 2000 to ~ 2200 Ma. Poles are listed in the appendix. *N1* and *N2* are the positions for the two main groups of Nipissing palaeopoles.

appear that generally low-grade rocks have been studied, although this is not always clear.

The Nipissing palaeopoles have been considered to fall into a diffuse northerly group (referred to as *N2*) and a tighter southerly group (*N1*). There is considerable uncertainty about the relative and absolute ages of these magnetizations. Symons (1967) believed that *N1* was the original magnetization and he described a contact test in support. Later Roy and Lapointe (1976) regarded *N1* as an overprint and *N2* as the original magnetization, basing their argument mainly on contact test for palaeopole *ND7*. Subsequently Morris (1979) has described a further contact test that could be applicable to the *N1* group of palaeopoles supporting Symons' original argument. A further complication is that the geological relationships between the rocks studied magnetically and those studied radiometrically is not always clear. Thus the *ND* palaeopoles of Fig. 23-10 could have been derived from rocks of several different intrusive events. For example, *ND11* is indistinguishable from poles for the Franklin intrusions (approximately 650 Ma old) and from palaeopoles from WNW-trending dykes of the Grenville swarm that are common in this area (Murthy, 1971; Park, 1974).

Hence *ND11*, ostensibly derived from Nipissing diabase, could have been derived from an intrusion that is very much younger than the Nipissing diabase, but is difficult to distinguish from it because of their petrological similarity. It is dangerous, therefore, to use *ND11* to support any APW path, as Roy and Lapointe (1976) and Embleton and Schmidt (1979) have done in order to extend the Coronation Loop eastward, and as Piper (1979) has done using the antipole in order to extend it westward!

The interpretation of results from the Nipissing diabase is further complicated by the uncertainty in the application of tilt corrections. The Nipissing diabase occurs as concordant or transgressive sills that have been intruded into Huronian sedimentary rocks. According to Card and Pattison (1973) the Nipissing intrusions "were emplaced after early faulting during the later stage of an early major folding of the Huronian sequence, but before secondary deformational events and regional metamorphism". In earlier studies attention was paid to this problem and Symons (1970) showed in some instances that the directions of magnetization were somewhat better grouped before than after corrections were applied for tilting observed in country rocks. In later studies palaeopoles have been calculated on the assumption that the various magnetizations were acquired when the rocks were in their present relative positions. This assumption would be justified if all Nipissing palaeopoles were in good agreement, but they are not. The effect of the absence of tilt correction is not clear, but it is noteworthy that the general strike of the Huronian rocks into which the Nipissing diabase is intruded is approximately east—west so that it is geometrically possible that the absence of proper attitudinal corrections could account for some of the meridional spread of Nipissing palaeopoles.

Clearly the palaeomagnetic results from the Superior Province and its southern border do not provide a clear picture of APW during the interval of time immediately prior to the Hudsonian Orogeny.

The palaeopoles from the Slave Structural Province are aligned roughly along a northwest to southeast trend. *DG*, from the Dogrib dykes, is probably the oldest because it is derived from the oldest body studied and although contact tests cannot be made (the country rock is magnetically unstable) the uniformity and stability of their distinctive magnetization directions suggest that the magnetization could be original. Estimates of the age of the Dogrib dykes fall in the range 2200 to $\sim$ 2600 Ma. *XD* is a palaeopole from an undated and unnamed small group of dykes that are probably of middle Proterozoic age. Palaeopole *ID* from the Indin dykes (2049 ± 86 Ma, Rb-Sr isochron) is supported by a contact test. The Big Spruce Complex (2066 ± 40 Ma, Rb-Sr isochron) has yielded palaeopoles *BC1*, *BC2* and *BC3* attributed to three distinct magnetizations (Irving and McGlynn, 1976a). A fourth magnetization, *BC4*, is probably very much younger. At first we considered *BC1* to be the primary magnetization and *BC2* and *BC3* to be overprints related to the younger Coronation Loop

(Irving and McGlynn, 1976a). *BC1* is statistically indistinguishable from *OS1* (the Otto Stock) and we originally considered this agreement to be good evidence against the idea of large relative motions between the Slave and Superior Structural Provinces since about 2000 Ma ago. However, we are now less certain that this is a sound argument for the following reasons.

The directions of remanent magnetization of the amphibolite from the contact aureole of the Otto Stock are close to the present earth's field, but they are very stable and show excellent square-shouldered demagnetization curves (Fig. 23-11). Other specimens have antiparallel reversed directions.

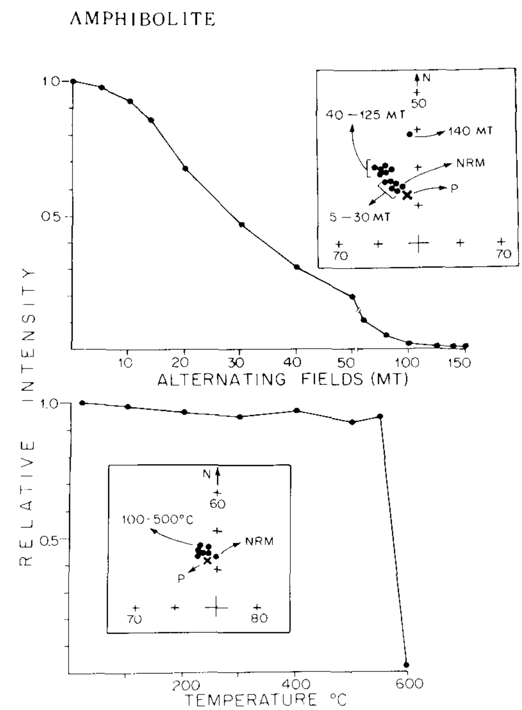

Fig. 23-11. Alternating field and thermal decay curves from amphibolite in contact aureole of the Otto Stock. In the insets the centres of equal-area nets are shown with the directions observed during step-size demagnetization. *P* is the present field. From Pullaiah and Irving (1975).

Similar magnetization directions occur in the late-stage lamprophyre dykes that are part of the Otto Stock. The direction in the dykes, when combined with the directions in the aureole, yield palaeopole *OS1*. The stock has vertical contacts and circular outline, so no attitudinal corrections are required. The magnetization is typical of that carried by single-domain magnetite and was probably acquired during cooling following intrusion of the Otto Stock (Pullaiah and Irving, 1975). The magnetization of the Big Spruce Complex is less simple. Results from a pyroxenite specimen are shown in Fig. 23-12. The pyroxenites have a predominantly (75%) soft magnetization that can be removed in low alternating fields. The remaining magnetization, when analyzed by step-wise thermal demagnetization, has two components. There is a larger *Z* magnetization making up about 20% of the natural remanent magnetization which is characterized by excellent square-shouldered demagnetization curves. Similar magnetizations are found elsewhere in the complex and, when combined, they yield palaeopole *BC3*. There is also a smaller but more tenacious *D* magnetization comprising about 5% of the natural remanent magnetization which remains after removal of the *Z* magnetization at 600°C. The direction of the *D* magnetization is not significantly different from the present field. Because *D* occurs commonly throughout the Big Spruce Complex, and because it is so resistant to heat, we considered *D* (palaeopole *BC1*) to be the original magnetization. This could not be checked by a contact test because the country rocks are magnetically unstable. However, the *Z* magnetization has the characteristics of a remanence of single-domain magnetite and could be the original thermoremanent magnetization of high-temperature chemical remanent magnetization acquired during cooling following intrusion. Is *D* or *Z* the original magnetization? We now tend to favour *Z* because of its ideal demagnetization characteristics (Fig. 23-12). *D* is directed along the present field and could be caused by incipient weathering or by some as yet unstudied process related to the effects of ice-loading or permafrost. Thus *D*, which is directionally more stable under extreme demagnetization than *Z*, may be a comparatively young magnetization. If this is so, then Track 5 which, as we have seen, no longer provides an adequate representation of results from Superior Province does not describe the results from the Slave Province either.

Because of the current uncertainties in the interpretation of the data from the Nipissing diabase and the Big Spruce Complex it is very difficult to construct APW paths for the two structural provinces either separately (McGlynn and Irving, 1975; Burke et al., 1976; and Cavanaugh and Seyfert, 1976) or collectively (Irving and McGlynn, 1976b; Roy and Lapointe, 1976). Hence the palaeomagnetic evidence, as it stands at present, does not determine relative movements between the Slave and Superior Structural Provinces. The question of whether or not the intervening Hudsonian orogen is a product of processes that involved large relative motions between the two remains unsolved.

Oldest Proterozoic and Archaean

Palaeopoles from Huronian rocks and their probable time equivalents are shown in Fig. 23-13. Some of them (e.g. *GG4*) are derived from rocks whose magnetization is post-folding, others (e.g. *GG3* and *GG5*) are from rocks

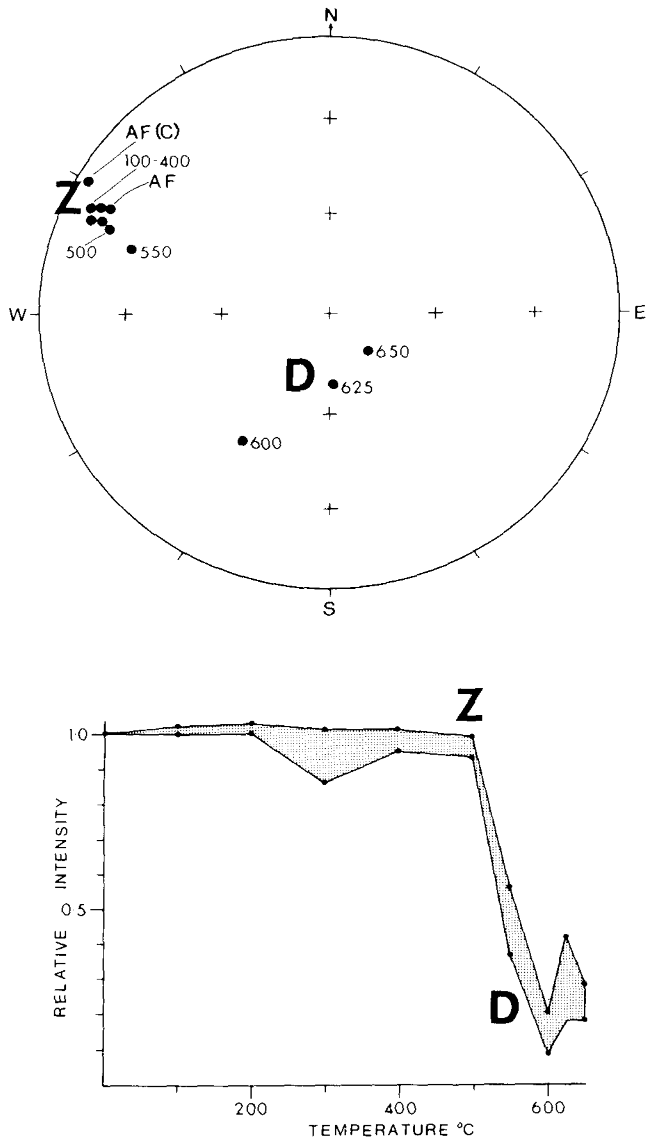

Fig. 23-12. The thermal demagnetization subsequent to a.f. demagnetization of pyroxenite from the Big Spruce Complex. The direction labelled *AF* is the mean direction (4 specimens throughout) after demagnetization in 20 mT. The mean directions after subsequent thermal demagnetization are labelled by their corresponding temperatures in °C. The direction labelled *AF(C)* is the direction of the vector obtained by subtracting the *D* magnetization from the vector *AF*. The general areas of the stereogram occupied by the *Z* and *D* magnetizations (corresponding to palaeopoles *BC1* and *BC2*, Fig. 23-10) are indicated. The thermal decay curve is shown below as the standard error envelope of the 4 curves. From Irving and McGlynn (1976a).

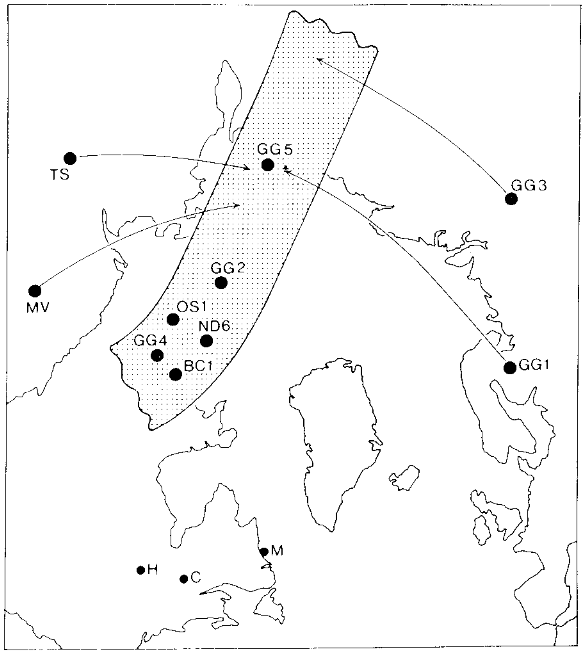

Fig. 23-13. Palaeopoles for the Huronian (*H*) and probable time equivalents, the Chibougamau Group (*C*) and the Mugford basalt (*M*). The arrows show how poles *MV*, *TS*, *GG1* and *GG3* could be fitted into a simple path if the sampling localities had undergone rotations about local vertical axes.

whose magnetization has not been shown to be pre-folding, and still others (*GG1* and *GG2*, the work of Morris, 1977a) are from rocks with pre-folding magnetization. Morris (1977a) and Symons and O'Leary (1978) have drawn complex polar loops through these poles. Alternatively, as is shown in Fig. 23-13, certain localities may have undergone local rotations and the polar path may, in reality, be comparatively simple. We do not wish to claim that such rotations have occurred but only that they may have occurred. The approximate agreement that Morris (1977a) has observed in pre-folding magnetization from two widespread Huronian localities (yielding *GG3*, Chibougamau and Cobalt) is evidence against such an effect, but it may be present elsewhere in the Huronian and requires evaluation in terms of the local geology.

The palaeopoles from the oldest rocks studied palaeomagnetically are shown in Fig. 23-14. Notable are the poles for some Archaean rocks (*SAG*, *DS*, *SW*, *KK*). The APW path is very tentative and is meant to indicate a very general age trend only. No tectonic significance should be attached to it.

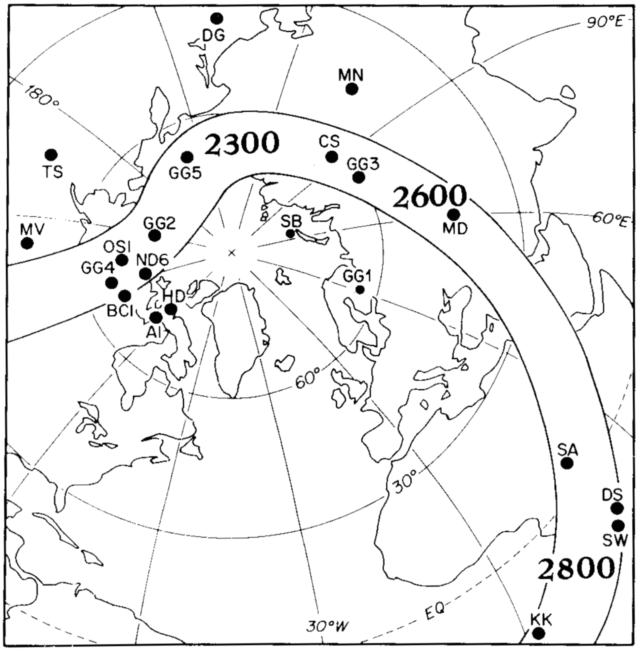

Fig. 23-14. Palaeopoles and tentative APW path for latest Archaean and earliest Proterozoic. Modified from Irving and Naldrett (1977).

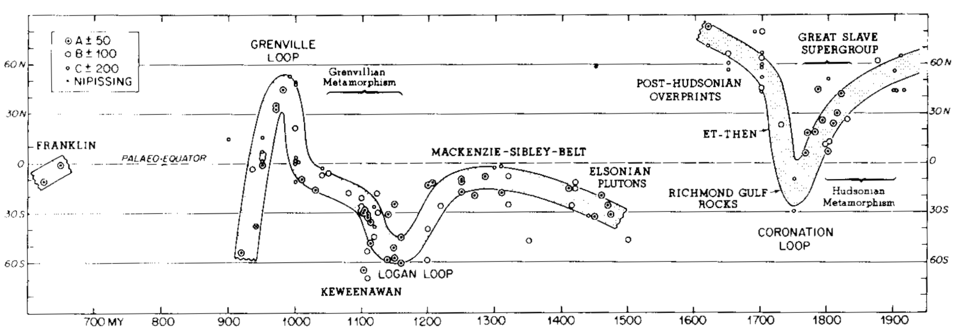

Fig. 23-15. Palaeolatitude variations for the Laurentian Shield. Calculated for Winnipeg (50°N, 97°W), using equation 9.1 of Irving (1964). Age uncertainty of individual points are in Ma.

TECTONIC SIGNIFICANCE

The palaeolatitudes determined for a central reference locality (Winnipeg) are shown in Fig. 23-15. The diagram is not extended back beyond ~ 1950 Ma because of the unsatisfactory nature of the earlier record. Most of the results can be contained within a belt 10° to 15° wide. Many results which are aberrant from the simply drawn APW paths of Figs. 23-4 and 23-7 now fall within the palaeolatitude belt and we attribute this, by analogy with the western Cordillera (Figs. 23-1 and 23-2), to the occasional occurrence in these Precambrian terrains of local rotations about vertical axes. The points that fall outside the generalized palaeolatitudinal belt are not more numerous than would be expected from errors in the positioning and dating of Precambrian palaeopoles.

The palaeolatitudinal belt has three distinctive signatures corresponding to the Coronation, Logan and Grenville Loops, respectively. During Coronation Loop time, from about 1900 to 1600 Ma ago, Laurentia moved from high to low palaeolatitudes and then back to high palaeolatitudes. The average rate of change in palaeolatitude for these 300 Ma is 5—6 cm per year. If APW is entirely caused by continental drift, then this is the minimum drift rate because movements in a palaeolongitudinal sense are not recorded. However, we do not know if there was a significant contribution to APW from true polar wander. If the latter is negligible then these minimum drift rates are very rapid indeed and comparable to the rates of motion of oceanic plates of today. They are much higher than the rates of minimum motion of continents during the past 300 Ma as recorded by APW paths. The minimum velocities for Gondwana, North America and Eurasia during the past 300 Ma are about 2 cm per year, and only reach 5 cm per year for short intervals of about 30 Ma (Gordon et al., 1979). Laurentia apparently moved relative to the pole at 5—6 cm per year for ten times this length of time during the later part of the early Proterozoic.

In the interval ~ 1450 to ~ 1200 Ma ago the average rates of palaeolatitudinal change are 1 to 2 cm per year, not greatly different from present rates, but in the interval about 1200 to ~ 930 Ma ago the rates once again become very rapid. The average is 5—6 cm per year and may reach as much as 10 cm per year. Sporadic as the record is, it does appear that the rates vary by a factor of at least 5 and that the minimum rates over long intervals during what might be loosely called the Hudsonian and Grenville Orogenies are very fast indeed.

The distribution of palaeopoles is not random (Lapointe et al., 1978). Using the present geographical frame as reference, 64% fall in latitudes of less than 30°, significantly more than the 50% that would be expected on a random distribution. At first sight this might seem to be important but tectonically it is not palaeopoles that are meaningful but the palaeolatitudes and rotations that can be calculated from them. The relevant questions relate to the distribution of these latter values which is now considered.

During the 1200 Ma for which a record is available the centrally placed reference town of Winnipeg spent 260 Ma (22%) in palaeolatitudes higher than 60°, 360 Ma (30%) in palaeolatitudes between 60 and 30°, and 580 Ma (48%) in palaeolatitudes less than 30°. This compares with 13, 37 and 50%, respectively, as expected for a uniform distribution on a sphere. Because of the ambiguity in the sign (whether north or south) of palaeolatitude it is impossible to study the symmetry of the distribution about the palaeo-equator. The Laurentia Shield does not seem to have been preferentially situated in any particular palaeolatitude zone.

The rapidity of the palaeolatitudinal changes and the great lengths of the polar tracks imply long (90°) unimpeded drift trajectories. This assumes that the contribution from true polar wander was small which, of course, is not known. Such trajectories, if rapidly executed, are difficult to envisage in an earth crowded with many independently moving continental plates. The inference is that during the part of the Proterozoic for which we have records, the continental crust was grouped into fewer, larger, faster-moving cratons than in the Phanerozoic.

The rapidity of APW and the corresponding palaeolatitudinal changes at certain times during the Proterozoic have important consequences for stratigraphy and for the problem of reconstructing relative motions of continents in the Proterozoic. The rapidity of the changes means that once an APW path has been accurately calibrated it may be used for stratigraphic correlation, and a resolution of 50 Ma or better may be possible within those intervals during which changes are fast.

While rapid palaeolatitudinal changes have advantages for stratigraphic resolution they present very serious problems for intercontinental comparisons. We shall argue that they are responsible for the mutually inconsistent tectonic interpretations that have been made, of which the following are examples. Piper (1976) has proposed that during much of the Proterozoic the palaeomagnetic results are consistent with a fixed position for Africa immediately southwest of Laurentia. Morris et al. (1979) have contended that the same data are consistent with Africa situated immediately southeast of Laurentia during the Proterozoic (as in Wegener's configuration, Wegener, 1924, quantified by Bullard et al., 1965, and Smith and Hallam, 1970). Embleton and Schmidt (1979) have argued that the data for much of the Proterozoic are consistent with Africa and Laurentia in their *present* positions. Finally, Morel and Irving (1978) and Irving and McGlynn (1979) believe that the palaeomagnetic data provide few constraints on the Proterozoic configuration of the continents and, in particular, that there was "no basis in the palaeomagnetic evidence for the proposition that the continental crust was assembled into a single unchanged Pangaea during the whole of the interval ~2200 to ~1300 Ma". How can such mutually inconsistent conclusions be drawn from the same data? Either the data are grossly inadequate or the procedures and assumptions are inappropriate. We shall attempt an answer by considering an example.

According to Embleton and Schmidt (1979) the APW paths for Africa and Laurentia more or less overlie each other. They show a southerly trending track between ~2300 and ~2100 Ma (old Track 5), followed by a westerly loop (Fig. 23-16). We have already argued that old Track 5 is no

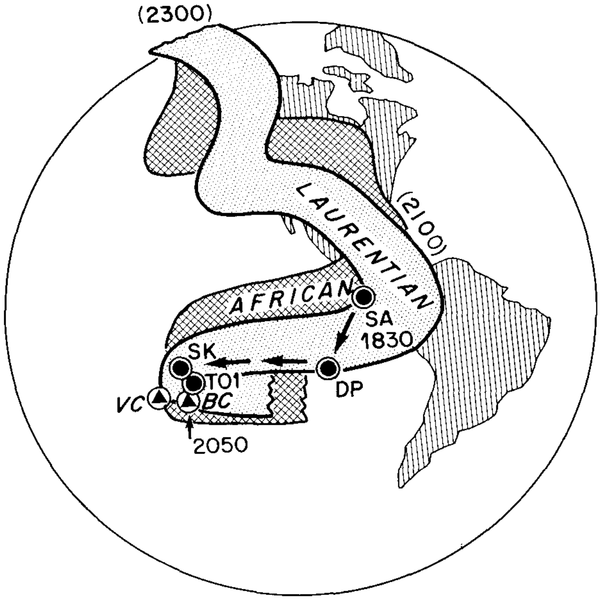

Fig. 23-16. The approximate identity of APW paths for Africa and Laurentia for the interval ~ 2300 to ~ 1900 Ma as proposed by Embleton and Schmidt (1979). Africa and Laurentia are in their *present* relative positions. The palaeopoles plotted are those that refer to the dating of the later parts of these paths as discussed in the text. They are: *BC*, Bushveld gabbro; *VC*, Vredefort Ring Complex; *SA*, Seton Formation; *DP*, Douglas Peninsula Formation; *ST*, Stark Formation; *TO1*, Tochatwi Formation. The stratigraphic order is *SA* (oldest)-*DP*-*ST*-*TO1* (youngest). Ages are in Ma. Ages in brackets are suggested by Embleton and Schmidt (1979), the other two are Rb-Sr ages (see text).

longer an adequate representation of the North American results, but in order to focus the discussion we shall consider only their western polar loop situated in the present mid-Pacific. The palaeopoles for Africa dating their loop are the Bushveld gabbro and the contemporaneous Vredefort ring complex of South Africa. The Rb-Sr isochron age of the Bushveld gabbro is 2050 ± 24 Ma (Hamilton, 1977). Embleton and Schmidt's Laurentian path also has a westerly loop defined by palaeopoles *SK* and *TO1* from the Stark and Tochatwi Formations. We believe that both these rock units have been rotated through a large angle and hence cannot be used to construct the reference path for Laurentia to say nothing of the whole world. Furthermore, the Stark and Tochatwi lie about 3000 m stratigraphically above the Seton Formation which contains volcanics dated at 1820 ± 10 Ma by a Rb-Sr isochron (Baadsgaard et al., 1973). If this age estimate is correct, the Stark and Tochatwi must be younger. Therefore to compare results from the Tochatwi and Stark Formations with results from the Vredefort Ring complex and the Bushveld complex is to compare results from rocks that could differ in age by 100 Ma or more. This conclusion is based on the available geochronological evidence obtained by the same method (Rb-Sr). From Fig. 23-15 it can be seen that such an age difference could correspond to a palaeolatitudinal change of 5000 km or more, sufficient to open an ocean. The differences in ages and the rapid palaeolatitudinal changes during this interval render the comparison made by Embleton and Schmidt (1979) questionable.

From our reading of the literature we find that similar questionable comparisons have been made frequently, and we suggest that they account for the contradictory tectonic interpretations based on palaeomagnetic results. It now seems to us that these contradictions arise because of the attempts by ourselves and others (for example, recently by Piper, 1980), to compare long segments of APW paths which represent motion over hundreds of millions of years duration. It might be more fruitful to consider the palaeomagnetic data in successive short time intervals and to construct a time sequence of trial maps with the continents in their correct palaeolatitude and azimuth, as has been done for the early Phanerozoic (see for example Smith et al., 1973). Such trial maps could then be compared with the geological evidence. By consideration of the tectonic and stratigraphic and metamorphic relationships of the rock units from which palaeopoles are obtained, and through improvements in radiometric dating, such procedures could provide results of fundamental value to studies of Proterozoic tectonics. Attempts to use this procedure have been made (Morel and Irving, 1978; Irving and McGlynn, 1979) and although the results are as yet inconclusive they do show that a very wide variety of continental configurations are consistent with the palaeomagnetic data. Moreover they provide no support for the idea that the continental configurations have remained fixed for long intervals of time during much of the Proterozoic, as is implied by the interpretations of Piper (1976, 1980), Morris et al. (1979) and Embleton and Schmidt (1979).

There are three APW loops and corresponding bends in the palaeolatitude graph for Laurentia (Fig. 23-15). The bend or hairpin of the Coronation Loop corresponds in time to the Hudsonian Orogeny. The hairpin of the Logan Loop corresponds to an interval of intense rifting in Keweenawan times and to the culmination of Grenvillian tectonism. The hairpin in the Grenville Loop corresponds to the period of post-Grenvillian uplift. These hairpins signify great changes in latitudinal motion of Laurentia, and they correlate in time with first order tectonic events. Their significance in terms of the interaction of Laurentia with other continents is still obscure because the rapidity of the palaeolatitudinal changes prevents satisfactory general comparisons with palaeomagnetic results from other continents from being made.

Plate tectonics would require that the Slave and Superior Provinces and the North Atlantic craton have moved relative to each other in the early Proterozoic, prior to the Hudsonian Orogeny. The palaeomagnetic record is as yet insufficiently precise to tell us if such movements have or have not occurred, but it does indicate that very large motions relative to the pole have taken place in the early Proterozoic as would be expected from plate tectonics. In the middle and later Proterozoic Laurentia, except for the Grenville Province, has behaved essentially as a single continent, and the results again provide abundant evidence that it has undergone large motions relative to the pole. The rates of motion are greater than in the later Phanerozoic, indicative perhaps of more rapid mantle convection driven by higher heat flow at times in the Proterozoic. Close to large faults and rifts, and in certain deformed zones, the palaeomagnetic results provide good but not conclusive evidence for the rotation of restricted blocks about local vertical axes. Such rotations are common in Phanerzoic deformed zones that can be related to plate-tectonic processes.

In summary, therefore, we conclude that the idea of the operation of plate-tectonic processes in the Proterozoic is not contradicted by the palaeomagnetic evidence from Laurentia.

ACKNOWLEDGEMENTS

We are grateful to P. F. Hoffman, A. Kröner and W. A. Morris for their criticism of the manuscript.

APPENDIX

Palaeopoles are keyed to the Earth Physics Branch Catalogue (see bibliography) or to the original if published recently.

AG	Amitsoq gneiss, Greenland (1)
AI	Algoman intrusives, Ont. (2)
AL	Aillik dykes, Labrador, 353
AN	Alona Bay Lavas, Ont. 139
AT1	Abitibi dykes primary, Ont., 110 (3)
AT2	Abitibi dykes, N. directions, Ont., 110
AT3	Abitibi dykes SW directions, Ont., 111
BA	Badcall dyke, Scotland (4)
BC1	Big Spruce Complex, N.W.T. (5)
BC2	Big Spruce Complex overprint, N.W.T. (5)
BC3	Big Spruce Complex overprint, N.W.T. (5)
BE1	Belt Series, Mont., 300
BE2	Belt Series, Alta.–Mont. (6)
BG	Baraga Co. dyke, Mich., 34
BR1	Beartooth Mts., dykes 4, 5, Colo., 386
BR2	Beartooth Mts., dykes 7, 8, Colo., 388
BY	Beaver Bay Complex, Minn., 159
CA	Cardenas lavas, Ariz., 397
CG	Chibougamau overprint, Que. (7)
CH	Charlton Bay Formation, N.W.T. (8)
CP	Coppermine Group, N.W.T., 472
CPP	Copper Harbour sandstone, Mich., 304
CRC	Croker Island Complex, Ont., 129
CS	Chibougamau sills, Que. (7)
CT	Castignon Complex, Labrador (9)
DD	Diabase dykes, Labrador (10)
DG	Dogrib dykes, N.W.T., 491
DO	Dolerite dykes, Greenland (11)
DP	Douglas Peninsula Fm., N.W.T. (49)
DS	Dundonald sill, Ont. (3)
DU	Dubawnt Group, N.W.T., 481
EP	El Paso rocks, Texas, 167
ES	Eileen sandstone, Wisc., 77

ET Et-then Group, N.W.T., 341
FA St. Francois rocks, Mo., 241
FFA Flin-Flon greenstones (younger) Man. (12)
FFB Flin-Flon greenstones (older) Man. (12)
FN1 Freda-Nonesuch Ss. combined, Mich., 303
FN2 Freda-Nonesuch overprint, Mich., (13)
FR Frontenac dykes, Ont., 363
FS Freda sandstone, Mich. (13)
GE Grenville metamorphics, Ont., 86
GFA Grenville Front anorthosites, Ont., 459
GG1 Gowganda—Chibougamau Fms. A, Ont.—Que. (14)
GG2 Gowganda—Chibougamau Fms. *B*, Ont.—Que. (14)
GG3 Coleman Member (Gowanda), Ont.
GG4 Gowganda Formation, Ont. (16)
GG5 Firstbrook Member (Gowganda), Ont. (15)
GM Grenville diorites, Que., 322
GO Gogebic rhyolites, Wisc., 366
GR Giant dykes, Gardar, Greenland (17)
GS Gila diabase, Ariz., 320
GT Gunflint Formation, Ont., 191
GU Grenville uplift overprint, Que.-Ont. (7)
GY Grand Canyon Supergroup, Ariz., 231
HBA Haliburton intrusives *A*, Ont. (18)
HBB Haliburton intrusives *B*, Ont. (18)
HBC Haliburton intrusives *C*, Ont. (18)
HD Haileybury diorites, Ont. (2)
HE Animikie hematite ore, Ont., 192
HG Hriddal dyke, Gardar, Greenland (17)
HL Harp Lake Complex, Labrador (19)
HLM Harp Lake marginal rocks, Labrador (19)
HP Harp dykes, Labrador (19)
ID Indin dykes, N.W.T., 492
IH Indian Head anorthosite, Labrador (20)
IL Ilimanssaq central syenite, Greenland, (21)
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IT Itivdledq dykes and gneisses, grnld (22)
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J2 Jacobsville Formation, Ont., 74
J1AB Jacobsville Formation, Mich. (23)
J1C Jacobsville Formation, Ont. (23)
J1SA Jacobsville Formation, Mich. (23)
KA Kaminak metamorphosed dykes, N.W.T. (24)
KC Kahocella overprint, N.W.T. (8)
KD1 Kangamuit dykes (1), Greenland (1)
KD2 Kangamuit dykes (3), Greenland (25)
KE Keweenawan intrusions, Minn., 308
KK Kamiskotia Complex, Ont. (3)
KL Kaminak lamprophyre dykes, N.W.T. (24)
KM1 Ketilidian metavolcanics (1), Greenland (26)
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KR Kungnat ring dyke, Greenland (11)
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MA Mackenzie diabase, N.W.T., 453
MAN Mamainse lavas normal, Ont., 141
MAR Mamainse lavas reversed, Ont., 140
MD Matachewan dykes, Ont. (3)
MDB Melville—Daly Bay metamorphics, N.W.T., 456
ME Menihek Formation, Labrador (9)
MGN Mamainse—Gargantua lavas N, Ont., 394
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M1V Michipicoten-volcanics, Ont., 142
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MN Munro Formation, Ont. (3)
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PRECAMBRIAN PALAEOMAGNETISM OF EUROPE AND THE POSITION OF THE BALTO-RUSSIAN PLATE RELATIVE TO LAURENTIA

R. P. E. POORTER

ABSTRACT

Palaeomagnetic poles from the Baltic Shield for the interval 2150—800 Ma and Eocambrian/early Palaeozoic data from the whole of Europe are reviewed. The results are in agreement with previously defined paths of apparent polar wander for the Svecofennian—Jotnian interval (2000—1200 Ma) and the Sveconorwegian episode of magnetization (1000—800 Ma). Eocambrian poles (640—580 Ma) average at 37°N, 124°E. The similarities between coeval paths of apparent polar wander from Baltica and Laurentia are best explained by assuming that both continents were close to one another during most of the Proterozoic. A major reorientation of the Baltic plate relative to Laurentia apparently took place during the Grenvillian orogeny. The post-Sveconorwegian and Devonian positions of Europe relative to North America are very similar, suggesting that both continents were not far apart during the intervening period.

INTRODUCTION

A review of Precambrian palaeomagnetism of Europe is necessarily focussed primarily on the Baltic Shield. Other vast Precambrian areas of Europe are the Russian Platform, which is actually covered with a thin layer of dominantly Eocambrian—Cambrian sediments, and the Ukrainian Shield. From the latter palaeomagnetic data are available. However, being established during the early period of palaeomagnetism, magnetic cleaning was not applied (Kruglyakova, 1961). Small Precambrian areas in western Europe originated mainly during or before the Cadomian orogeny and have been affected by the latter or by the subsequent Caledonian or Variscan events (Zwart and Dornsiepen, 1978). The Lewisian Platform of northwest Scotland is commonly considered to be a former part of the Laurentian Shield.

This paper is largely devoted to the Proterozoic palaeomagnetic stratigraphy of the Baltic Shield, based on apparent polar wandering (hereafter referred to as APW). This stratigraphy is achieved according to the principles outlined by Irving and McGlynn (1976), which include the consideration that during slow post-metamorphic cooling the magnetization is acquired at a substantially lower temperature than observed during laboratory experiments. Blocking temperatures of about 380°C for pure magnetite and 450°C for pure hematite can be expected for rocks that cooled at a rate of less than 1°C/10^5 year (Pullaiah et al., 1975). Conceivably, the cooling ages of an igneous-metamorphic complex as for instance obtained by K-Ar dating of hornblende and biotite are useful to bracket the magnetization age (Buchan et al., 1977). Estimates of the diffusion threshold temperature for significant argon loss vary between 225°C (Buchan et al., 1977) and 400°C (Andriessen, 1978; Verschure et al., 1979) for biotite and between 500°C (Hart et al., 1968) and 620°C (Buchan et al., 1977) for hornblende.

After a discussion of late Precambrian and early Palaeozoic pole positions the final part of this paper contains tentative reconstructions of the Baltic Shield relative to the Laurentian Shield as inferred from the respective palaeomagnetic stratigraphies.

GEOLOGICAL FRAMEWORK

Wynne-Edwards and Hasan (1970) and Zwart and Dornsiepen (1978) have shown that the orogenic belts of the North Atlantic region correlate well if the Mesozoic drift is taken into account. According to the latter authors the Laurentian, Fennosarmatian (Baltic and Ukrainian Shields with the Russian Platform) and African blocks confine a large Y-shaped area of repeated orogenic activity, possibly ever since the Grenvillian orogeny up to the Variscan orogeny. This suggestion leads to a plate-tectonic model of recurrent interaction between the three blocks. The only quantitative test of this model is to compare the respective APW stratigraphies and the attention in this paper is directed mainly to the Laurentian and Baltic areas. The APW stratigraphy of the latter is provisionally considered to be representative for the whole Fennosarmatian block.

The oldest palaeomagnetic record is from the 2200—1900 Ma old sedimentary-volcanic Jatulian sequence (Neuvonen, 1975) which overlies the 2800—2600 Ma old Archaean basement in the northeastern part of the Baltic Shield (Fig. 24-1). The Svecofennian (or Svecokarelian) orogeny rejuvenated most of the shield during the 2000—1800 Ma interval and the late-orogenic magmatism was almost without interval succeeded by the Rapakivi igneous episode (1670—1590 Ma; Vaasjoki, 1977). The igneous activity proceeded intermittently, in particular in southern Sweden, until the Sveconorwegian orogeny (1200—1000 Ma).

The age of the Jotnian volcanic-sedimentary sequence is bracketed by the intersecting late-Jotnian dolerites and sills, which were intruded between 1220 and 1270 Ma ago (Kouvo, 1976; Patchett, 1978) and the underlying Ragunda and Nordingrå granites, dated at 1320 and 1380 Ma, respectively (Welin and Lundqvist, 1975). This is in agreement with the preliminary Rb-Sr isochron age of 1320—1370 Ma obtained from Jotnian sediments in southwestern Finland (Kouvo, 1976).

In southern Sweden a dolerite dyke swarm occurs in a zone at least 100 km wide, stretching from the Dalarna district to the island of Bornholm.

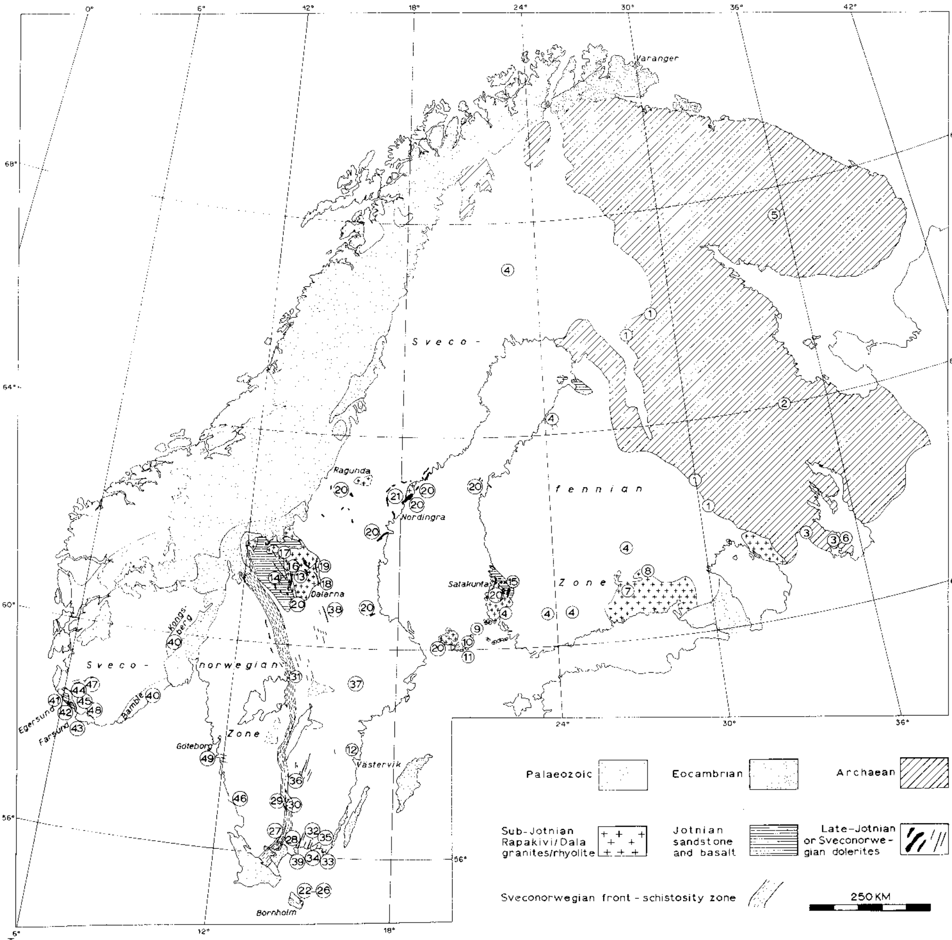

Fig. 24-1. Major geological units of the Baltic Shield and locations related to pole positions listed in Table 24-I.

These dykes, striking parallel to the Central Schistosity Zone (Fig. 24-1), were intruded between 1000 and 870 Ma ago and are possibly related to the post-orogenic uplift of the Sveconorwegian zone (Patchett, 1978).

The Sveconorwegian orogeny, which is the European counterpart of the Grenvillian orogeny, was active in southern Norway and southwestern Sweden. The poly-orogenic character of the Sveconorwegian zone is clearly demonstrated by radiometric evidence of relict ages of at least 1600—1700 Ma (O'Nions and Baadsgaard, 1971; Skiöld, 1976; Welin and Gorbatschev, 1976; Jacobsen and Heier, 1978).

REVIEW OF APPARENT POLAR WANDER TRACKS AND CURRENT RECONSTRUCTIONS

Two apparent polar wander tracks have been derived for the Baltic Shield. The middle—late Proterozoic track was proposed by Neuvonen (1970) mainly according to his and his co-workers' investigations of Svecokarelian and Jotnian rocks in Finland. Palaeopoles that have been published subsequently for the 1800—1200 Ma interval were largely in agreement with this initial track, which possibly can be extended back to about 2150 Ma. The related poles are listed in Table 24-I and are plotted in Fig. 24-3 (poles 1—24). Several authors noted that the APW tracks of North America (compare track 4a in Irving and Mc Glynn, 1976) and the Baltic Shield for the 1800—1200 Ma interval display a similar trend. Reconstructions of the position of the Baltic Shield relative to North America have been made accordingly (Fig. 24-2), matching the pertinent APW track elements (Donaldson et al., 1973; Neuvonen, 1974; Poorter, 1976a, b).

The Sveconorwegian polar pattern emerged after combining palaeomagnetic results from southwest Norway and southern Sweden (Poorter, 1972a). This pattern was confirmed by later studies of Sveconorwegian and related rocks and comprises the period from 1000 to 850 Ma ago (poles 25—49 in Table 24-I and Fig. 24-3). In an early stage of the investigation the similarity of the Grenvillian and Sveconorwegian APW tracks was already noted (Hargraves and Fish, 1972; Stewart and Irving, 1974), enabling the reconstruction of the position of the Baltic Shield relative to North America (Fig. 24-2) shortly after the Grenvillian—Sveconorwegian orogeny (Ueno et al., 1975). The reconstruction supported Wynne-Edwards and Hasan's (1970) hypothesis of a contiguous Grenvillian—Sveconorwegian belt.

The possible position of Baltica relative to Laurentia for the period before 1200 Ma and the 1000—850 Ma interval (Fig. 24-2) has two interesting aspects: the apparent unity of Laurentia and Baltica and the 90° clockwise rotation of the Baltic Shield relative to North America between 1200 and 1000 Ma ago. This rotation has been related to the origin of the Grenvillian—Sveconorwegian orogeny (Poorter, 1976a, b; Patchett et al., 1978).

DISCUSSION OF THE APPARENT POLAR WANDER STRATIGRAPHY OF THE BALTIC SHIELD

Table 24-I provides a listing of Proterozoic palaeomagnetic poles from the Baltic Shield which are depicted in Fig. 24-3. Radiometric data have been normalized where possible according to the constants recommended by Steiger and Jäger (1977). The ages assigned to the pole positions from the slowly cooled igneous-metamorphic Sveconorwegian zone (poles 40—46) were deduced according to the outline given above.

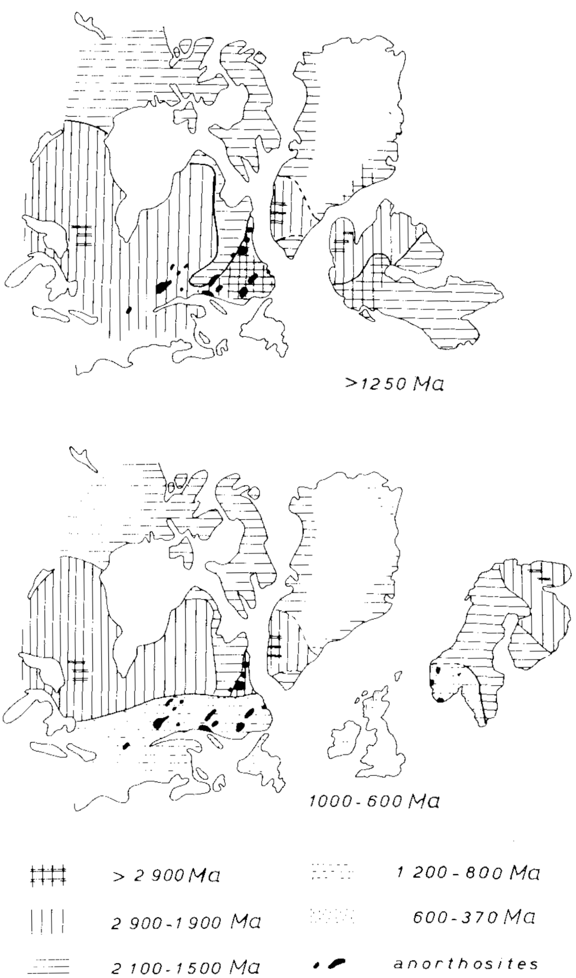

Fig. 24-2. Preliminary reconstructions of the Baltic Shield relative to Laurentia (from Poorter, 1976b).

The oldest part of the Proterozoic track comprises palaeopoles (2, 3, 5, 6) from the Russian part of the Baltic Shield. They are close to the well established mean pole 4 from synorogenic Svecokarelian intrusives in Finland and Sweden, which have an average intrusion age of 1870—1900 Ma. Post-orogenic cooling ages are not available for these intrusives, although a late- or post-orogenic igneous episode is mentioned at about 1800—1850 Ma ago (Kouvo, 1976). Accordingly, an age of 1800—1900 Ma is assigned to pole 4 and related poles 2—6 (Fig. 24-4), taking into account a protracted cooling period. Pole 1 from the 2150 Ma old Jatulian rocks in Finland is not greatly

TABLE 24-I

Proterozoic palaeomagnetic pole positions from the Baltic Shield and Northwest Scotland

Rock unit	Catal.[a]	Age (Ma)	Method/Ref.[b]	Sites/Samples[c]	Pole position (°)	dp (°)	dm (°)	Ref.
1. Jatulian igneous rocks, Finland	15.188	2150	U/(1)	4/12	60.0N 215.0E	α_{95} = 20		(1)
2. Jatulian redbeds and lavas, USSR	OT 405	1610–1870	K(WR)/—	12/38	58.0N 238.0E	6	12	—
3. Jotnian sandstones, Karelia, USSR	OT 257	1850–1950	K,Pb-Th/—	?/86	49.0N 236.0E	—	—	—
4. Svecokarelian gabbros, Finland	15.187	1870–1900	U/(2)	7/0	45.5N 241.0E	α_{95} = 7.2		(2)
5. Pap intrusion, Kola, USSR	OT 345	1820–1880	K(WR)/—	1/12	40.0N 223.0E	9	16	—
6. Ryaboretskii sill, Onega, USSR	OT 344	1600–1800	?/—	0/29	32.0N 231.0E	1	2	—
Igneous rocks associated with Rapakivi granites, Finland								
7. Kuisaari quartz diabase	—	1590–1670	text	0/9	22.0N 190.0E	7	14	(3)
8. Luontarivesi granodiorite	—	1590–1670	text	0/25	67.0N 198.0E	8	10	(3)
9. Åva intrusives	OT 198	1590–1670	text	0/15	41.0N 169.0E	9	17	(4)
10. Kumlinge dolerites	OT 202	1590–1670	text	0/7	13.0N 201.0E	11	19	(4)
11. Föglo dolerites	OT 200	1590–1670	text	0/8	31.3N 186.5E	6	12	(5)
Gothian–Svecofennian								
12. Loftahammar gabbro, Sweden	15.212	1480–1658	text	5/26	23.0N 179.0E	3	5	(6)
13. Upper Dala volcanics, Sweden	OT 098	1480–1624	text	2/11	22.7N 176.0E	—	—	(7)
d. Ragunda Complex		1293 ± 30	Rb(WR)/d)	19/111	54.0N 164.9E	6	9	(d)
Jotnian sedimentary sequence								
14. Öje basalt, Sweden	OT 099	1250–1350	text	2/16	31.7N 185.1E	—	—	(7)
15. Satakunta sandstone, Finland	OT 475	1250–1350	text	0/10	3.0N 180.0E	8	12	(8)
Late-Jotnian dolerites and monzonites								
16. Bunkris dike, Sweden	—	1516 ± 62, 1546 ± 84	Rb(M)/(9)	1/5	24.3N 215.5E	9	17	(7)
17. Bunkris dike at Glysjön, Sweden	—	—	—	1/5	33.9N 168.0E	3	6	(7)
18. Emådalen sill, Sweden	—	1223 ± 36	Rb(M,WR)/(9)	1/5	27.1N 163.0E	4	7	(7)
19. Älvo sill, Sweden	—	1215 ± 18, 1290 ± 63	Rb(M,WR)/(9)	1/5	17.3N 170.2E	9	13	(11)
20. Combined, Sweden and Finland	—	1220–1270	text	10/0	0.4S 155.9E	α_{95} = 5.6		text
21. Nordingrå dolerite, Sweden	—	1254 ± 20	K(WR)/(10)	48/322	7.2S 156.9E	2.0	3.1	(12)
e. Post-Ragunda dolerites		< 1293		12/64	14.5N 194.4E	8	15	(d)
Bornholm dolerite dikes, Denmark								
22. Bolshavn, A component				1/9	12.8N 152.9E	6	10	(13)
23. Vigehavn				1/7	15.0N 158.8E	4	7	(13)
24. Vaseaa				1/11	16.0S 127.8E	7	10	(13)
25. Bölshavn, B component				1/6	20.7N 223.6E	8	14	(13)
26. Listed				1/32	13.8N 250.1E	2	4	(13)

TABLE 24-I Continued

Rock unit	Catal.[a]	Age (Ma)	Method/Ref.[b]	Sites/Samples[c]	Pole position (°)		dp (°)	dm (°)	Ref.
Dolerite dikes striking parallel Sveconorwegian front, Sweden									
27. Osby	OT 101	781 ± 25	K(WR)/(7)	1/5	2.0S	213.0E	6	8	(7)
28. Hägghult	OT 101	838 ± 25, 852–958	K(WR,M)/(7, 14)	1/50	30.0S	211.0E	2	2	(7)
29. Hjortsjö	OT 101	1573 ± 50	K(WR)/(7)	1/5	27.0S	216.0E	3	3	(7)
30. Målaskog	OT 101	886 ± 45	K(WR)/(7)	1/5	10.0S	239.0E	2	3	(7)
31. Kristinehamn	OT 101	1516 ± 50	K(WR)/(7)	1/5	2.0S	241.0E	5	8	(7)
32. Bräkne–Hoby	—	880 ± 120, 975 ± 58	Rb(M)/(9)	1/7	22.2N	251.9E	3	6	(15)
33. Fäjö				1/6	23.3N	248.9E	3	6	(15)
34. Tärnö		871 ± 21, 880 ± 28	Rb(M)/(9)	1/5	0.3N	237.0E	4	6	(15)
35. Väby	14.528			2/7	6.5N	242.8E	—	—	(15, 16)
36. Nilstorp		984 ± 47	Rb(M)/(9)	1/6	9.0N	238.5E	6	11	(15)
37. Årby		995 ± 65	Rb(M)/(9)	1/6	7.3S	227.4E	7	10	(15)
38. Falun		914 ± 13, 966 ± 20	Rb(M)/(9)	1/10	6.1S	237.6E	5	7	(15)
39. Karlshamn	14.527	871 ± 25, 936 ± 56	Rb(M)/(9)	2/17	41.6S	210.2E	—	—	(15, 16)
Slowly cooled igneous-metamorphic Sveconorwegian zone									
40. Kongsberg–Bamble hyperites and amphiboles, Norway	14.523	970–1000	text	6/48	3.0S	207.0E	$\alpha_{95} = 6.4$		(16)
41. Egersund anorthosite 1, Norway		900	text	0/26	37.0S	206.5E	8	8	(17)
42. Egersund anorthosite 2, Norway	OT 327/476	900	text	7/33	40.0S	203.6E	14	14	(18, 19)
43. Farsund complex, Norway	15.198	900	text	7/35	43.3S	194.0E	10	10	(20)
44. Bjerkreim–Sokndal lopolith, Norway	OT 328	870–900	text	9/45	41.8S	229.3E	13	14	(19)
45. Rogaland migmatite complex, Norway	OT 329	870–900	text	7/27	37.3S	231.8E	16	19	(19)
46. Falkenberg amphibolite, Sweden		900	text	1/18	27.0S	246.0E	5	6	(16)
Post-Sveconorwegian dolerite dykes									
47. Hunedalen ENE system, Norway	OT 332	900	text	8/37	34.0S	208.0E	8	9	(19)
48. Egersund WNW system combined, N.	OT 156/331	< 870	text	7/68	24.4S	231.1E	9	11	(19, 21)
49. Göteborg WNW dikes, Sweden	14.526	< 870	text	0/10	17.0S	239.0E	2	3	(22)
Northwest Scotland									
50. Scourian granulite	15.206	1800		0/22	43.8N	273.8E	7	10	(23)
51. Scourie dikes	15.204	1800		9/181	37.3N	274.8E	7	10	(23)
52. Badcall dike	15.205	1440		0/32	7.9N	250.8E	2	5	(23)
53. Stoer Group	OT 422	970 ± 24	Rb(WR)/(24)	20/50	35.0N	234.0E	5	9	(24)
54. Basal Torridon Group	OT 423	779 ± 24	Rb(WR)/(24)	11/30	28.0S	228.0E	19	24	(24)
55. Torridon Group, normal		779 ± 24	Rb(WR)/(24)	28/—	0.0N	236.0E	6	10	(24)
56. Torridon Group, reversed		779 ± 24	Rb(WR)/(24)	53/—	9.0S	217.0E	5	7	(24)

[a] OT refers to Irving and Hastie (1975); 14. and 15. refer to McElhinny and Cowley (1977, 1978).
[b] U: U-Pb ages mainly of Zircon; K: K-Ar ages; Rb: Rb-Sr isochron ages; WR: whole-rock; M: minerals.
[c] Pole position at sample level for single site (1) or limited sampling over an area (0); pole position at site level for 2 or more sites; pole position at intrusion level, indicated by number of intrusions involved and sample number 0.

References: 1–2, Neuvonen (1975, 1974); 3–4, Neuvonen (1978, 1970); 5, Neuvonen and Grundström (1969); 6, Poorter (1976a); 7, Mulder (1971); 8, Neuvonen (1973); 9, Patchett (1978); 10, Welin and Lundqvist (1975); 11, Patchett et al. (1978); 12, Piper (1979a); 13, Abrahamsen (1977); 14, Klingspor (1976); 15, Patchett and Bylund (1977); 16, Poorter (1975); 17, Murthy and Deutsch (1974); 18, Hargraves and Fish (1972); 19, Poorter (1972a); 20, Murthy and Deutsch (1975); 21, Storetvedt and Gidskehaug (1968); 22, Abrahamsen (1974); 23, Beckmann (1976); 24, Stewart and Irving (1974); d and e, Piper (1979c).

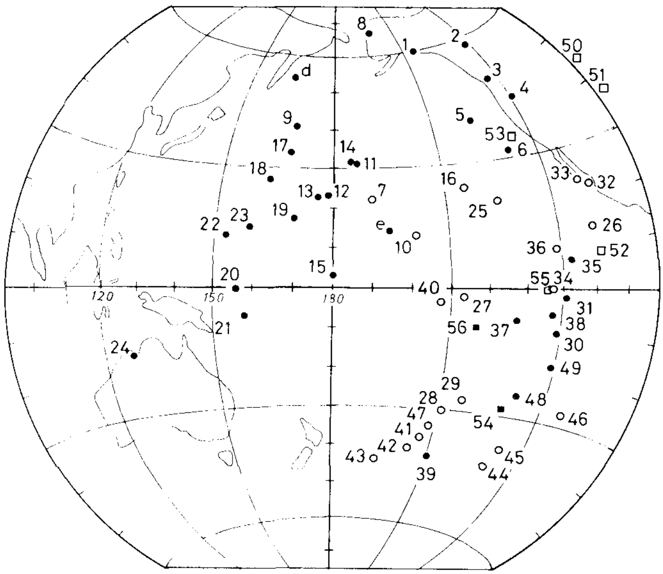

Fig. 24-3. Proterozoic pole positions from the Baltic Shield and northwest Scotland. Closed symbols: normal defined polarity; open symbols: reversed defined polarity. Numbers as in Table 24-I.

different from poles 2—6 (1800—1900 Ma), which leads to the conclusion that either plate motion during the 2150—1900 Ma interval was limited or that the magnetization has a younger age and possibly occurred during large-scale regional uplift of northern Sweden and central Finland according to Neuvonen (1975). The only evidence for an additional APW track for the 2150—1900 Ma interval analogous to track 4b for Laurentia (Irving and McGlynn, 1976) is a 2078 Ma pole at 8°N, 230°E from the Uralian belt (1.407, Irving and Hastie, 1975).

Neuvonen and co-workers investigated igneous rocks close to contacts with Rapakivi granites (poles 7—11). The actual magnetization of these rocks, which are older than or contemporaneous with the Rapakivi granites, is likely to have originated during the cooling of the neighbouring granite bodies. The emplacement of the Rapakivi granites took place between 1670 and 1590 Ma ago according to the U-Pb zircon ages reported by Vaasjoki (1977). Poles 7 and 11 are considered as the more reliable representatives for the cooling episode of the Rapakivi granites at about 1600 Ma, whereas poles 8, 9 and 10, which plot outside the proposed APW track (Fig. 24-4), are regarded as the less reliable pole positions (Neuvonen, 1970, 1978).

The palaeopoles observed from the Loftahammar gabbro and the Upper Dala Sub-Jotnian volcanics (12 and 13) are very close. A maximum magnetization age for the Loftahammar gabbro is imposed by Rb-Sr isochron

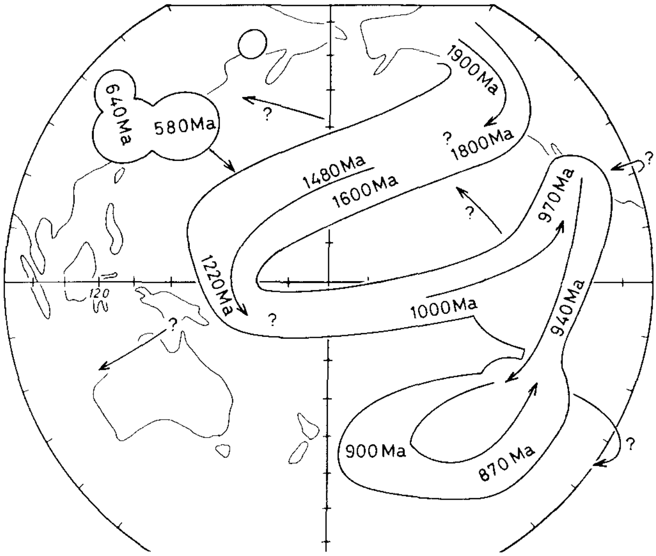

Fig. 24-4. Apparent polar wander swathe relative to the Baltic Shield incorporating ovals or circles of confidence.

ages for the associated Loftahammar gneissic granite of 1658 ± 35 Ma (Priem and Bakker, 1973) and 1620 ± 40 Ma (Åberg, 1978). A Rb-Sr isochron age of 1624 ± 42 Ma applies to the Upper Dala volcanics (Priem et al., 1970). These ages are in concert with the Rapakivi igneous episode. Additional geochronological information on the Loftahammar area of southeastern Sweden suggest that the actual magnetization may have taken place later. Biotite ages according to K-Ar dating in the Västervik area near Loftahammar average out at 1480 Ma (11 values: 3 from Priem et al., 1969, and 8 from Åberg, 1978). A K-Ar biotite reference line suggests 1565 Ma (Åberg, 1978). Three Rb-Sr biotite ages are about 1430 Ma and a whole-rock Rb-Sr isochron for a pegmatite yielded 1475 ± 40 Ma (Priem et al., 1969). Rejuvenation of K-Ar biotite ages in southeastern Sweden is reflected by ages from 1400 Ma near the Central Schistosity Zone, increasing to 1500 Ma eastward near the Baltic Shield coast. This resetting is possibly due to large-scale igneous activity during the so-called Gothian event (Åberg, 1978). Accordingly, the lower limit to the magnetization age of the Loftahammar gabbro is 1480 Ma.

Mulder (1971) determined the magnetization of the Jotnian Öje basalt from 9 sites and found only 2 sites yielding a very similar characteristic magnetization (pole 14). Its position near the Rapakivi poles (7 and 11) and the Loftahammar and Upper Dala poles (12 and 13) suggests a period of stationary pole position for at least the 1480—1300 Ma interval. At the end of this interval, however, there are two deviating pole positions, separated by 50° from each other and situated outside the suggested middle—late Proterozoic APW track. One is from the Satakunta sandstone (pole 15); the other is from the Ragunda granite complex recently reported by Piper (1979c), with a quoted age of 1293 ± 30 Ma (pole added in Table 24-I as entry d). According to Piper (1979c), these poles may be indicative of an additional APW loop. Post-Ragunda dykes show a palaeomagnetic pole (Table 24-I, entry e) between the middle—late Proterozoic and the Sveconorwegian APW tracks (Piper, 1979c). Poles 16 and 17 are from localities supposed to be in one dolerite dyke, thought to be of Sveconorwegian age because of its strike, but yielding suspect Rb-Sr isochron ages of 1516 ± 62 and 1546 ± 84 Ma as determined for the locality at Bunkris (Patchett, 1978). The pole position (16) actually supports a Sveconorwegian age rather than the older radiometric age. Pole 17 approaches the poles representing the 1600 to 1200 Ma interval; however, a multicomponent magnetization is indicated (Mulder, 1971, fig. 6, ZDE2). The difference between the magnétization reflected by poles 16 and 17 can be explained by assuming two intrusions of different age.

The pole positions reported from late-Jotnian dolerite dykes and sills appear to be in close agreement, indicating a short period of emplacement. The late-Jotnian igneous episode is coeval with the Mackenzie igneous episode in North America and the Gardar episode in southern Greenland (Patchett et al., 1978). Pole 20 is the average pole position of intrusions from Finland and Sweden including Märket, Satakunta and Vaasa (Neuvonen, 1965, 1966; Neuvonen and Grundström, 1969) in Finland; in Sweden: Nordingrå (Magnusson and Larson, 1977), Nordingrå, Gnarp and Gävle (Poorter, 1976a), Lybergsgnupen (Mulder, 1971) and Älvdalsåsen and Sundsjö/Giman (Patchett et al., 1978). Between-site differences are of the same order of magnitude as between-intrusion differences. Pole 20 is considered to be representative for the late-Jotnian igneous episode of the Baltic Shield, covering results from the whole area with a related mean Rb-Sr age of about 1218 ± 68 Ma (Patchett, 1978). Pole 21 from the Nordingrå dolerite complex, based on a large number of sites and samples (Piper, 1979a), possibly represents a shorter interval. The K-Ar isochron age for the Nordingrå dolerite is 1254 ± 20 Ma (Welin and Lundqvist, 1975); U-Pb zircon ages of about 1270 Ma apply to the Märket and Satakunta intrusives (Kouvo, 1976). Significantly different from the overall late-Jotnian pole positions are poles 18 and 19 from two associated minor sills, although their Rb-Sr ages agree with the average late-Jotnian age. Poles 22—24 from undated dolerite dykes on Bornholm are near the late-Jotnian poles. According to Abrahamsen (1977) pole 24 possibly reflects a late-Jotnian intrusion, whereas poles 22 and 23 are due to Caledonian overprinting or intrusion.

It is worth noting that the poles confining the APW path for the 2150—1200 Ma interval (Fig. 24-4) are all based on the same polarity, except for the Rapakivi poles 7 and 11 (Neuvonen, 1978), a small percentage of reversed directions in the Jotnian sandstone of the USSR (pole 3) and the Ragunda poles d and e (Piper, 1979c). We define the predominant polarity normal in concert with Neuvonen (1978) and McElhinny and Cowley (1977, 1978). The NNE dolerite dykes, striking parallel to the Sveconorwegian front, exhibit Rb-Sr isochron mineral ages in the range 100 to 870 Ma (Patchett, 1978). Dolerites (hyperites) within the Schistosity Zone show K-Ar whole-rock ages between 700 and 1600 Ma (Mulder, 1971; Klingspor, 1976). The higher ages are anomalous when compared to the Rb-Sr data reported by Patchett (1978) on the dykes east of the Schistosity Zone and the lower limit imposed by the intersected 1450 Ma old Karlshamn granite. Pole positions from 13 dykes (27—39) (Mulder, 1971; Patchett and Bylund, 1977) show an arc-like streaking over about 70°. Being near this locus of poles, two Bornholm dykes (poles 25 and 26) are believed to have intruded during the 1000—870 Ma interval (Abrahamsen, 1977). The position of pole 16 of the Bunkris dolerite likewise suggests a Sveconorwegian age.

The age of pole 40 derived from the Kongsberg and Bamble regions can be deduced from K-Ar biotite ages in the range 1000 to 970 Ma (O'Nions et al., 1969). Pertinent K-Ar hornblende ages are 1040—1010 Ma, imposing the higher limit. The Kongsberg and Bamble pole accordingly defines the oldest part of the Sveconorwegian APW track.

Pole positions from slowly cooled rocks of southwestern Norway delimit the southern part of the Sveconorwegian track. Hornblende samples from the migmatite complex (pole 45) and the Bjerkreim—Sokndal lopolith (pole 44) reveal K-Ar hornblende ages between 972 and 937 Ma (16 determinations), averaging at 953 ± 10 Ma (Dekker, 1978). K-Ar biotite ages of the migmatite complex are about 870 Ma, an age supported by Rb-Sr biotite dating on the same samples. Near the Caledonian front K-Ar ages of brown biotite yield about 800 Ma, but at the same place a younger generation of green biotite reflects Caledonian ages (Verschure et al., 1979). Since a temperature of about 400°C is estimated for the biotite-producing reaction, argon loss in biotite seems to become significant at 400°C rather than at 300°C or less as reported for other geological environments (Jäger et al., 1967). At a greater distance from the Caledonian front, Caledonian reheating obviously did not exceed 400°C. Accordingly, the magnetization age of the migmatites is fixed at 870 Ma as indicated by the K-Ar ages of the original biotite. According to Pasteels et al. (1979) the emplacement of the Bjerkreim—Sokndal lopolith (pole 44) and the nearby Farsund complex (pole 43) started at about 955 Ma ago and intrusion of late magmatic differentiates continued until 910 Ma. Inferring a common cooling history for the anorthosites, the Farsund complex and the Bjerkreim—Sokndal lopolith, the magnetization would not have taken place before 910 Ma. A tentative age of 900 Ma is therefore adopted, taking into account the high coercive magnetization mainly residing in exsolved hematite-ilmenite. The APW shift between poles 41 and 43 and poles 44—45 reflects an actual drift episode between 900 and 870 Ma. This is in agreement with the Karlshamn dolerite pole 39 which is located between the poles from southwestern Norway and has a similar age.

Pole 47 for the Hunedalen dykes indicates simultaneous magnetization with the anorthosites during the post-orogenic cooling. The dykes apparently intruded into a relatively hot environment as shown by backveining and amphibolitization. The Egersund dykes (pole 48) intruded a cold environment since glass is present in the chilled margins and the adjacent country rock has been remagnetized (Poorter, 1972a). The palaeopole for the Egersund dykes indicates a relatively young APW shift. Close to this pole and revealing the same polarity is pole 49 from the similarly WNW-striking dykes near Gothenborg in southwestern Sweden. The Gothenborg and Egersund dykes are actually aligned, which also suggests a contemporaneous origin.

Amphibolites sampled at three locations in southwestern Sweden and at one in the extreme south of the Bamble sector yielded directions characteristic for southwestern Norway (Poorter, 1975). Their average pole position is near poles 43 and 44, but is not included in Table 24-I ($\alpha_{95} = 37°$, 4 sites); only the most reliable pole 46 is given.

A straightforward magnetostratigraphic interpretation of the Sveconorwegian polar pattern is indicated by the arrows in Fig. 24-4. The suggested APW track accomodates the majority of the radiometric data, except for the suspect K-Ar whole-rock ages.

The slowly cooled igneous-metamorphic rocks of the Sveconorwegian zone obviously display a single polarity which is here conventionally defined as reversed. The dolerite dykes exhibit both reversed and normal polarities with a tendency to cluster.

There is no decisive proof for a link between the pre-1220 Ma APW track and the Sveconorwegian track. The connection suggested in Fig. 24-4 is simply the shortest one. An alternative has been put forward by Abrahamsen (1977) with the intention of incorporating pole 24. He suggested that the anti-polar track for the 1900—1220 Ma interval be connected via pole 24 with the Sveconorwegian track.

APPARENT POLAR WANDER RELATIVE TO NORTHWEST SCOTLAND

Table 24-I comprises poles 50 to 56 from northwest Scotland. After correction for Mesozoic drift (88.4°N, 27.7°E, + 38.0), poles 50 to 52 indeed fall near the Laurentian Proterozoic APW path (Beckman, 1976) and not on the APW path relative to the Baltic Shield. The results from the late Precambrian Stoer and Torridon Groups fit the Laurentian polar path for the Grenville episode after drift correction and, in fact, also the Sveconorwegian polar pattern (Figs. 24-3 and 24-6). Pole 53 is not far from the

coeval pole 32. The poles from the 780 Ma Torridon Group are situated within the Sveconorwegian pole path.

LATE PRECAMBRIAN AND EARLY PALAEOZOIC POLE POSITIONS

Palaeomagnetic pole positions from Europe for the interval 800—640 Ma are virtually lacking, which hampers a connection between the Sveconorwegian polar track and the Phanerozoic APW path. A comparable situation exists for North America where palaeomagnetic poles of this age are relatively scarce. An interesting suggestion has been made by Morris and Roy (1977) to connect the Grenville track with late Precambrian poles via a great circle — the Hadrynian track — although their use of the Grenville dyke poles should only be reluctantly accepted. The age of the Grenville dykes is not yet well established and some of the dykes are considered to yield a secondary magnetization (Murthy, 1971).

The pattern displayed by all available late Precambrian—early Palaeozoic poles of Europe is diffuse. In a first attempt to elucidate the late Precambrian APW relative to Europe, the pole positions listed in Table 24-II were selected according to the following criteria: (1) minimum number of samples 10; (2) α_{95} less than 20°; (3) "in situ" directions not parallel to the present field direction, unless recent remagnetization is disproven (e.g. by foldtest or baked contacts). These criteria are largely in agreement with Hicken et al. (1972) for admittance to "A" data. Following McElhinny (1973) Cambro-Ordovician data from the Bohemian Massif were not included because of their inconsistency. These data, however, reasonably fit with models of relatively high APW rates (e.g. Hagstrum et al., 1980; Van der Voo et al., 1980).

Fig. 24-5 shows that the Eocambrian poles 13 and 14 from the Armorican Massif and 20 and 22 from the Baltic Shield are situated close together. These poles in the range 640—580 Ma average at 37° N, 124° E (α_{95} = 15). Most other poles in the range 640—530 Ma are homogeneously dispersed about this average pole. The deviating pole 2 from Eocambrian tillites in Scotland is considered to be younger than pole 1 and probably reflects overprinting during the Caledonian orogeny (Tarling, 1974). The magnetic anisotropy associated with the remanent magnetization of these tillites appeared to be almost negligible (Abouzakhm and Tarling, 1975). Poles 4—7 from late Precambrian inliers of the Midland Craton of England and Wales are interpreted by Piper (1979b) in terms of large APW movement between about 60°S and 60°N during the 700—600 Ma interval. A difficulty in interpreting these results, however, remains the discrepancy between the poles from the Uriconian volcanics (poles 5 and 6). A palaeomagnetic study elsewhere of the Uriconian and the overlying Longmyndian (pole 8) revealed a remanence considered to be secondary (Lomax and Briden, 1977), the latter determined to be older than the intersecting dolerites (pole 9).

TABLE 24-II

Late Precambrian and Early Palaeozoic pole positions from Europe

Rock unit	Age (Ma)	Method/Ref.	Sites/Samples	Pole position (°)	dp (°)	dm (°)	Ref.
Scotland							
1. Port Askaig formation	Eocambrian		5/29	39.0N 190.0E	7	13	(1)
2. Port Askaig formation	Eocambrian		4/28	17.0S 129.0E	16	22	(1)
3. Mean pole (Arenigian lavas, Aberdeen gabbro, Foyer complex)	Ol			12.0N 179.0E	α_{95} = 16		(2)
England and Wales							
4. Stanner–Hanter complex	418?	K(WR)/(3)	15/84	31.4N 279.1E	11	16	(3)
5. Western Uriconian volcanics	677	K(WR)/(3)	6/39	36.9N 235.3E	9	16	(3)
6. Eastern Uriconian volcanics	632	K(WR)/(3)	9/44	13.9N 91.5E	7	13	(3)
7. Post Uriconian dolerites	638	K(WR)/(3)	11/70	37.0S 257.0E	13	19	(3)
8. Longmyndian sandstone, secondary comp., in situ	< 600	Rb(WR)/(4)	69/449	60.0S 209.0E	7	13	(4)
9. Dikes and baked contacts in Longmyndian	Silurian	(5)	13/139	19.0N 223.0E	6	11	(4)
10. Leicester diorites	534	Rb(WR)/(6)		63.0N 141.0E			(6)
11. Caerfai series	Є-Ol			26.0N 169.0E	7	13	(7)
12. Mean pole (Breiden Hill Inlier, Builth volcanics, Ashgillian intrusives, Borrowdale volcanic Group, Eycott Group, Carrock Fell complex)	Ol-u			7.0N 188.7E	α_{95} = 12		(8, 9, 10)
Armorican region							
13. Spilites de Paimpol	640	Rb(WR)/(11)	12/53	34.0N 117.0E	6	12	(11)
14. Diorite de St. Quay	583	Rb(WR)/(11)	7/40	34.0N 139.0E	2	4	(11)
15. Rhyolites de Lézardrieux et de St. Germain-le-Gaillard	546	Rb(WR)/(11)	8/49	16.0N 163.0E	12	20	(11)
16. idem, sec. component			3/12	38.0N 262.0E	8	14	(11)
17. Gabbro de Keralain			3/36	31.0N 268.0E	2	4	(11)
18. Syncline de Zone Bocaine/Syncline de May	Є-Єu		2/0	7.0N 211.0E			(12)
19. Paimpol–Bréhec/Erquy–Cap Fréhel	470		2/0	38.0N 233.5E			(13)
Norway and Sweden							
20. Båtsfjord dolerites	640	K(WR)/(14)	6/34	40.9N 105.3E	α_{95} = 7		(14)
21. Fen complex	595–600	K(M)/(18)	1/19	63.0N 142.0E	3	4	(15)
22. Nexö sandstone[a]	Eocambrian		15/56	37.6N 134.3E	7	13	(16)
Russian Platform							
23. Mean pole	Ol			39.0N 150.0E			(2)
Ural Mountains							
24. Mean pole	Єl			13.4N 166.4E	α_{95} = 21		(17)

[a] recalculated pole based on thermally cleaned remanence; pole based on Af cleaning: 47°N, 150.5°E.

1. Tarling (1974); 2. Morel and Irving (1978); 3. Piper (1979b); 4. Lomax and Briden (1977); 5. Piper (1978); 6. Duff (1977); 7. Briden et al. (1979); 8. Piper and Stearn (1975); 9. Piper and Briden (1973); 10. Faller and Briden (1977); 11. Hagstrum et al. (1980); 12. Jones (1978); 13. Jones et al. (1979); 14. Kjøde et al. (1978); 15. Poorter (1972b); 16. Prasad and Sharma (1978); 17. Krs (1978); 18. Verschure et al. (1979).

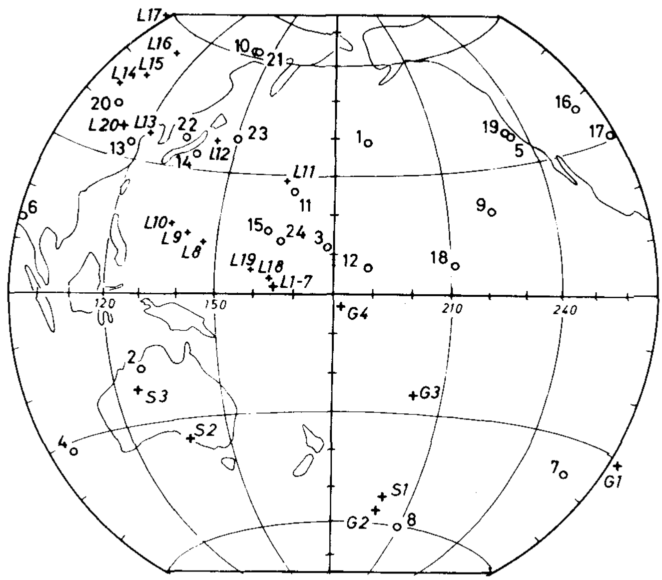

Fig. 24-5. Late Precambrian/Cambrian pole positions from Europe (Table 24-II) and Laurentia (L) (Table 24-III); mean poles for Siberia (S) and Gondwana (G) (Table 24-III).

PRECAMBRIAN PLATE TECTONICS

The Proterozoic magnetostratigraphy of the Baltic Shield and the whole of Europe respectively allow tentative reconstructions relative to the Laurentian Shield. Preliminary reconstructions depicted in Fig. 24-2 illustrate the coherence of the Laurentian and Baltic plates during certain Proterozoic intervals.

Although the reconstruction method based on matching coeval APW tracks has been criticized because of the constraint that no relative movement should have occurred during this interval (Morel and Irving, 1978), the fact that APW tracks can be matched at all essentially rules out the possibility of extensive motion of Laurentia relative to Baltica during the periods in question.

The reconstructions obtained using the above method are shown in Fig. 24-6. The pole positions of North America are plotted together with the coeval pole positions of the Baltic Shield after rotation. The poles and angles of rotation are listed in Table 24-IV.

Instead of the reconstruction shown in Fig. 24-2 for the pre-1200 Ma interval (1200—1800 Ma) the positions A and B in Fig. 24-6 were obtained by matching the 1200—1500 Ma and 1500—2100 Ma polar tracks. Position A is speculative, being subject to the many uncertainties concerning the age of the various pole positions that are matched. Position B is fixed largely on superposition of the keypoles 20 (average of late-Jotnian dolerites) and

TABLE 24-III

Late Precambrian and Early Palaeozoic pole positions from North America, Siberia and Gondwana

Pole	Rock unit	Age (Ma)	Pole position (°)
North America (1, 2)			
L1	Mafic rocks, Michigan	PreЄ—Є?	9S 159E
L2	Franklin rocks group A, Canada	675—625	4S 161E
L3	Intrusives, Colorado	704—485	5N 174E
L4	Cloud Mountain basalt, New Foundland	605	5N 172E
L5	Upper Keweenawan overprint, Michigan	PreЄ—Є?	4N 169E
L6	Franklin rocks group B, Canada	675—625	8N 166E
L7	Tapeats sandstone, Arizona	Cl-m	5N 158E
	Average of L1—L7		2N 165E
L8	Ophiolite complex, Quebec	550?	13N 146E
L9	Intrusives, Colorado	704—485	15N 142E
L10	Tudor gabbro, Ontario	1200—670	17N 137E
L11	Bradore formation, New Foundland	Єl	29N 167E
L12	Rome formation, Tennesee	Єm	38N 144E
L13	Intrusives, Colorado	704—485	37N 122E
L14	idem (3)	525	39N 100E
L15	idem	idem	48N 107E
L16	Muav formation, Arizona	Єm-u	55N 110E
L17	Abrigo formation, Arizona	Єm-u	59N 89E
L18	Wichita granites, Oklahoma (4)	525	4S 344E
L19	Wilberns formation, Texas (5)	Cu	6S 339E
L20	Mean Middle Ordovician pole (6)	Om	27N 112E
Siberia			
S1	Mean pole (7)	600	52S 199E
S2	Mean Cambrian pole (7)	Єl-u	35S 138E
S3	Mean Ordovician pole (7)	Ol-u	22S 127E
Gondwana			
G1	Mean pole (7)	600	31S 273E
G2	Mean Cambrian pole (7)	Єl-u	56S 198E
G3	Mean Ordovician pole (7) X	Ol-u	25S 202E
G4	Mean Ordovician pole (7) Y	Ol-u	3S 182E

1. Van Alstine and Gillet (1979); 2. Gillet and Van Alstine (1979); 3. Patterson et al. (1978), 4. Spall (1970); 5. Van der Voo et al. (1976); 6. Scotese et al. (1979); 7. Morel and Irving (1978).

340 (Mackenzie igneous episode), which are contemporaneous (Patchett et al., 1978). Position C is very close to reconstructions previously presented by various authors (Ueno et al., 1975; Poorter, 1976b; Patchett and Bylund, 1977). The Grenville track is assumed here to reflect the time period 800—1000 Ma, consistent with K-Ar mineral ages for the Grenville Province (McElhinny and McWilliams, 1977).

The inferred transition from position B to C required a considerable

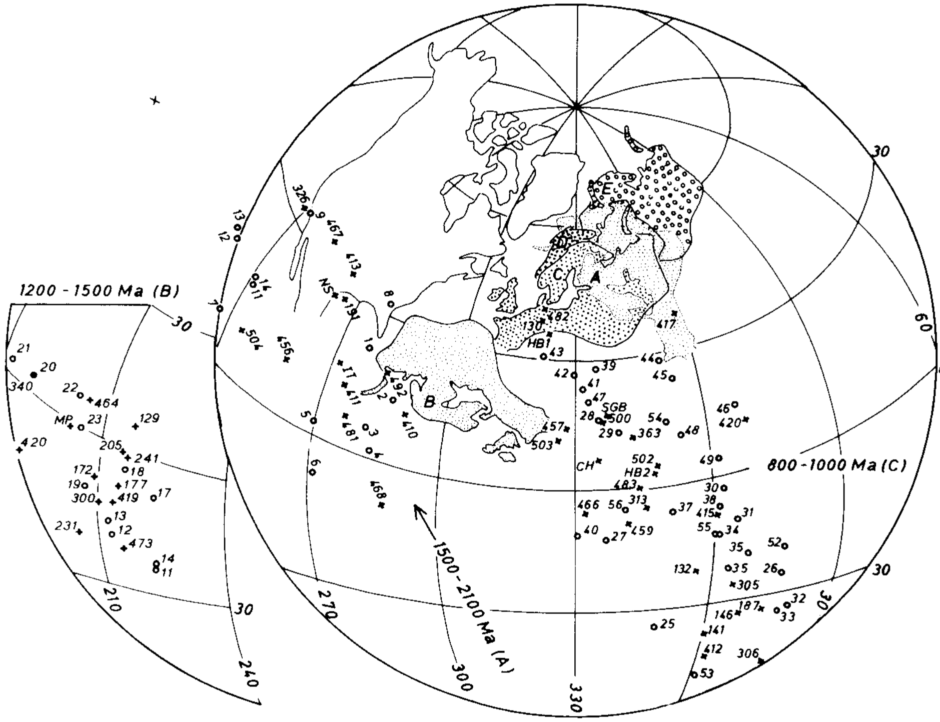

Fig. 24-6. Reconstruction of the Balto-Russian plate relative to Laurentia; see text for explanation.
A = 1500—2100 Ma reconstruction; B = 1200—1500 Ma reconstruction; C = 800—1000 Ma reconstruction; D = Devonian reconstruction (Scotese et al., 1979); E = Jurassic reconstruction (Sclater et al., 1977).
Pole positions for the intervals A, B and C: crosses Laurentia (3-digit numbers refer to Irving and Hastie, 1975; others from McElhinny and Cowley, 1978): *MP* = 15.162; *IT* = 15.197; *HB1* = 15.157; *HB2* = 15.159; *CH* = 15.172; *SGB*: Park and Roy (1979); *NS*: Beckmann et al. (1977); circles Baltica (numbers as in Table 24-I) after respective rotations.

amount of tectonic reorganization between 1200 and 1000 Ma ago, which obviously is related to the Grenvillian—Sveconorwegian orogeny. The onset of the Grenvillian orogeny was preceded by the widespread igneous activity expressed by the Mackenzie and late-Jotnian episodes, possibly initiated by regional tension due to rifting of the Baltic Shield from its B position.

The post-Sveconorwegian position C (Fig. 24-6) of the Baltic Shield relative to Laurentia is very similar to the Devonian reconstruction D made by Scotese et al. (1979). These authors argue that the ultimate position E was attained in the Carboniferous. It is tempting to believe that the Baltic Shield retained its position relative close to North America after the Grenvillian—Sveconorwegian orogeny, except for a small anti-clockwise rotation coinciding with the Caledonian orogeny, until the Devonian. Unfortunately,

TABLE 24-IV

Euler rotations applied to obtain reconstructions of the Balto-Russian Shield (positions A—E in Fig. 24-6) and the position of Siberia and Gondwana (Fig. 24-7). All rotations are with respect to North America. Gondwana reassembled according to Scotese et al. (1979) with respect to Africa

		Latitude (°)	Longitude (°)	Angle of rotation (°)
Baltica	1500—2100 Ma (A)	33.6S	197.2E	− 57.5
Baltica	1200—1500 Ma (B)	14.4N	354.0E	+ 94.6
Baltica	800—1000 Ma (C)	75.7S	99.7E	+ 58.0
Baltica	Devonian (D) (1)	69.4S	42.8E	+ 41.3
Baltica	Jurassic (E) (2)	84.4N	146.6E	− 29.2
Siberia	late Precambrian	10.3S	47.6E	+ 78.6
Gondwana (Africa)	late Precambrian	25.7S	52.3E	+ 118.5

palaeomagnetic data from Baltica are not available and those from North America are too scarce to support this idea at least for the period from about 800 to 640 Ma.

The coherence of the Eocambrian poles 13 and 14 from Armorica and 20 and 22 from Baltica (Fig. 24-5; Table 24-II) and likewise the coherence of lower Cambrian poles from Armorica (15), the Ural Mountains (24) and Wales (11) point to a contiguous European plate during the Eocambrian—Cambrian period. A recent study of the Caerfai Bay Shales in Wales (pertaining to pole 11) yielded a pole position (19°N, 187°E) not far from pole 11, but closer to lower Ordovician pole positions (pole 12) and a lower Ordovician age was suggested accordingly (Claesson and Turner, 1980). The position of the Russian Platform relative to the remainder of Europe cannot be revealed on the basis of the only available pole 23 (Fig. 24-5), which is not close to any other coeval pole from Europe.

The North American and European pole positions show a reasonable correlation for the Eocambrian—Cambrian interval (Fig. 24-5); applying the rotation related to the Devonian position on the position of the European poles the correlation is somewhat better than if the rotation related to the post-Sveconorwegian position is applied (Table 24-IV). With the present knowledge of the APW relative to Europe, the tentative conclusion can be drawn that the European and North American plates were not far apart during the Eocambrian and Cambrian periods, decreasing the possibility of an intervening ocean of considerable extension like for instance suggested by Wright (1976), but increasing the possibility that the Baltic Shield remained close to North America after the Grenvillian orogeny. The latter contention was also made by Van der Voo et al. (1980).

In order to obtain the position of Gondwana and Siberia relative to the European plate it is preferable to compare the pole positions of Gondwana, Siberia and Laurentia for the Eocambrian—Cambrian interval instead of comparing them with the less well established European data. The reconstructions shown in Fig. 24-7 are obtained by matching the 600 and 500 Ma average pole positions of Siberia (S1 and S2) and Gondwana (G1 and G2) from Morel and Irving (1978) with the coeval average pole positions of Laurentia L1—7 and L11—17, respectively. The related poles and angles of rotation are listed in Table 24-IV. The resulting positions of Gondwana and Siberia relative to Laurentia are shown in Fig. 24-7. A similar reconstruction was proposed by Scotese et al. (1979) for the late Cambrian. The position of the Balto-Russian plate, however, is different in being retained in its relative position between 800 and 450 Ma (Fig. 24-7), except for some rotation. Scotese et al. (1979) based the Cambrian reconstruction of the Baltic Shield southeast of Laurentia on a poorly documented pole for Baltica (Bergström, 1977). The reconstruction in Fig. 24-7, where Siberia lies with its present north coast to the south, shows an overlap with Southern Europe. If it is assumed that Siberia was situated somewhat more to the southeast, juxtaposed to the southern margin of Europe, it appears

Fig. 24-7. Late Precambrian/Cambrian position of Siberia and Gondwana with respect to North America (present position). Balto-Russian plate as in Fig. 24-6 (position C dashed, position D solid). Crosses delimit the Siberian plate according to the reconstruction described in the text. Dashed and solid outline after adaption to position C and D, respectively, of the Balto-Russian plate. Gondwana is reassembled according to Scotese et al. (1979); positions of Kazakhstania and China are omitted.

likely that the Cadomian and Baikalian belts are situated on both sides of the supposed suture. These belts are both of Eocambrian/Cambrian age and are possibly related to interaction of the European and Siberian plates.

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Chapter 25

PALAEOMAGNETISM OF THE BALTIC SHIELD — IMPLICATIONS FOR PRECAMBRIAN TECTONICS

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ABSTRACT

Palaeomagnetic data of the Precambrian (age 2700—800 Ma) rocks from the Baltic (Fennoscandian) Shield are reviewed. In the light of new palaeomagnetic pole positions and radiometric age data, a revised apparent polar wander path (APWP) is presented for the Baltic Shield. Apart from the Sveconorwegian (1000—800 Ma ago) results there are no major discrepancies in the pole positions of widely separated rock units, suggesting that the Baltic Shield has behaved as a coherent unit since early Precambrian times. The shape of the APW path of the Baltic Shield differs from that of the Laurentian Shield, indicating independent drift of the shields at least during part of the Precambrian. The dissimilar palaeolatitude curves of the two shields support this idea. However, during the Svecokarelian/Hudsonian orogeny (age 1900 Ma) and during the Jotnian/MacKenzie magmatic interval (1300—1200 Ma ago), the two shields had similar palaeolatitudes and may have been in juxtaposition. A collision of the two shields of these times could have triggered the Svecokarelian/Hudsonian orogenies 1900 Ma ago and the global rifting and magmatic episode (Jotnian/MacKenzie) about 1250 Ma ago. An intimate connection between the Shields was evident also during the Grenville/Sveconorwegian orogeny. This is suggested by palaeomagnetic and geological data.

The APW speed (c. 0.35°/Ma) of the Baltic Shield during the Precambrian is significantly lower than that of the Laurentian Shield (0.50°/Ma). Remanent magnetization with "normal" polarity is dominant in the Baltic Shield.

INTRODUCTION

Palaeomagnetic data of Precambrian rocks can be used to analyze whether pre-Mesozoic horizontal motions have taken place within or between different shields. Relative movements between lithospheric units are indicated by differences in the APW paths of the units. If the different provinces of a shield show different APW paths, the plate tectonic model can be used to describe the evolution of the shield (Piper et al., 1973; Cavanaugh and Seyfert, 1977). However, if the APW curves are similar, the ensialic within-shield tectonics is the proper model (McElhinny and McWilliams, 1977). Dissimilar APW curves of different shields indicate that the shields have drifted independently and emphasize that the opening and closing of oceans is a recurrent feature in the earth's history (Wilson, 1966; Spall, 1973).

In order to find out whether relative motions of lithospheric units have taken place, accurate APW paths are needed for each shield. In this paper we present a new APW path for the Baltic Shield which supersedes the simpler paths previously proposed (e.g. Neuvonen, 1970; Spall, 1973; Poorter, 1975). The new APW curve is based on all the available palaeomagnetic data of the Baltic and Ukrainian Shields. The data include more than ten new palaeopole determinations, all of which have radiometric age control (Neuvonen et al., 1981; Pesonen et al., in prep.; Pesonen and Suominen, in prep.; Pesonen, in prep.; O. Kouvo, pers. comm., 1980). The APW curve, which covers the age interval 2700—800 Ma, is compared with that of the Laurentian Shield. Finally, the palaeolatitude curves of the Baltic and Laurentian Shields are compared and their relevance to Precambrian tectonics is discussed.

GEOLOGICAL SETTING

The Baltic Shield (Fig. 25-1) can be subdivided into five different blocks: (1) the Pre-Karelian basement, (2) the Granulite Complex of Finnish Lapland, (3) The Svecokarelian belt, (4) the Sveconorwegian province and (5) the Caledonian belt (Phanerozoic) which is not discussed in this paper (see Poorter, this volume, Chapter 24).

The palaeomagnetic data are arranged into groups according to the provinces mentioned. In addition, two major post-orogenic or anorogenic events involving igneous activity are recognized within the Svecokarelian belt: the sub-Jotnian (1650—1350 Ma ago) and the Jotnian (1300—1200 Ma ago). These magmatic intervals are treated as individual groups, because palaeopoles of these rocks are of crucial importance in defining the APWP of the Baltic Shield. The sampling areas are shown in Fig. 1 and are summarized in Table 25-I.

The pre-Karelian (early Precambrian) basement in eastern Finland, Sweden and the northwestern USSR is composed of granites, gneisses and supracrustal rocks with U-Pb ages generally older than 2500 Ma (Kouvo, 1976; Gaál et al., 1978). The rocks of this area were, however, greatly affected by deformation and metamorphism during the Svecokarelian orogeny (1900 Ma ago). The early Karelian rocks (age 2400—2000 Ma), which overlie the basement in its marginal area, are grouped together with the pre-Karelian data. Thirteen palaeopoles belong to this group (Table 25-I).

The infracrustal rocks of the Granulite Complex form an arcuate zone in northern Finland (Fig. 25-1). This province has suffered low- to high-grade metamorphism several times, as shown by radiometric ages which appear to peak around 2500 Ma, 2100 Ma and 1900 Ma (Meriläinen, 1976). It is palaeomagnetically interesting that the last major metamorphism about 1900 Ma ago coincides with the Svecokarelian orogeny. Two palaeomagnetic poles are available from this province (Fig. 25-1; Table 25-I).

The Svecokarelian belt consists of various types of infracrustal and

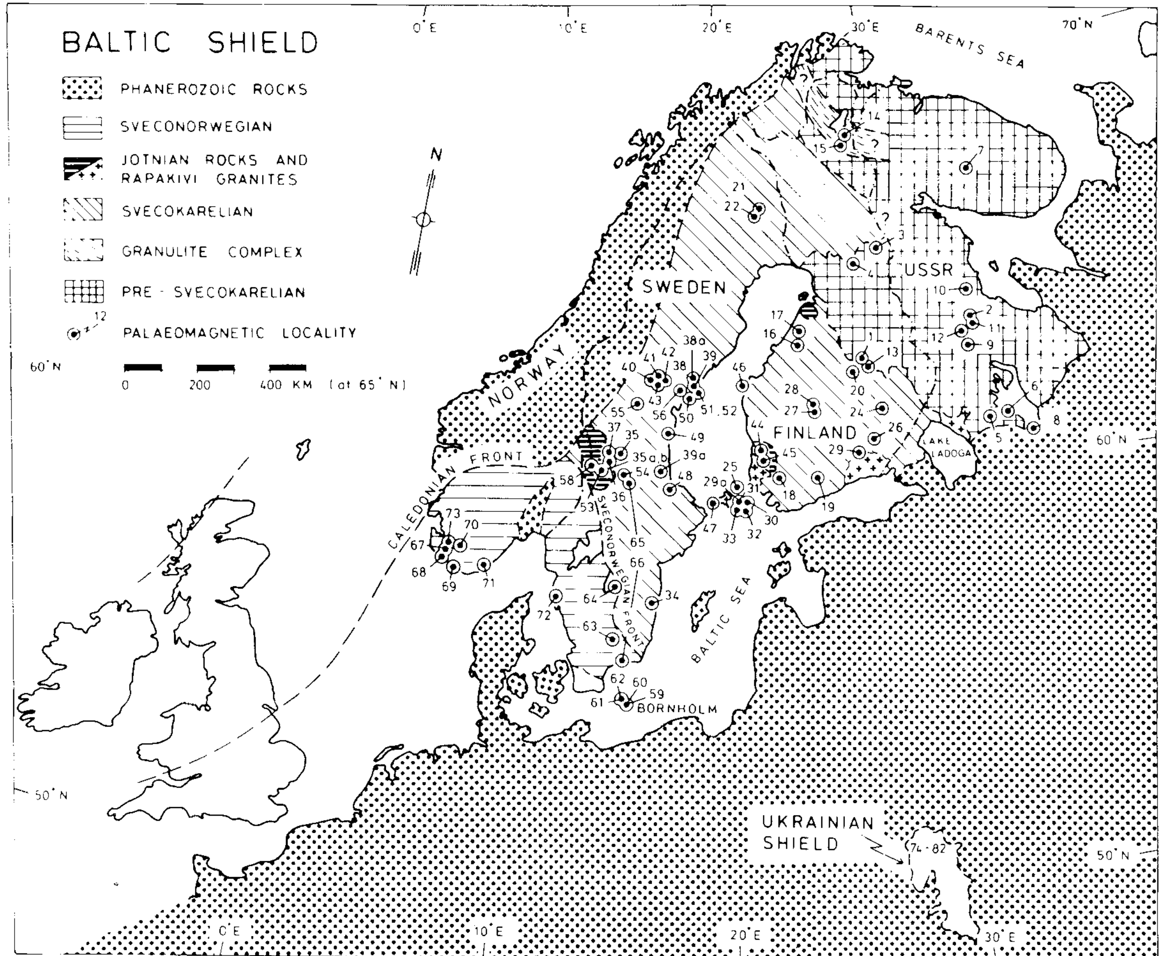

Fig. 25-I. Simplified geologic map of the Baltic Shield showing the five main provinces and the magmatic episodes. The continuation of the Granulite Complex across the Finnish border is not known is detail. The pre-Svecokarelian basement also includes both older (age > 2900 Ma) and younger (presumably Jatulian or Svecokarelian) inliers which are not shown. Palaeomagnetic sampling sites are marked with numbered circles (see Table 25-I).

Precambrian (2700—800 Ma) palaeomagnetic poles from the Baltic and Ukrainian Shields

Symbol	Rock unit	Country	S/B/N	Age	Method	Palaeomagnetic pole Lat. (N), Long. (E)	dp, dm (A_{95})	Pol.	Cat.	Refs./Notes
I. Pre-Svecokarelian										
1	Nilsiä diorites	F	1/8/64	2684	c	63.9, 313.0	7, 8	N	A	(1), (2)
2	Jatulian beds and lavas	SU	1/12/38	1870—1610		58.0, 238.0	6, 12	N	A	(3)
3	Jatulian dikes and greenstones	F	1/4/47	2150	c	47.2, 233.4	11, 18	N	A	(4), (5)/b
4	Syöte gabbro	F	1/1/11	2440	c	35.7, 221.2	2, 4	N	B	(4), (6)/b
5	Pukhta—Pedaselsk group	SU	1/-/57	1950—1850	a, d	44.0, 231.0	3, 5	N, R	A	(7)
6	Shoksha group	SU	1/-/29	1950—1850	a, d	54.0, 240.0	3, 5	N	A	(7)
7	Kola gabbro-norites	SU	1/-/12	1880—1820		40.0, 223.0	9, 16	N	A	(7)
8	Onega gabbro-diabases	SU	1/-/15	1800—1600		32.0, 230.0	1, 2	N	A	(7)
9	Central Karelia dikes	SU	1/2/12	1870—1610		49.0, 235.0	5, 7	N	A	(7)
10	Northern Karelia sandstones	SU	1/2/7	1870—1610		65.0, 217.0	8, 10	N	B	(7)
11	Central Karelia sandstones	SU	1/5/5	—		57.0, 239.0	15, 21	N	B	(7)
12	Central Karelia sandstones	SU	1/3/14	—		62.0, 264.0	5, 7	N	B	(7)
13	Varpaisjärvi dikes	F	1/11/17	—		47.0, 188.0	4, 6	N	B	(1), (2)
II. Granulite Complex										
14	Akujärvi quartz diorite	F	1/3/10	1925	c	41.3, 245.5	4, 7	N	A	(8), (9)
15	Laanila dikes	F	1/3/18	—		— 3.5, 218.1	5, 8	N	B	(8)
III. Svecokarelian										
16	Ylivieska gabbro	F	1/5/42	1885	c	43.3, 242.4	5, 8	N	A	(10)
17	Pohjanmaa intrusives	F	1/5/45	~ 1900		37.9, 239.1	8, 15	N	B	(10)
18	South Finland intrusives	F	1/7/82	~ 1900		39.7, 242.6	10, 18	N	B	(11)
19	Hyvinkää gabbro	F	1/11/11	1875	c	44.1, 246.2	5, 9	N	A	(12), (13)
20	Pielavesi gabbro	F	1/5/20	~ 1900		35.9, 238.2	2, 4	N	B	(8), (14)
21	Tärendö gabbro	S	1/5/17	2000—1800		44.6, 228.7	10, 17	N	A	(15)
22	Tärendö silicic rocks	S	1/3/12	2000—1800		42.0, 248.0	—	N	B	(15)
23	Mean synorogenic intrusives	F, S	7/41/229	~ 1900		41.2, 240.8	(4.3)	N	A	—/(d)
24	Rantasalmi lamprophyres	F	1/11/24	1837	c	49.0, 226.0	2, 3	N	A	(1), (2)
25	Åva intrusives	F	1/15/-	1815—1803	c	41.0, 169.0	9, 17	N	B	(16)
26	Luontarivesi granodiorite	F	1/25/211	1790—1765	c	67.0, 198.0	10, 17	N	A	(17), (18)
27	Keuruu dikes	F	1/11/34	~ 1880	c	45.4, 218.0	4, 7	N	A	(19), (2)/(c)
28	Keuruu dikes	F	1/3/12	⩽ 1900		26.3, 257.8	2, 3	R	B	(19), (2)/(c)
Sub-Jotnian igneous interval (~ 1650—1320 Ma)										
29	Kuisaari dolerite	F	1/8/84	1650	c, e	22.0, 190.0	7, 14	R	A	(17), (20)
29a	Åland rapakivi	F	1/1/6	1659—1589	c	19.5, 194.4	15, 18	N	B	(21), (20)
30	Kumlinge—Brändö dikes	F	1/7/25	1602—1556	b	12.2, 182.0	5, 9	R	A	(22)
31	Föglö—Sottunga dikes	F	1/6/25	1523—1518	b, c	27.9, 187.5	6, 13	N	A	(22)
32	Kumlinge dikes	F	1/4/17	1602—1556	b	14.3, 190.1	8, 14	R	A	(16)/(b)
33	Föglö dikes	F	1/8/56	1523—1518	b, c	31.3, 186.5	6, 12	N	A	(23)
34	Loftahammar gabbro	S	1/5/26	1694—1460	b	23.0, 179.0	2, 5	N	A	(24), (25)
35	Dala porphyries	S	1/2/11	806— 701/1570	a/b	23.0, 184.0	—	N	A	(26), (27)
35a	Dala metasediments	S	1/6/27	—		33.0, 177.0	7, 14	M	B	(28)
35b	Dala volcanics	S	1/8/46	—		27.0, 189.0	5, 11	M	B	(28)
36	Öje basalts	S	1/2/11	931— 745/1405	a/b	32.2, 186.0	5, 10	N	A	(26), (27)

Table 25-I (Continued)

Symbol	Rock unit	Country	S/B/N	Age	Methods	Palaeomagnetic pole Lat. (N), Long. (E)	dp, dm (A_{95})	Pol.	Cat.	Refs. Notes
37	Dala dolerites	S	1/4/19	—		23.0, 178.0	21, 40	N	B	(26), (27)
38	Nordingrå gabbro-anorthosite	S	1/7/39	⩾ 1385	e	34.1, 135.7	8, 14	R	A	(28)/(b)
38a	Nordingra gabbro-anorthosite	S	1/12/69	⩾ 1385		38.5, 141.6	4, 7	N	A	(28)/(b)
39	Nordingra granite	S	1/1/28	1415		33.0, 149.0	8, 14	R	A	(28)/(b)
39a	Gävle granite	S	1/1/12	—		35.0, 165.0	7, 14	N	B	(28)
40	Ragunda intrusive rocks	S	1/11/-	~ 1320		51.3, 169.5	7, 11	N	A	(25)/(b)
41	Ragunda intrusive rocks	S	1/4/-	~ 1320		53.2, 163.7	15, 23	R	B	(25)/(b)
42	Ragunda dolerites	S	1/3/-	1320—1250		9.2, 195.1	19, 34	N	B	(25)/(b)
43	Ragunda dolerites	S	1/9/-	1320—1250		16.2, 194.2	10, 20	R	B	(25)/(b)
Jotnian igneous interval (~ 1300—1150 Ma)										
44	Satakunta sandstone	S	1/10/40	~ 1300	a	3.0, 180.0	8, 12	N	B	(29), (30)
45	Satakunta dolerites	F	1/18/62	1236—1225	b	2.0, 158.0	3, 5	N	A	(31), (32)
				1263	c					
46	Vaasa dolerite dikes	F	1/15/58	1225/1270	b/c	7.0, 164.0	3, 6	N	A	(33), (32)
47	Märket dolerite dikes	F	1/8/58	1270	b, c	− 5.9, 145.5	7, 11	N	A	(23), (32)
48	Gävle dolerites	S	1/2/15	1245		8.0, 150.0	2, 4	N	A	(24), (35)
49	Gnarp dolerites	S	1/1/6	1245	b	− 11.0, 159.0	7. 10	N	B	(24), (35)
50	Nordingra dolerites	S	1/2/12	1245	b	− 5.0, 158.0	5, 9	N	A	(24), (35)
50a	Nordingrå baked rocks	S	1/13/80	1250	e	− 2.3, 157.4	3, 5	N	A	(28)
51	Ulvö dolerites	S	1/-/32	1245	a	1.0, 161.0	5, 9	N	A	(36)
52	Ulvö dolerites	S	1/-/20	1245	a	− 4.0, 153.0	4, 6	N	A	(37)
53	Ålvdalsasen dolerite sills	S	—	1231	b	− 1.4, 148.6	5, 6	N	B	(35)
54	Ålvho dolerite sills	S	—	1290—1215	b	17.3, 170.2	9, 13	N	B	(35)
55	Sundsjö—Gimån dolerite dikes	S	—	1229—1156	b	2.9, 147.4	6, 11	N	B	(35)
56	Väst-Norrland dolerites	S	1/43/252	1245	a	− 7.5, 156.5	2, 3	N	A	(38)
57	Mean Jotnian dolerites	S, F	13/-/-	~ 1250		1.0, 156.0	(5.3)	N	A	—/(d)
58	Dala dolerites and basalts	S	1/1/5	~ 1200		− 10.0, 115.0	15, 27	N	B	(39)
59	Vasea dike	DB	1/1/11	~ 1200		− 16.0, 127.8	6, 10	N	B	(40)/(a)
60	Listed dike	DB	1/1/32	~ 1000		13.8, 250.1	2, 4	N	B	(40)/(a)
61	Bolshavn dike	DB	1/1/6	~ 1000		20.7, 223.6	7, 14	N	B	(40)/(a)
62	Kjeldsea dike	DB	1/3/11	~ 1100—1000		7.0, 179.0	6, 10	R	B	(41), (40)/(a)
IV. Sveconorwegian (~ 1100—800 Ma)										
63	Swedish hyperites	S	1/2/10	1516— 886	a	− 6.0, 240.0	(18.0)	R	A	(26), (27)/(b)
64	Swedish hyperites	S	1/3/60	1573— 781	a	− 19.8, 213.3	(24.1)	R	A	(26), (27)/(b)
65	Falun—Karlshamn dolerites	S	1/4/28	987— 890	b	− 9.7, 233.4	(32.1)	R	A	(42)/(b)
66	Falun—Karlshamn dolerites	S	1/4/24	1005— 890	b	13.7, 243.8	(15.1)	N	A	(42)/(b)
67	Mean Egersund dolerites	N	2/10/88	880— 663	a, e	− 25.0, 231.5	(13.3)	R	A	(43), (44)/(d)
68	Mean Egersund anörthosites	N	2/9/-	~ 1000— 850	a	− 39.6, 203.6	(16.1)	N	A	(45), (46)/(d)
69	Egersund farsundites	N	1/7/35	920— 900	a	− 43.3, 194.0	9, 10	N	A	(47)
70	Hunnedalen dikes	N	1/8/37	⩽ 950— 850	a, b	− 34.0, 208.0	9, 10	N	A	(43)/(d)
71	Bamble—Kongsberg basement	N	1/6/48	1120— 975	a, b, c	− 8.0, 208.0	6, 8	N, R	B	(34)/(d)
72	Tuve dolerite	S	1/1/10	—		− 17.0, 238.9	2, 3	R	B	(48)
73	Mean Rogaland basement	N	1/3/19	950— 850	a, b	− 36.0, 233.0	10, 12	N	A	(43)/(d)
V. Ukrainian Shield										
74	Bug—Bodolsky complex	UR	—	2300—2000	a, b, c	54.0, 286.0		R	B	(49), (50), (51)

Table 25-I (Continued)

Symbol	Rock unit	Country	S/B/N	Age	Methods	Palaeomagnetic pole Lat. (N), Long. (E)	dp, dm (A_{95})	Pol.	Cat.	Refs. Notes
75	Ingulets complex	UR	—	2000—1700	a, b, c	64.0, 325.0		R	B	(49), (50), (51)
76	Ingulets complex	UR	—	2000—1700	a, b, c	57.0, 316.0		R	B	(49), (50), (51)
77	Bokovyansk intrusion	UR	—	1700—1500	a, b, c	52.0, 301.0		R	B	(49), (50), (51)
78	Takova granite	UR	—	1600—1400		51.0, 326.0		R	B	(49), (50), (51)
79	Turchinki gabbro	UR	—	1400—1200		12.0, 212.0		R	B	(49), (50), (51)
80	Uman granite	UR	—	1750	a, b, c	26.0, 240.0		R	B	(49), (50), (51)
81	Korosten complex	UR	—	1750	a, b, c	24.0, 232.0		R	B	(49), (50), (51)
82	Korosten complex	UR	—	1750	a, b, c	27.0, 238.0		R	B	(49), (50), (51)

Symbols	refer to Figs. 25-1 and 25-2.
Country	DB = Bornholm (Denmark); F = Finland; N = Norway; S = Sweden; SU = Soviet Union; UR = Ukraine.
S/B/N	number of studies/sites/samples.
Age/Method	radiometric ages (Ma) recalculated, when possible, with the new decay constants (Steiger and Jäger, 1977). a = K-Ar; b = Rb-Sr; c = U-Pb; d = U-Pb-Th; e = geological evidence. For further details of the ages see references listed.
Lat. (N), Long. (E)	*latitude and longitude (degrees) of the palaeomagnetic pole.*
dp, dm	95% confidence oval of the pole (single study).
(A_{95})	95% confidence circle of the mean pole (several studies combined).
Pol.	polarity of the pole: N = normal, R = reversed (see p. 629).
Cat.	Category of the pole. A means reliable data, B less reliable in a manner outlined on p. 631.
Notes	(a) for other Bornholm poles and their interpretation models, see Abrahamsen (1977); (b) normal and reversed poles recalculated by the present authors from original data; (c) see text; (d) some recalculations performed in this study.

References: (1) Neuvonen et al. (1981); (2) O. Kouvo (pers. commun., 1980); (3) Irving and Hastie (1975); (4) Neuvonen (1975); (5) Sakko (1971), (6) Kouvo (1977); (7) McElhinny and Cowley (1977); (8) Pesonen (in prep.); (9) Meriläinen (1976); (10) Pesonen and Stigzelius (1972); (11) Neuvonen (1974); (12) R. P. Puranen (pers. commun., 1980); (13) Härme (1978); (14) Helovuori (1979); (15) Cornwell (1968); (16) Neuvonen (1970); (17) Neuvonen (1978); (18) Korsman and Lehijärvi (1973); (19) Pesonen et al. (in prep.); (20) Vaasjoki (1977); (21) Johnson (1979); (22) Pesonen and Suominen (in prep.); (23) Neuvonen and Grundström (1969); (24) Poorter (1976a); (25) Piper (1979b); (26) Mulder (1971); (27) Priem et al. (1968); (28) Piper (1980); (29) Neuvonen (1974); (30) Simonen (1980); (31) Neuvonen (1965); (32) Suominen (in prep.); (33) Neuvonen (1966); (34) Poorter (1975); (35) Patchett et al. (1978); (36) Magnusson and Larson (1977); (37) Larson and Magnusson (1976); (38) Piper (1979a); (39) Dyrelius (1970); (40) Abrahamsen (1977); (41) Schønemann (1972); (42) Patchett and Bylund (1977); (43) Poorter (1972); (44) Storetvedt and Gidskehaug (1968); (45) Hargraves and Fish (1972); (46) Murthy and Deutsch (1974); (47) Murthy and Deutsch (1975); (48) Abrahamsen (1974); (49) Kruglyakova (1961); (50) Spall (1973); (51) Semenenko et al. (1968).

supracrustal rocks which were affected by the major Svecokarelian orogeny about 1900 Ma ago (Kouvo, 1976). Thirteen palaeopoles have been obtained from this belt. The sampling localities extend from northern Sweden to southeastern Finland (Fig. 25-1).

The sub-Jotnian palaeomagnetic data consist of 20 palaeopoles located within the Svecokarelian belt in Finland and Sweden. Similarly, 14 palaeopoles are available from the Jotnian sedimentary and igneous rocks of the belt in Sweden and Finland (Table 25-I).

The youngest Precambrian tectonic province in the Baltic Shield is the Sveconorwegian block for which radiometric ages range from 1100 Ma to 800 Ma (Versteeve, 1975; Lundqvist, 1979), although older relict ages are also reported (Skiöld, 1976). These ages, especially those determined by the K-Ar method, most likely reflect the cooling and uplift of the Sveconorwegian province. The province is surrounded in the west by the Caledonian belt and in the east by the Sveconorwegian front, which appears to be a 50—100 km broad, subvertical tectonic boundary (the schistosity zone; Fig. 25-1; e.g. Patchett and Bylund, 1977). The similarity in geology and radiometric age data of the Sveconorwegian and Grenville provinces (Laurentian Shield) is striking and has been noted by several workers (e.g. Wynne-Edwards and Hasan, 1970; Zwart and Dornsiepen, 1978; *see also Baer, this volume, Chapter 14, ed.*). There are eleven palaeopole measurements from this province, including those from the schistosity zone and adjacent dyke swarms (the Falun-Karlshamn dykes).

In addition to the Baltic Shield, Precambrian bedrock is exposed as numerous minor shield inliers in Europe (e.g. Zwart and Dornsiepen, 1978). In this paper we have included the palaeomagnetic data only from the Ukrainian Shield (Fig. 25-1) from which nine poles are reported by Kruglyakova (1961; Table 25-I).

THE APWP OF THE BALTIC SHIELD

Table 25-I lists all the palaeomagnetic poles from the Baltic and Ukrainian Shields in the age interval 2700—800 Ma. Methods and results of radiometric dating are also given along with references to original studies.

The polarity has been defined so that north-seeking magnetization, which yields palaeomagnetic poles in the Pacific region (Fig. 25-2), is called "normal". The poles that would plot on the opposite hemisphere are called "reversed" and are inverted by 180 degrees before plotting. The true polarity of Precambrian poles is, however, uncertain because there are large gaps in the APW paths (Fig. 25-2), and segments of polar wander paths cannot always be reliably connected (Spall, 1973; Irving and McGlynn, 1976). Moreover, the "normal" polarity with respect to the Baltic and Laurentian Shields may differ.

In this paper we distinguish between the two possible polarities (although

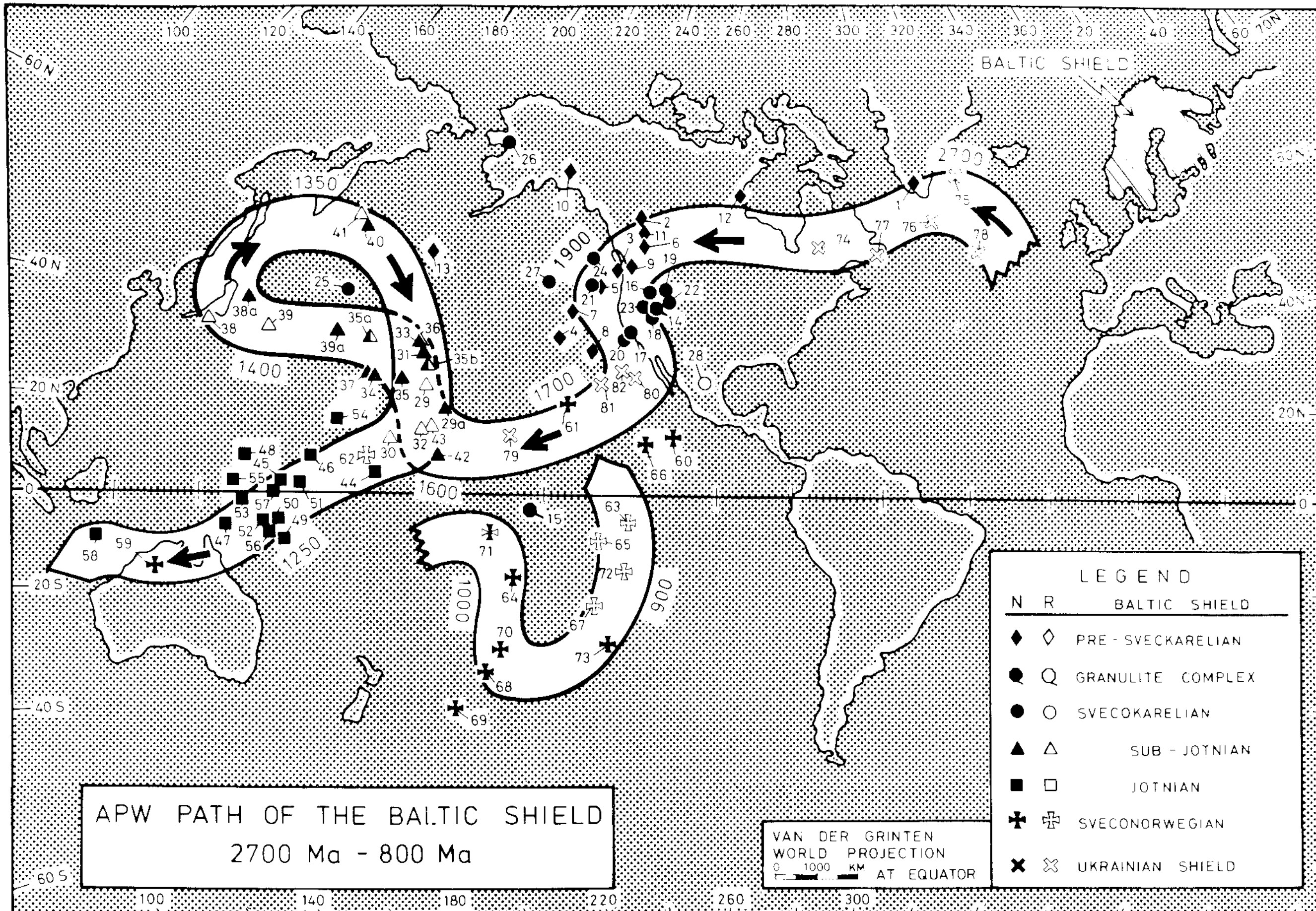

Fig. 25-2. The apparent polar wander path (APWP) for the Baltic Shield (2700—800 Ma ago). The pole positions have the same numbers as the sampling areas of Fig. 25-1. The 95% confidence limits for the poles are given in Table 25-I. Open (closed) symbols denote reversed (normal) polarity. For other symbols see legend.

this involved some recalculation), because systematic departures from the 180° symmetry of antipoles may render important clues as to unremoved secondary components, non-dipole disturbances or apparent polar wander during the polarity change (see detailed discussion in Pesonen, 1978).

Each pole in Table 25-I is characterized by a reliability category symbol, A or B. Category A includes the more reliable data for which a radiometric age is available and reported demagnetization studies show no indication of unremoved secondary components (see also Irving and Hastie, 1975; Irving and McGlynn, 1976). The majority (53%) of the data belong to category A. The remaining data (47%) are less reliable and comprise category B. A few results (about 5%), which are claimed to be unreliable by the original authors (e.g. Neuvonen, 1967), were excluded from this study.

GENERAL DESCRIPTION OF THE APWP

The simplest APW path for the Baltic Shield is a smooth swathe (Piper et al., 1973) with a width of 15° (Fig. 25-2).

The APW swathe starts from pole 1 of probable Archaean age and located in Greenland (Fig. 25-2), passes through the North American continent, makes a large hairpin turn in the Pacific (the sub-Jotnian loop) 1650—1350 Ma ago before arriving in northern Australia (poles 58—59) at the end of the Jotnian interval (1200 Ma ago). Then there is a major gap in palaeomagnetic data as no poles are available for the age interval 1200—1000 Ma. Therefore, the Sveconorwegian poles (63—73) are treated as a separate track representing the age interval 1000—800 Ma (Fig. 25-2).

The main differences between the APW path shown in Fig. 2 and the previously proposed paths are the new early Precambrian—Svecokarelian track (2700—2000 Ma ago) and the large northerly sub-Jotnian loop (1650—1350 Ma ago) (Pesonen, 1979b; Piper, 1980).

Pre/early Karelian data (2700—1900 Ma)

The only early Precambrian pole comes from dioritic rocks in the Nilsiä region of east-central Finland (Figs. 25-1 and 25-2), for which radiometric dating yields an age of 2680 Ma (Neuvonen et al., 1981). Fold or baked contact tests have not been conducted, but other evidence indicates an early Precambrian age for the remanent magnetization. The pole position differs by more than 60° from the early Karelian (2—13) and synorogenic Svecokarelian poles (16—28) with ages between 2400 Ma and 1800 Ma (Pesonen and Stigzelius, 1972; Neuvonen, 1974) but some of the Ukrainian poles (74—78) which appear to have either Archaean or early Karelian ages (2400—1700 Ma) plot in this same area (see also Semenenko et al., 1968; Spall, 1973).

The early Karelian palaeomagnetic poles (age 2400—1900 Ma) are located on the western coast of North America (poles 2, 3, 5, 6, 9 and 11). These poles appear to form a cluster which, in the APW swathe, slightly predates the subsequent Svecokarelian (1900 Ma) poles (16—24). The pre- and early Karelian poles are of "normal" polarity except for the Pukhta-Pedaselsk group (pole 5), which displays short "reversed" intervals (Katseblin, 1968). The Ukrainian poles (74—78) are all of "reversed" polarity (Fig. 25-2).

Early Karelian poles deviating from the APW swathe also occur. Poles 4, 7 and 8 (Fig. 25-2) show small deviations which may be explained by differences in the radiometric and palaeomagnetic uplift ages (e.g. Pullaiah et al., 1975; Poorter, this volume, Chapter 24). Poles 10 and 13 depart significantly from the APW path, as discussed later (p. 633).

Granulite Complex data

Two poles (14, 15) are available from the Granulite Complex (Figs. 25-1 and 25-2). The first one is obtained from the Akujärvi quartz diorite, for which radiometric dating yields an U-Pb age of 1925 Ma (Meriläinen, 1976). This remanent magnetism with normal polarity was probably recorded during the uplift and slow cooling of the quartz diorite, which took place after granulite facies metamorphism about 2000—1900 Ma ago (Pesonen, in prep.). The other pole from the Granulite Complex is derived from a swarm of olivine diabase dykes (pole 15; Pesonen, in prep.) which cut the granulites in the Laanila region. No radiometric age data are available for these dykes. The pole position (Fig. 25-2) suggests either an age of about 1700 Ma or about 900 Ma. Both of these ages are realistic on geological grounds. Diabase dykes with an age of 1730 Ma exist in the Lake Inari area about 100 km north of Laanila (Meriläinen, 1976), and igneous activity at about 1150—900 Ma is known in northern Finland (R. Lauerma, pers. commun., 1980), in Sweden (Kresten et al., 1977) and in the Kola Peninsula (T. Mutanen, pers. commun., 1980).

Svecokarelian palaeomagnetic data

In defining the APW path for the Baltic Shield, the pole for the synorogenic Svecokarelian intrusives (age 1900—1800 Ma; Neuvonen, 1974) is of crucial importance. The mean pole (23) is determined from seven rock units in Sweden and Finland. Consistency between the Swedish (21, 22) and Finnish (16—20) poles is excellent (Fig. 25-2).

Poles 27 and 28 are also of considerable importance in defining and interpreting the APW path for the Baltic Shield. These poles come from the Keuruu diabase dykes (age 1880 Ma; Pesonen et al., in prep.) in central Finland (Fig. 25-1 and 25-2). The dykes must have been intruded into an already reasonably cold crust as they have chilled margins against the synorogenic Keuruu gabbro. Furthermore, both normal and reversed polarities are present in the dykes (Pesonen et al., in prep.) in contrast with the synorogenic rocks, which are all of normal polarity (Fig. 25-2; Table 25-I). The reversal in the Keuruu dykes is asymmetric (not 180°). Baked contact tests and detailed demagnetization studies suggest that the asymmetry is not caused by secondary remanence. For this reason the poles (27, 29) have not been averaged. The APW interpretation is, however, difficult as age data are available only for a normally magnetized dyke. We will return to the problem of asymmetric reversals on p. 637.

Poles 24--26 are significantly younger (age about 1815 Ma) than the synorogenic intrusives (age about 1900 Ma; Table 25-I). The pole positions also differ from those of the synorogenic intrusives (Fig. 25-3). Accordingly, we consider that a prominent anti-clockwise polar loop is possible during the age interval 1850—1750 Ma as shown in Fig. 25-3. Because it is based on only four poles, this post-orogenic APW loop must be regarded as tentative.

Sub-Jotnian loop

A gap of almost 60° exists in the APW path in the age interval 1750—1650 Ma.

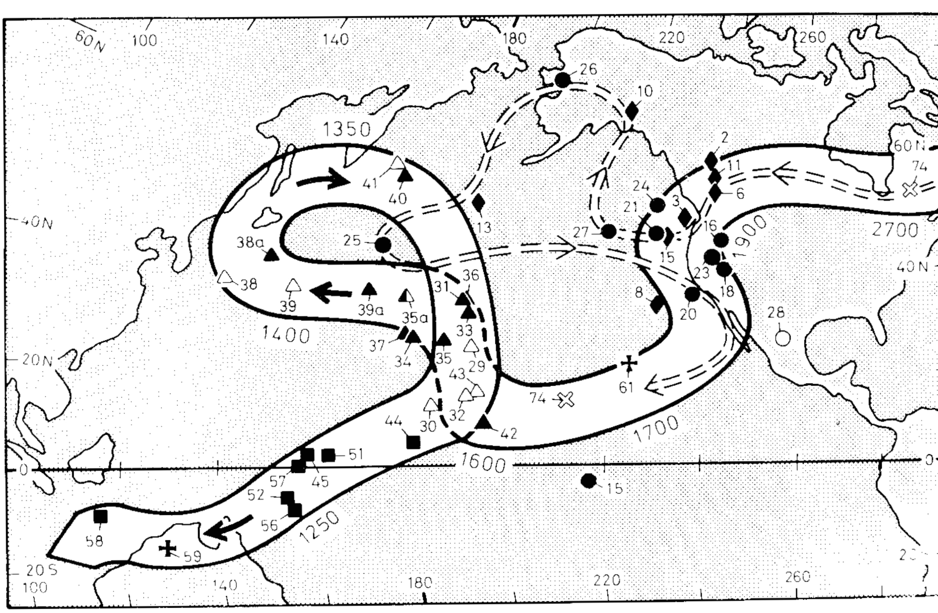

Fig. 25-3. An alternative APW path for the Baltic Shield. A more complex path is shown as a dashed line superimposed on the APW path of Fig. 25-2. Symbols as in Fig. 25-2.

Only a group of Ukrainian poles (80—82) and possibly poles 15 and 61 fit this part of the APW track (Fig. 2).

The oldest sub-Jotnian results come from the Kuisaari dolerite (29) (Neuvonen, 1978), the Åland Rapakivi massif (29a)(Johnson, 1979) and the Kumlinge dykes in the Åland archipelago (30, 32)(Neuvonen and Grundström, 1969; Pesonen and Suominen, in prep.). The Kuisaari dolerite was magnetized at the time of Ahvenisto rapakivi intrusion (1650 Ma ago; Vaasjoki, 1977), as verified by positive baked contact tests (Neuvonen, 1978). The Kumlinge dykes were magnetized during the intrusion of the dykes, which took place somewhat later (about 1600 Ma ago) than the emplacement of the Åland rapakivi massif (about 1650 Ma ago; Vaasjoki, 1977; Suominen, in prep.). This is supported by baked contact tests and by the reversed polarity of the dykes in contrast with the normal polarity of the Åland rapakivi massif (Fig. 25-2; Table 25-I; Pesonen and Suominen, in prep.).

Poles 33—37 form a tight cluster and represent the age interval of about 1570—1500 Ma. They are all of normal polarity (Neuvonen and Grundström, 1969; Mulder, 1971; Poorter, 1976a; Piper, 1980; Pesonen and Suominen, in prep.). In this group the Swedish poles (34—37) agree well with the Finnish poles (31, 33), demonstrating internal consistency of the Baltic Shield palaeomagnetic data. The depth of the sub-Jotnian loop is established by the poles of the Gävle granite (pole 39a), the Nordingrå intrusions (poles 38, 38a and 39; ages about 1415—1385 Ma) and the Ragunda intrusion (age about 1320 Ma)(poles 40, 41; Piper, 1979b, 1980). The return part of the loop is defined by poles 42 and 43 of the post-Ragunda dolerites (age $\leqslant$ 1320 Ma), which cut the Ragunda intrusion (Piper, 1979b).

Jotnian results

The palaeomagnetic poles (45—47) of the Jotnian magmatic episode (age 1300—1200 Ma) form a cluster situated slightly to the northeast of Australia (Fig. 25-2; Neuvonen, 1965, 1966, 1970; Neuvonen and Grundström, 1969; Poorter, 1976a; Larson and Magnusson, 1976; Magnusson and Larson, 1977; Patchett et al., 1978; Piper, 1979a, b, 1980). The poles of the Ragunda dykes (age $\leqslant$ 1320 Ma), Jotnian sandstone (age 1300 Ma) and Jotnian dyke rocks and sills (age 1250 Ma) all lie in the APW swathe in accordance with geological and geochronological data (Table 25-I). Poles 58 (Dalå dolerites; Dyrelius, 1970) and 59 (Vaseå dyke; Abrahamsen, 1977) deviate from the other Jotnian poles to the younger side on the APW swathe. This deviation, associated with category B poles, could result from causes other than apparent polar wander. The deviation is, however, in accordance with the age (1200 Ma) of these poles which suggests them to be younger than the rest of the Jotnian poles (Fig. 25-2). The mean pole of the Swedish Jotnian dolerites (Lat. = $-$ 0.2° N, Long. = 156.0° E, A_{95} = 6.3°) is in excellent agreement with the corresponding Finnish pole (Lat. = 5.0° N, Long. = 155.8° E, A_{95} = 5.0°). All the Jotnian poles are of "normal" polarity, consistent with the corresponding data from MacKenzie intrusive rocks of North America (Patchett et al., 1978).

The second major hiatus in the APW swathe is the age interval 1200—1000 Ma. Therefore, the path between the Jotnian poles and the Sveconorwegian poles is left open (e.g. Abrahamsen, 1977).

The Sveconorwegian loop

The oldest Sveconorwegian poles are derived from highly metamorphosed basement rocks (71, 73; Poorter, 1972, 1975) and from synorogenic anorthosite intrusives (68, 69; Hargraves and Fish, 1972; Murthy and Deutsch, 1974, 1975). These rocks have been magnetized during their slow cooling and uplift after the Sveconorwegian metamorphism. Their magnetization age can be only roughly fixed between the time of metamorphism and final uplift to the approximate time interval 1000—900 Ma ago.

Problems arise not only in age definitions but also in interpretation of complex magnetizations. Multicomponent remanence is indicated by the non-linear vector diagrams and the asymmetric reversals in these rocks (Mulder, 1971; Poorter, 1972, 1975).

The younger arm of the Sveconorwegian loop is composed of poles from several dyke swarms, some of which are located in the Sveconorwegian province (Hunnedalen (70), Tuve (72) and Egersund (67) dolerites; Storetvedt and Gidskehaug, 1968; Poorter, 1972; Abrahamsen, 1974). Others are situated either in the schistosity zone (Swedish hyperites (63—64; Mulder, 1971) or in the adjacent terrain (Falun-Karlshamn dykes (65—66), Patchett et al., 1978).

The Hunnedalen dykes probably record an uplift magnetization as their pole position is very close to that of the anorthosites (68) and farsundites (69; Fig. 25-2). The Egersund and Tuve dolerites record intrusion magnetization. The Egersund dykes are clearly younger (880—663 Ma) than the basement (ages 1000—900 Ma; Table 25-I). The dykes also have chilled margins and opposite polarity compared with the basement rocks (Storetvedt and Gidskehaug, 1968).

The complex magnetization history of the Swedish hyperites (63—64) is manifested by their multicomponent remanence (Mulder, 1971; Poorter, 1972). Results of baked contact tests are not available, and it is not known whether the magnetization reflects a Sveconorwegian overprint or the true intrusion age. The presence of two polarites (with strong asymmetry) supports the latter idea but the wide K-Ar age range (1573—781 Ma; Priem et al., 1968) favours the former alternative.

Positive results were obtained from baked contact tests performed on radiometric and palaeomagnetic data of the Karlshamn-Falun dykes (Patchett et al., 1978). This suggests that the dykes were magnetized during their intrusion into a cold Svecokarelian terrain just outside the schistosity zone. These dykes give the first palaeopoles of Sveconorwegian age (1000—800 Ma), obtained from rocks outside the Sveconorwegian block. The reversal in these dykes is asymmetric, a feature which appears to be characteristic of the Sveconorwegian reversals (Fig. 25-2).

Four poles (59—62) are available from the dolerite dykes of the Bornholm island (Schønemann, 1972; Abrahamsen, 1977). Their magnetism is either of Jotnian (poles 59, 62) or of Sveconorwegian (poles 60, 61) age. The geology of Bornholm is, however, very complex as the island lies close to the Sveconorwegian front and to the Polish Trough (Fig. 25-1)(Abrahamsen, 1977; Pozaryski and Brochwicz-Lewinski, 1978). Hence, the Bornholm poles have been included in category B.

PROBLEMS IN THE BALTIC SHIELD APWP

Before we attempt to interpret the palaeomagnetic results of the Baltic Shield in terms of various tectonic models, it is worthwhile to consider causes other than plate motion which can affect the polar paths of Figs. 25-2 and 25-3. Errors in sampling and measurement, insufficiently eliminated secular variation and discrepancies between radiometric and magnetization ages can cause considerable scatter of palaeopoles (*see also Irving and McGlynn, this volume, Chapter 23, and McWilliams, this volume, Chapter 26, ed.*).

Secondary components

The Precambrian rocks of the Baltic Shield have been affected by several metamorphic events (e.g. Kouvo, 1976; Meriläinen, 1976; Gaál et al., 1978), each of which has added another component to the magnetic and radiometric record carried by the rocks. A single metamorphic event involving total remagnetization permits the treatment of the new characteristic remanence by standard methods, and the resulting radiometric and palaeomagnetic ages relate only to the times of metamorphism. In the case of partial remagnetization the new NRM consists of at least two components. Recent palaeomagnetic studies in North America (e.g. Buchan and Dunlop, 1976; McGlynn and Irving, 1978) demonstrate that multicomponent NRMs are common in Precambrian metamorphic rocks. The secondary components can be isolated by modern demagnetization techniques (Roy and Lapointe, 1978), vector subtraction (Buchan and Dunlop, 1976) great circle methods (Halls, 1976), fully executed baked contact tests and palaeointensity measurements (Pesonen, 1978, 1979a). Such investigations of multicomponent NRMs have considerably modified the APWP of North America (Fig. 25-5).

There are no detailed studies on rocks from the Baltic Shield where secondary components have been properly separated. Yet in many cases the existence of secondary remanence is clearly indicated by non-linear vector diagrams (Mulder, 1971; Poorter, 1972, 1975) or asymmetric reversals (e.g. Mulder, 1971; Patchett et al., 1978; Pesonen et al., in prep.). The secondary overprints due to Svecokarelian, Sveconorwegian or Caledonian orogenies should be interesting subjects for future studies.

Priem et al. (1968) suggested that the sub-Jotnian palaeopoles represent remagnetization due to the Caledonian orogeny because these poles lie very close to the Palaeozoic poles of Europe. The Caledonian remagnetization hypothesis can, however, be rejected (Neuvonen, 1970; Spall, 1973) on the following grounds. The poles from the oldest (2700—1900 Ma) to the youngest (1300—1200 Ma) rocks (Fig. 25-2) form a chronologic sequence which continues through and beyond the Palaeozoic poles. It also seems unlikely that all the diverse rock types, representing widely separated areas and different metamorphic grades, would be completely remagnetized. Further, although there is a wide range of K-Ar ages, none of them is Palaeozoic. Finally, the baked contact test for both normal and reversed sub-Jotnian dykes indicates remanence acquisition during intrusion and not during a regional remagnetization (Pesonen and Suominen, in prep.).

The internal consistency of the palaeomagnetic data of various ages in the Baltic Shield (e.g. the Svecokarelian, sub-Jotnian and Jotnian rocks), the positive results of baked contact tests (e.g. Patchett et al., 1978; Neuvonen et al., 1981; Pesonen et al., in prep.) and the discovery of new field reversals (Piper, 1979b, 1980; Pesonen et al., in prep.) make it unlikely that the smooth chronological sequence of poles in Fig. 25-2 has been caused by remagnetization. Future studies should, however, be carried out to clarify whether the scatter of poles within the swathes and the differences between the swathes of Figs. 25-2 and 25-3 can be explained by unremoved secondary components.

Asymmetric reversals

It is noteworthy that almost all of the palaeomagnetic reversals observed in the Baltic Shield show departures from the 180° symmetry (e.g. reversals 27—28, 30—31, 63—64 and 65—66). In some cases this may be caused by unremoved secondary components, but the baked contact tests, geochemical data and radiometric ages suggest that the asymmetry is not always due to secondary overprinting (Pesonen, in prep.). The asymmetry can also be caused by non-dipolar geomagnetic field disturbances, short period geomagnetic excursions or apparent polar wander (e.g. plate motion) during the polarity change (e.g. Pesonen, 1979a). The last possibility is interesting as it would mean that relatively detailed APWP portions can be established. The portions connecting two poles of opposite polarity can be fixed if the relative ages of the two poles can be determined by geological means (see detailed discussion of this topic in Pesonen, 1978).

TECTONIC IMPLICATIONS OF THE APWP

From the discussions presented above we may conclude that the APWP is produced by tectonic causes such as motion of the Baltic Shield as a whole, differential uplift within the shield (Ueno and Irving, 1976; Neuvonen, 1978) or relative motion of individual parts of the shield. The last two mechanisms are of special interest regarding the evolution of the shield itself, depending on whether ensialic within-shield tectonics or Wilson-cycle plate tectonics dominated (McElhinny and McWilliams, 1977; *see also McWilliams, this volume, Chapter 26, ed.*).

Differential uplift and APWP

Morgan (1976) and Beckmann et al. (1977) have shown that smooth APW tracks can be produced if the uplift after regional metamorphism takes place differentially, so that one block rises more rapidly than nearby blocks (see also Ueno and Irving, 1976). In principle, the entire Precambrian APW path could be caused by differential uplift of subunits of the shield. For example, a suitable combination of regional tilt and uplift gives rise to a smooth polar wander path (Morgan, 1976).

Neuvonen (1978) tested this model in the Baltic Shield from the APW data with ages between 1800 Ma and 1600 Ma. He showed that the model cannot explain the APW track. Neuvonen based his arguments on the internal consistency of the Svecokarelian poles (25 and 26) and of the sub-Jotnian poles (29 and 32), which were derived from widely separated rock units (Fig. 25-1). This result does not preclude the possibility that small, especially post-metamorphic parts of the APW paths were produced by differential uplift.

Tectonics within the Baltic Shield

The internal consistency of the Svecokarelian (poles 21—22 and 16—20), sub-Jotnian (34—37 and 31—33) and Jotnian poles (poles 48—56 and 45—47) in Sweden and Finland suggests that major tectonic rotations or relative displacements have not taken place in this province since the Svecokarelian orogeny.

The Sveconorwegian poles of Sweden and Norway (e.g. poles 67 and 72) also show good consistency, but the areal distribution of sampling sites is restricted (Fig. 25-1). The consistency of palaeomagnetic poles from the Archaean and the Granulite Complex cannot be determined from the scarce data available.

Comparison of palaeomagnetic poles of rocks with similar age but from different tectonic blocks of the Baltic Shield can only be made in a few cases because of lack of data.

The 2400—1900 Ma old poles derived from different inliers of the early Precambrian province (e.g. poles 3—4 vs. poles 5—12; Fig. 25-1) all plot in the same region (Fig. 25-2). Similarly, the poles of the Akujärvi quartz-diorite from the Granulite Complex (pole 14) and the synorogenic intrusives from the Svecokarelian province (poles 16—22) lie close to one another although these provinces are separated by large Archaean rock sequences. These examples suggest that no major movements have taken place between the tectonic blocks of the Baltic Shield. Therefore, it seems that within-shield tectonics has dominated during the evolution of the Baltic Shield since pre-Svecokarelian times. The tectonic style during the Archaean (before 2400 Ma) cannot yet be determined on palaeomagnetic grounds as there is only one early Precambrian pole available.

Ukrainian Shield vs. Baltic Shield

The palaeomagnetic poles from the Ukrainian Shield (Kruglyakova, 1961) are plotted in Fig. 25-2 together with the poles from the Baltic Shield. It is evident that the Ukrainian poles from two groups display a reasonable fit with the APW curve of the Baltic Shield (see also Spall, 1973).

The first group formed by poles 74—78 fits into the early Precambrian—Svecokarelian APW track (Fig. 25-2) although the radiometric ages with a wide range (about 2400—1700 Ma; Semenenko et al., 1968) suggest that they may be younger. The other group of poles (80—82) is characterized by an age of 1750 Ma (Table 25-I), which is in accordance with the Baltic Shield palaeomagnetic data. Further, the poles and ages (79) of the Turchingi gabbro (1400 Ma) and Ragunda dykes (42—43; $\leqslant$ 1320 Ma) are rather similar. Although the demagnetization studies of Ukrainian rocks are limited, we can tentatively conclude that the Ukrainian results fit well with the APW curve of the Baltic Shield. This suggests that no major rotations have taken place between these two European Shields since Archaean time.

Baltic Shield vs. Laurentian Shield

The APW path of the Baltic Shield is compared to that of the Laurentian Shield in Fig. 25-4. The APW curve of the Laurentian Shield is compiled from the APW paths determined by Irving and McGlynn (1976) and McElhinny and McWilliams (1977), except for the 1350—1000 Ma interval which is taken from Pesonen (1978). The comparison to follow is restricted only to the data from North American cratons. Piper (1980) has made a similar comparison including the data from Greenland and Scotland. Several major differences between the two APW paths are recognized (Fig. 25-4). The

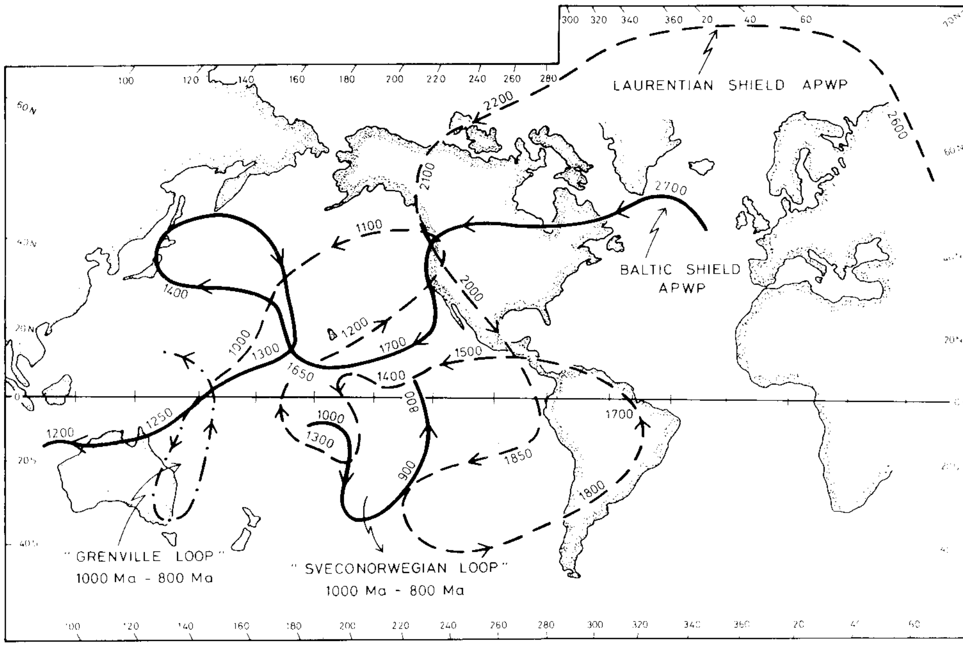

Fig. 25-4. Comparison of the APW paths of the Baltic and Laurentian Shields. The APW path of Fig. 25-2 is used for the Baltic Shield. The Laurentian APW path is compiled from data by Irving and McGlynn (1976; age interval 2200—800 Ma), McElhinny and McWilliams (1977; age interval 2200—2700 Ma) and Pesonen (1979a; age interval 1300—1000 Ma). Both curves are shown in present geographic coordinates. Closing the Atlantic Ocean into its pre-Mesozoic state (Bullard et al., 1965) would move the Baltic Shield poles approximately 38° to the west.

overall shapes of the two paths are clearly different. For example, the curve for the Baltic Shield shows a prominent sub-Jotnian polar wander loop during the age interval 1650—1320 Ma. In the Laurentian APW path a corresponding loop is missing, although recent palaeomagnetic results from Greenland (Piper, 1980) give indications of a somewhat similar loop. However, there is a major anticlockwise loop (Logan loop) in the Laurentian path during the age interval 1200—1000 Ma (Pesonen, 1979a). Such a loop has not yet been detected in the Baltic APW path. This may, however, simply result from the lack of palaeomagnetic data for this age interval. If the APW path of Fig. 25-3 is adopted for the Baltic Shield, there exists an additional anticlockwise loop at about 1850—1750 Ma ago. A somewhat similar loop is recognized in the Laurentian APW path (1850—1600 Ma) but the shapes of the two loops are different (Fig. 25-4). Moreover, as shown by McElhinny and McWilliams (1977), the Laurentian loop of this age interval may extend much more to the east than shown in Fig. 25-4.

Some similarities in the two APW paths are also observed. For example, the early Precambrian—Svecokarelian/Hudsonian tracks (2700—1900 Ma) have similar shapes, although the Laurentian track of this age interval is much longer. The only tracks which are strikingly similar in both curves are the Sveconorwegian and Grenville loops of the age interval 1000—800 Ma. In both cases the poles form a hairpin-shaped, anticlockwise loop in the Southern Pacific region. Moreover, the closing of the Atlantic Ocean (Bullard et al., 1965) would cause the two loops to coincide (see also Donaldson et al., 1973; Buchan, 1978). Thus the palaeomagnetic data of the Baltic Shield support the idea of contiguity of the Grenville and Sveconorwegian provinces during the Sveconorwegian/Grenville orogenies (Poorter, 1976b; Patchett and Bylund, 1977). Similarities in geology (Wynne-Edwards and Hasan, 1970) and geochronological data (Patchet and Bylund, 1977) also support this conclusion (*see also Baer, this volume, Chapter 14, ed.*).

Recent palaeomagnetic results from the Laurentian Shield (Morris and Roy, 1977) appear to have resolved the "Grenville problem" and showed that the Grenville (about 1000—800 Ma old) and the Keweenawan (1200—1000 Ma old) tracks can be linked together without plate-tectonic suture mechanisms.

The corresponding problem in the Baltic Shield is the tectonic relation between the Sveconorwegian province and the rest of the shield (Fig. 25-2). This problem awaits solution as there are as yet no palaeomagnetic data for the age interval 1200—1000 Ma.

COMPARISON OF PALAEOLATITUDES

The palaeolatitude curves of the Baltic and Laurentian Shields are compared in Fig. 25-5. Both palaeolatitude curves have been derived directly from the APW paths of Fig. 25-4 using the method of Irving (1964, p. 186). The Baltic Shield curve was calculated with respect to the city of Kajaani which is located in the central part of the Shield (Fig. 25-1). The reference city for the Laurentian Shield is Winnipeg (see also Donaldson et al., 1973). In places where gaps exist in the APW curves, interpolations were made assuming constant APW speed. Although this method to produce the palaeolatitude curves has its limitations, the validity of the method was verified by comparing the palaeolatitude curve with actual palaeolatitude data (Baltic Shield) calculated directly from the category A palaeomagnetic results. This comparison (Fig. 25-5) shows that the data agree well with the smooth palaeolatitudinal curve.

The overall shapes of the two palaeolatitude curves are not similar. As the APW curves were also different, it seems probable that the shields have drifted independently during most of the Precambrian (Spall, 1973; Neuvonen, 1974). This conclusion is also supported by the different APW speeds of the shields. The velocity derived from the Baltic Shield curve (Fig. 25-2 and 25.3) is approximately 0.35°/Ma (range 0.31—0.40°/Ma), depending on

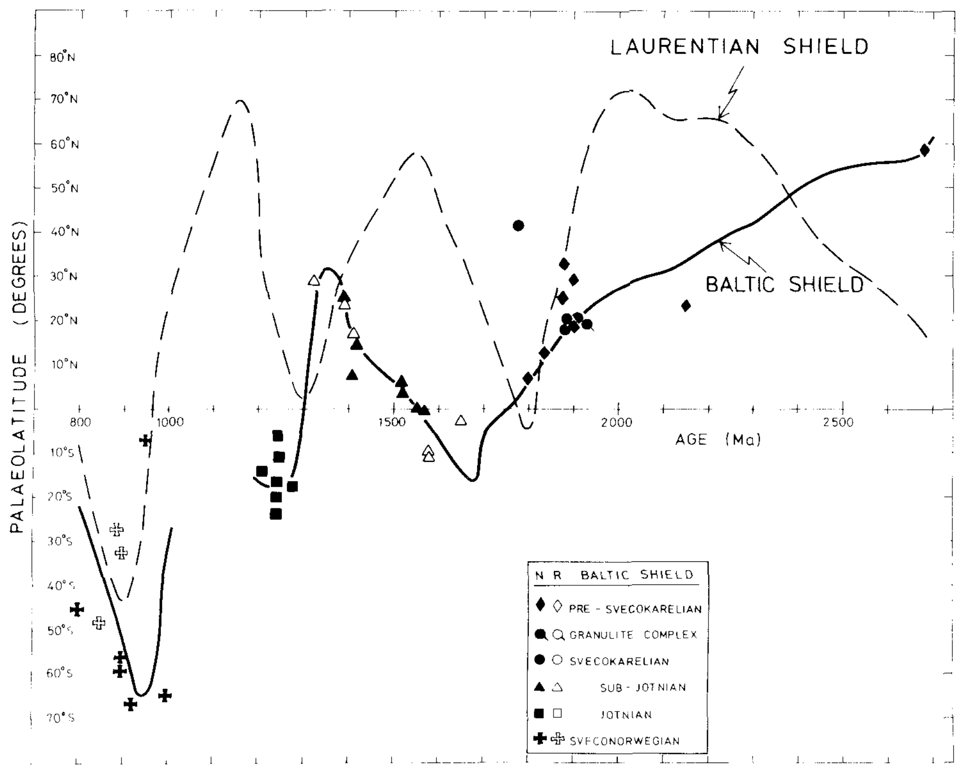

Fig. 25-5. Palarolatitude curves for the Baltic and Laurentian Shields. See p. 641 for explanation. The reference cities for the palaeolatitudes are Kajaani (Lat. = 64°13′, Long. 27°44′E; Baltic Shield) and Winnipeg (Lat. = 49°53′, Long. 97°10′W; Laurentian Shield).

whether the curve of Fig. 25-2 or 25-3 is used. The velocity of about 0.5°/Ma (Fig. 25-5) obtained for the Laurentian Shield is significantly higher.

According to Fig. 25-5 the two shields may have been in juxtaposition during three periods. The first period, indicated by crossing palaeolatitudes, is located in the age interval 1900—1750 Ma. Therefore, the cause of the Svecokarelian and coeval Hudsonian orogenies may have been a collision of the two shields about 1900 Ma ago. After this period the palaeolatitude curves depart and meet again about 1350—1250 Ma ago (Fig. 25-5). This second crossing of the latitudes may have some relation to the worldwide rifting and magmatic activity during the age interval 1300—1200 Ma (Jotnian/MacKenzie episodes; Neuvonen, 1974; Sawkins, 1976). The intimate connection of the Laurentian and Baltic Shields during this time is supported by the similarities in geology and magnetic properties ("normal" polarities) of the Jotnian/MacKenzie dykes (Patchett et al., 1978).

The palaeoclimatological conditions revealed by the metasediments on both shields agree well with the palaeolatitudes calculated and consequently support the idea of mutual collisions at about 1900 Ma and 1300 Ma ago (Neuvonen, 1974). Similar geological features in addition to the palaeomagnetic data support the contiguity of the Grenville and Sveconorwegian provinces during the Grenville/Sveconorwegian deformation.

CONCLUSIONS

The following conclusions can be drawn from the presently available Precambrian palaeomagnetic data of the Baltic Shield.

(1) The systematic distribution of palaeopoles along a smooth APW swathe in chronologic sequence between 2700 Ma and 800 Ma ago is due to a motion of the Baltic Shield relative to the pole. Sampling and measurement errors, unremoved secondary components and differences between radiometric and magnetization ages explain the scatter of poles within the APW path.

(2) The palaeomagnetic data for rocks with the same age but from different tectonic provinces are consistent. This suggests that the Baltic Shield has behaved as a coherent unit at least since late Archaean times. Further, the consistency indicates that ensialic tectonics has been dominant during the evolution of this shield. The Sveconorwegian poles differ significantly from those of the other provinces of the Baltic Shield. Whether this reflects suturing between the Sveconorwegian and other blocks of the shield can be tested only after palaeomagnetic data are obtained for the age interval 1200—1000 Ma.

(3) The APW curves of the Baltic and Laurentian Shields are not similar, suggesting that both shields have drifted independently during the Precambrian. However, the palaeolatitude data indicate that the shields may have been in juxtaposition about 1900 Ma, 1350—1250 Ma and 1000—800 Ma ago. Collision of the two shields 1900 Ma ago could have initiated the major orogenies of Svecokarelian and Hudsonian times (1900—1800 Ma ago). The crossing palaeolatitude curves of the two shields 1300 Ma ago may bear some relationships to the worldwide rifting and igneous activity (Jotnian/MacKenzie interval) about 1250 Ma ago. The contiguity during the Grenville/Sveconorwegian deformation (1000—800 Ma ago) is supported by the similarity of the two areas.

(4) The APW velocity of the Baltic Shield (0.35°/Ma) is significantly lower than that (0.50°/Ma) of the Laurentian Shield. In order to verify this difference, additional data are needed for the Baltic Shield, where "normal" polarity data appear to dominate.

(5) The palaeomagnetic data from the Ukrainian Shield are consistent with the results from the Baltic Shield. This suggests that no major relative movements have taken place tween these two European shields since the Archaean (*see also Poorter, this volume, Chapter 24, ed.*).

(6) Future studies of secondary components, asymmetric reversals and scatter of palaeopoles should be carried out on the basis of detailed demagnetization experiments. Further palaeomagnetic work should be focused on rocks in the age interval 1200—1000 Ma in order to solve the problem whether the Sveconorwegian province has always been part of the Baltic Shield (ensialic tectonics) or whether it has collided and sutured with the rest of the shield (plate tectonics).

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Chapter 26

PALAEOMAGNETISM AND PRECAMBRIAN TECTONIC EVOLUTION OF GONDWANA

M. O. McWILLIAMS

ABSTRACT

Precambrian palaeomagnetic data from Gondwana are reviewed with the goal of assessing the duration and magnitude of major intercratonic movements which may have occurred within the supercontinent during its evolution. The data suggest that Gondwana existed only from latest Precambrian or early Palaeozoic times up until its breakup in the Mesozoic. Prior to latest Precambrian times at least two major fragments are identifiable, East Gondwana (Australia, India, Antarctica) and West Gondwana (Africa, South America); these probably collided along the Pan-African Mozambique belt.

The data further suggest that (within the limits of uncertainty in the palaeomagnetic results) the older cratons within west Gondwana retained their early Palaeozoic configuration back as far as ~ 1000 Ma and perhaps longer. Thus not all Pan-African and older mobile belts mark the sites of major ocean closure, but rather formed without the destruction of vast amounts of oceanic lithosphere.

Data from each Gondwana suggest that after compensating for a small late Precambrian—early Palaeozoic rifting episode, the Australian cratonic assemblage may have maintained a stable configuration back to about 1500 Ma. Although no differential motion between the Yilgarn and Pilbara cratons is indicated by the palaeomagnetic data, the available data are insensitive to local rotations and thus the Ophthalmian—King Leopold—Halls Creek mobile zone system *could* have resulted from closure of a 1000—2000 km wide ocean.

The accrued geological evidence for distinct oceanic and continental crust in Precambrian times, coupled with the palaeomagnetic evidence for irregular movement of the continental lithosphere with respect to the spin axis, is consistent with a form of "plate tectonics" operating in the Precambrian oceanic domains. However, as a result of higher heat flux, the Precambrian continental crust may not have responded to external torques in the same way as the present continental crust. A direct uniformitarian application of modern plate tectonic theory to Precambrian continental crust is not wholly supported by the palaeomagnetic data.

INTRODUCTION

A fundamental goal of this volume is to examine the evidence for and against the existence of Mesozoic—Cenozoic-style plate tectonics in Precambrian times. The belief that plate tectonics has been operative in the recent geological past is now widely held, largely due to detailed studies in the disciplines of marine geology and geophysics. Due to subduction at convergent plate boundaries, the oldest remaining parts of the oceanic lithosphere in most oceans are latest Palaeozoic or Mesozoic in age. As a consequence the rich "direct" record of older plate kinematics locked within the plate is lost. The only possible way to recover the history of relative movements in earlier times is to study the much more fragmentary record left on the predominantly unsubductable continental lithosphere.

Three fields of investigation important to the study of this remaining continental record are: (1) structural analysis of features possibly produced by plate interaction such as orogenic (mobile) belts and proto-continental margins, (2) geochronology and (3) palaeomagnetism. Taken alone, none of these sources can provide unequivocal evidence for the existence or absence of plate tectonics in Precambrian times. Structural analysis of deformed tectonic zones and of sedimentary basins is of paramount qualitative importance, for with the correct set of rules, it can be used to infer what sort of plate interaction might have taken place to produce the large-scale structural features observed. Geochronology can yield critical information on the timing of possible plate interaction. In favourable circumstances, palaeomagnetic studies can yield important quantitative information as to the kinematics of plate interaction (e.g., direction and magnitude of relative displacement, minimum estimates of absolute plate velocities, etc.). The prime objective is to determine from these three fields of study whether or not a uniformitarian application of the same set of rules which have been invoked to explain Mesozoic—Cenozoic structures is valid for Precambrian times. The scope of this review is limited to the pre-Devonian evolution of the southern continents (South America, Africa, India, Australia and Antarctica). As initial conditions it is assumed that: (1) at about 400 Ma ago these continents were configured in a supercontinental assemblage called Gondwana: and (2) that plate tectonics is the causal mechanism for the post-Devonian fission of the supercontinent.

THE WILSON CYCLE

In his now classic paper, J. T. Wilson (1965) laid the foundations for the geometrical rules of modern plate tectonics by showing how the earth's lithospheric plates interact at their various divergent, transform and convergent boundaries. As developments in plate tectonics accelerated (McKenzie and Parker, 1967; Morgan, 1968; Le Pichon, 1968), Wilson showed the relevance of considering global geological history in terms of cyclical opening and closing of major oceans (Wilson, 1966; 1968). In recognition of his contributions, Dewey and Burke (1974) introduced the term "Wilson cycle" to refer to this ocean opening-closing cycle.

Well before the Wilson cycle was actually given a formal name, earth scientists were quick to interpret Phanerozoic mountain belts as sites of plate convergence preceded by the opening of an ocean (see e.g. Dewey, 1969; Hamilton, 1970; Bird and Dewey, 1970; Dewey and Bird, 1970). By comparison with previous orogenic hypotheses, the remarkable success of these Phanerozoic interpretations led to a uniformitarian application of the Wilson cycle to Precambrian orogenic domains (Burke and Dewey, 1972, 1973; Dewey and Burke, 1973; Talbot, 1973; Walker, 1973).

Any discussion of whether the Wilson cycle was operative in Precambrian times must be prefaced by a consideration of what evidence for its existence might be expected to be preserved on the continents today. As all older oceanic material would have been subducted (a small fraction would have been obducted), all potential evidence must come from the older parts of the continental crust. Obviously, evidence produced by three types of plate interaction must be considered: divergent, convergent and transform motion.

Divergent plate motion produces Atlantic-type rifted continental margins, with characteristic initial graben structures, shelf-slope deposits, possibly aulacogens and certainly a distinctive depositional evolution (see e.g. Dewey and Bird, 1970; Dickinson, 1970). If the Wilson cycle operated in Precambrian times, many such Atlantic-type margins would have been created; however, many of the distinctive features of these continental margins would be extensively modified (and perhaps totally obliterated) by continent—continent collision in the later convergent phase of the cycle. A number of Precambrian sedimentary sequences of various states of preservation have been interpreted as ancient analogs of Atlantic-type continental margins (see e.g. Hoffman, 1973; Hoffman et al., 1974; Dunnet, 1976; Erikson, 1979).

Evidence for transform motion would take the form of offset of pre-existing structures such as metamorphic-plutonic belts or narrow depositional troughs. Considering that: (1) most continental transform motion taking place today occurs near continental margins where it might later be obliterated; and (2) that transforms do not produce striking and uniquely recognizable features which would be preserved for long periods, direct geological evidence for Precambrian transform motion might not be expected to be widely apparent.

Structures produced during continent—continent collision (orogenic belts and suture zones) are probably the most frequently preserved indicator of Wilson cycle operation for two reasons. Firstly, the subduction/collision processes produce the most dramatic effects in pre-existing continental crust at and near the continental margin (deformation, metamorphism, intrusion, thrusting and deformation, crustal thickening etc.). Secondly, continental collision is the terminal phase of any individual Wilson cycle. Although subsequent Wilson cycles are probably often initiated at or near the sites of previous collision, in most cases the new opening will not occur about precisely the same rotation pole of closure (Wilson, 1966; Dewey and Bird, 1970). Further, in those cases where collision has not been followed by a new Wilson cycle, the collisional suture so produced will remain intact, subject only to probable deep erosion and possible cover by younger sediments (Burke and Dewey, 1973; Dewey, 1977).

During the closure of a large ocean, it might be expected that a number of arc—continent collisions would occur prior to terminal continent—continent suturing. Clearly, much of the evidence for these arc—continent events might be destroyed at the culmination of each Wilson cycle, but depending upon the completeness of suturing, considerable evidence for arc—continent collision may remain intact (Dewey, 1977). Thus, if the Wilson cycle did operate in Precambrian times, it might be expected that evidence for its existence would be most frequently preserved in continental collision (suture) orogens, and arc—continent collision orogens, and less frequently by rifted (Atlantic-type) continental margins and transform-produced offsets.

The identification of old collisional sutures is a complicated and often equivocal task. The great complexity of these sutures makes it difficult to recognize the original lines of closure and dismemberment, particularly where quite irregular continental margins have collided (Dewey and Burke, 1973, 1974; Dewey, 1976, 1977; Sengör, 1978). A compounding factor is that much of the geological evidence which is characteristic of suture zones is developed at relatively shallow structural levels, while older possible suture zones have been eroded to a deep level. Further complications arise from the need to identify a multitude of possible "non-terminal" sutures (Dewey, 1977), i.e. the attachment of various sedimentary wedge, island arc and ophiolitic fragments to the leading continental edge during the subduction process but before the terminal continent—continent collision. In subducting a large oceanic plate, a considerable amount of this oceanic debris may accumulate in the process of producing a major continental suture. As noted by Dewey and Bird (1971), much of the material preserved in collisional orogens is associated with the closing of smaller marginal basins rather than the larger oceans which close behind them.

Considering the potential complexities and state of preservation of Precambrian suture zones, it is obvious that a direct assessment of Precambrian Wilson cycle operation is difficult solely from geological evidence. A uniformitarian approach would dictate that an exhaustive investigation of each possible collisional suture is unnecessary. Since orogenesis occurs today concomitant with plate convergence and has done so in the very recent past, older orogens too are the result of convergence and collision (see e.g. Dewey and Burke, 1973; Talbot, 1973; Walker, 1973; Burke et al., 1977). Detailed study of many orogenic belts (dominantly Precambrian in age and especially in Africa and Australia) has led to contradictory inference, i.e. that some major orogens do not result directly from collision subsequent to large-scale horizontal displacement (see e.g. Shackleton, 1973, 1976; Rutland, 1973; Duff and Langworth, 1974; Watson, 1976; Kröner, 1977a,b, 1979b). Uniformitarians would of course argue that no signs of large-scale horizontal motion would be expected to be observed, precisely because of the operation of the Wilson cycle: (1) all oceans are small twice and most of the complex evidence for Wilson cycle operation is created at these times; (2) much or all of this evidence is extensively modified in terminal collision; and (3) little of the diagnostic suture zone structure remains after deep erosion.

An important problem in determining the possible ways in which the Wilson cycle may have operated is an assessment of the size of the oceans that may have opened and closed. The convergent phase of the Wilson cycle can begin at any time when the oceanic lithosphere has cooled and become sufficiently dense to be subducted. This could conceivably occur when an ocean was as narrow as 1000 km or wider than 10 000 km, depending upon spreading rates. In either case, the terminal sutures created, especially those remaining after erosion has exposed the deep roots of the collisional orogen, will be essentially identical (Dewey, 1977). Thus, the undisputed existence of a Precambrian suture is not a sufficient condition to confirm the existence of the Wilson cycle, as such evidence only indicates that an ocean of unknown size closed. Confirmation of horizontal displacements of the scale of the Mesozoic—Cenozoic Wilson cycle requires quantitative displacement information which can only come from palaeomagnetic data.

PALAEOMAGNETIC EXPECTATIONS AND SOURCES OF ERROR

While detailed geologic analyses can suggest the presence of an older collisional suture, palaeomagnetism is presently the only tool which can provide quantitative information about the scale of relative cratonic displacement prior to collision. Consequently, the interpretation of palaeomagnetic data has formed an important part in the debate about the existence or absence of the Wilson cycle in Precambrian times. But palaeomagnetism can provide accurate quantitative information only in favourable circumstances. This consideration is especially important (and sometimes ignored) in Precambrian studies where the required circumstances occur less frequently than in Phanerozoic times. The difficulties in applying palaeomagnetic data to Precambrian orogens as compared to their Mesozoic—Cenozoic analogs can be traced to three sources: (1) uncertainties in magnetization age; (2) uncertainty in structural control (tectonic attitude of sampled units); and (3) relatively poor temporal data density. The first two sources lead to the lower quality of an individual palaeomagnetic result compared to Phanerozoic data; these sources, on conjunction with poor data density, lead to complications about the construction of apparent polar wander paths (APWPs) for orogenically or suture-bounded cratons. A fourth uncertainty enters in the form of magnetic polarity assignments; in comparing discrete APWP segments which do not connect directly with the relatively better known Phanerozoic APWP, it is often possible to construct two APWP segments (normal or reversed) from the same data set. Tectonic interpretations made from the two paths may be quite different.

The Precambrian geomagnetic field is assumed to be described by an axial geocentric dipole when averaged over sufficiently long intervals in all Precambrian palaeomagnetic studies. This admittedly uniformitarian assumption is supported by observations of Precambrian geomagnetic field behaviour which are remarkably similar to field behaviour in Mesozoic and Cenozoic times, including antipolar reversals, excursions and intensity variations.

Magnetization age

Modern palaeomagnetic investigations employ a wide variety of demagnetization techniques to isolate the various components of magnetization which may be present in a rock sample. For details of techniques and magnetization types, consult McElhinny (1973) and Collinson (1975). The magnetization sought in most studies is the primary magnetization acquired by the rock at about the time it was formed. In igneous rocks this component is usually a thermoremanent magnetization (TRM) but can be a high-temperature, chemical remanent magnetization (CRM). In sedimentary rocks the primary magnetization is either a CRM or detrital remanent magnetization (DRM). For older rocks, especially those which may have been heated within or near an orogenic belt after acquiring a primary magnetization, there is a reasonable probability that a younger secondary magnetization may be superimposed upon the primary component. Depending upon prevalent time and temperature conditions, a secondary magnetization may partially or totally replace a pre-existing magnetization.

Apart from thermally acquired secondary magnetizations due to orogenesis, burial or intrusion, secondary magnetizations may be acquired at lower temperatures by thermal relaxation (VRM: viscous remanent magnetization) or chemically due to effects of weathering or subsurface alteration by groundwater. Hydrothermal effects of fluids circulating near intrusive and plutonic rocks can also be an important source of secondary magnetizations at lower temperatures.

The necessity to determine uniquely each of the magnetic components in rock samples led to techniques of multicomponent analysis. Interest in multicomponent magnetizations has led to considerable study of demagnetization data reduction techniques and data interpretation (see e.g. Zijderveld, 1967; Halls, 1976; Hoffman and Day, 1978; Stupavsky and Symons, 1978). It is now fairly common in palaeomagnetic studies of older rocks to obtain two or more components of magnetization from an individual unit; often one magnetization which is interpreted as primary and one or more secondary magnetizations which can in some cases be linked with a later thermal or thermochemical event.

While the direction of the various components present in a sample or collection of samples can often be determined (frequently with an estimate of their relative ages), a fundamental obstacle in Precambrian studies is estimating an absolute magnetization age. In studies of Phanerozoic sedimentary rocks it is often possible to associate a primary magnetization with a particular palaeontological reference, and thus ages can be estimated that are accurate to within a subdivision of a geologic period. In Precambrian sedimentary rocks, body fossils are absent, trace fossils are diagnostically difficult and age estimation by stromatolite morphology remains a developing discipline. Thus, absolute magnetization ages for sedimentary rocks are often constrained only by radiometric ages from basement rocks and cross-cutting or overlying igneous rocks. In studies of thick Precambrian sedimentary sections devoid of radiometric ages, one approach has been to construct a relative magnetization chonology from stratigraphic relationships, and to link the endpoints of this APWP with radiometrically dated results from elsewhere on the craton (Elston and Grommé, 1975; McWilliams and McElhinny, 1980; Kröner et al., 1980).

Assignment of absolute magnetization ages in igneous and metamorphic rocks is commonly done by equating a particular radiometric age (K-Ar, Rb-Sr or U-Pb) with the magnetization age. While this is often the only estimate of magnetization age available, the practice can be a source of considerable error. Radiometric ages reflect the time at which a particular mineral-isotope or rock-isotope system became closed or "blocked" to isotopic diffusion. The problem as it relates to magnetization age is that isotopic and magnetic systems have a potentially wide thermal range for closure. Blocking of a particular magnetic system (magnetite-ulvospinel or hematite-ilmenite solid solution series) may occur before or after blocking of the isotopic system. A further complication in Precambrian studies is that certain isotopic systems, notably Rb-Sr whole-rock systems when considered over a large enough sample volume, are insensitive to later thermal events which can quite drastically affect the magnetic system (Grant, 1964). For example, an anorthosite body dated at 1400 Ma which was reheated at 900 Ma could yield an original 1400 Ma whole-rock Rb-Sr age and a 900 Ma K-Ar mineral age. While the reheating may have only reached 300—400°C, it is possible (and indeed probable), that the primary magnetization would be overprinted with a 900 Ma secondary magnetization. Given sufficient time at this temperature (several Ma), the primary magnetization may be totally replaced by the secondary magnetization. In this case, equation of primary crystallization age (Rb-Sr whole rock) with the isolated magnetization would be incorrect and could lead to serious problems in APWP construction and interpretation.

Recognition of this problem in recent years has led to detailed consideration of the interrelationship of the magnetic and isotopic systems (see e.g. Buchan et al. 1977; McWilliams and Dunlop, 1978; York, 1979a,b). Although no exact rules for the assessment of magnetization age from isotopic age can yet be constructed, in many cases mineral K-Ar ages (especially hornblende and biotite) may be the most accurate indicators of true magnetization age. The most definitive magnetization ages for secondary as well as primary magnetizations will probably come in the future from the incremental heating techniques, as the $^{40}Ar/^{39}Ar$ release spectra and associated ages can be compared directly with blocking temperature spectra of multi-component magnetization systems.

Structural control

When attempting to construct APWPs for individual cratons, isolation and age determination of the magnetic components present in a particular rock unit is obviously not a meaningful exercise unless compensation for post-magnetization tectonic disturbances such as folding, tilt or rotation about a local vertical axis can be achieved. Unless tectonic rotation information is sought (e.g. Beck, 1977; Simpson and Cox, 1977; Irving, 1979), palaeomagnetic studies must be confined to areas where local tectonics can be readily unravelled. Palaeomagnetic studies of Precambrian sedimentary and layered volcanic rocks are usually done in areas where the sampled strata exhibit simple structures; folds in the beds can be used to test the age of magnetization relative to folding. Rotations about a local vertical axis can be tested for by sampling over a large lateral extent; concordant results suggest either that no rotation has occurred or at least indicate the minimum size of the rotated block.

Because plutonic and intrusive bodies often contain stable magnetic carriers and can be isotopically dated, many Precambrian palaeomagnetic results are derived from such rocks. In most cases, tectonic control for these rocks is poor or absent; arguments for the absence of post-magnetization tectonic disruption are usually derived from the observation of concordant directions of magnetization over a wide lateral sampling extent. If significant disturbances have occurred, it might be expected that results from widely spaced localities would disagree. Still, the uncertainty of restoring plutonic and intrusive bodies to their correct palaeo-orientation is a significant potential source of error in constructing cratonic APWPs.

Data density

Examination of published compilations of palaeomagnetic data shows that at present the data density (poles per Ma) for the period 2800 to 570 Ma ago is depleted by a factor of 10—15 when compared to the 200 to 0 Ma interval. Briden (1976) has illustrated the parallel between data density of the Precambrian palaeomagnetic database today and the data density of the Phanerozoic palaeomagnetic database in the late 1950's when the first APWPs were constructed and the reality of continental drift became apparent to some. The primitive paths of those times have been extensively augmented and refined; yet the underlying conclusions of these early data remain intact. By analogy, it might be expected that a similar future awaits Proterozoic studies. While it is true that there are not sufficient data at present to conclusively prove or disprove the existence of any particular global tectonic model, the overall quality and density of the data have reached the stage where the gross features of apparent polar motion are visible and therefore important inferences can be made.

Since Precambrian APWPs are severely underconstrained, the tectonic implications of these paths must necessarily suffer the same disadvantages. As the following parts of this review show, in many cases the fundamental problem is one of interpretation. Comparison of APWPs is most often done in a non-rigorous fashion, and thus two sparsely populated paths may be interpreted as following identical trends because they show no gross signs of dissimilarity, yet strictly they are different because they do not coincide. The apparent lack of coincidence is viewed by some as evidence for relative displacement in a uniformitarian model, while opponents view the disparity as simply a sampling problem. In the latter interpretation, as more data accumulate to fill the gaps in the record, a clearly defined single APWP would then be evident.

The construction of a single cratonic APWP suffers from interpretive problems. When attempting to join poles in an ordered time sequence on an unknown path, it is nearly always possible to combine the data on a relatively smooth path which does not violate any of the applicable constraints, given: (1) relatively large uncertainties in pole position and magnetization age: (2) uncertainty in polarity assignments; and (3) the general sparse nature of the data. Frequently several such interpretations can be made which invoke different polarity assignments for parts of the dataset, leading to a different APWP connection. Usually, a simplifying condition such as minimizing total path length (Irving, 1979) or inferred absolute lithospheric velocity (Gordon et al. 1979) is invoked to constrain the path of apparent polar motion.

PRECAMBRIAN WILSON CYCLE SIGNATURES

Given the preceding caveats regarding the basic nature of Precambrian palaeomagnetic data, the important question to be answered before a detailed analysis begins is "will continental displacements such as those envisaged for Phanerozoic Wilson cycles be visible in the Precambrian palaeomagnetic record?". Apparent polar wander paths have often been depicted as narrow lines connecting successive poles (see e.g. McElhinny, 1973). A fuller appreciation of the data has led to the concept of APW swathes some 10—15 degrees in width (Beck, 1970; Piper et al., 1973). The swathes are regions within which motion of the palaeomagnetic pole is most likely to have occurred, given the available data. The wide swathes serve as a guide to emphasize the overall fundamental APW trend while attenuating higher frequency APW components which may not reflect real cratonic displacement. So long as the apparent polar motion of the cratonic fragments exceeds the limitations of the swathe, operation of the Wilson cycle can be detected

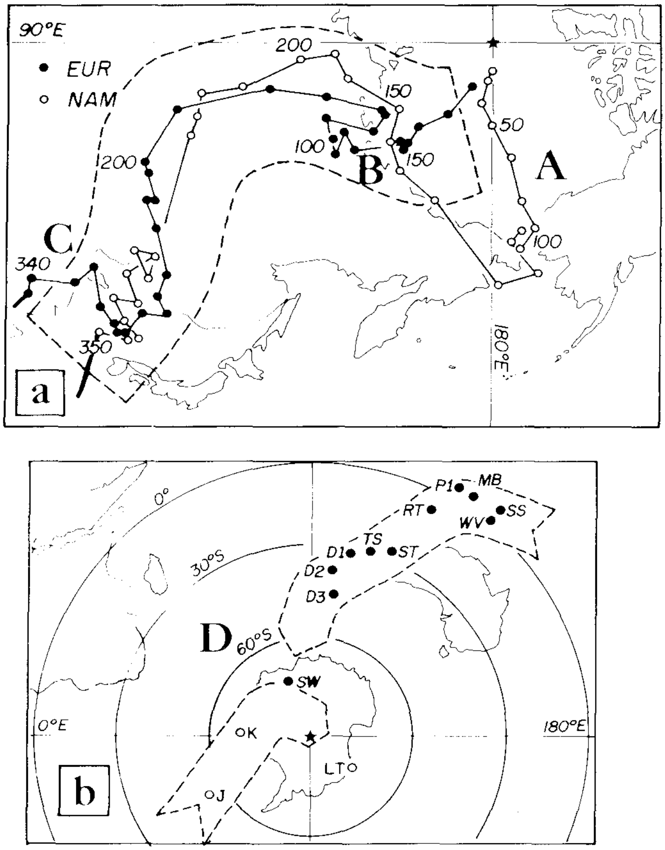

Fig. 26-1. a. Smoothed APW paths for North America (*NAM*) and Eurasia (*EUR*), from Irving (1979). Points plotted at 10 Ma intervals; Eurasian data corrected for opening of North Atlantic. Superimposed swathe width 15 degrees at equator.
b. APW paths for Peninsular India (solid circles) and China (open circles), illustrating separate, convergent paths. Reversed Eurasian path of Fig. 1a also collides with Indian path. Swathe width 15 degrees at equator.
Data for India for Klootwijk (1976): *SS* = speckled sandstone (Early P); *WV* = Wardha Valley redbeds (Late P—Early Tr); *P1* = Panchet beds (Late P—Early Tr); *MB* = Mangli beds (Late P—Early Tr); *RT* = Rajmahal traps (Early K); *ST* = Satavedu sandstone (Late K); *TS* = Tirupati sandstone (Late K); *D1* = Deccan traps, lower reversed (Early T, 65—60 Ma); *D2* = Deccan traps, upper normal (Early T, 65—60 Ma); *D3* = Deccan traps, upper normal (Early T, 65—60 Ma); *SW* = Siwalik beds (Miocene).
Data for China from McElhinny (1973): *J* = Jurassic; *K* = Cretaceous, *LT* = lower Tertiary.

if data are available which span the time interval of ocean opening and closing. However, this can only occur with a favourable configuration of sampling site, palaeomagnetic pole, and rotation (Euler) pole (see e.g.

McElhinny, 1973, p. 237; Gordon et al., 1979, fig. 2). In unfavourable circumstances, when palaeomagnetic and rotation poles are in close proximity, opening and closing of an ocean produces no significant APW effect. Assuming a complete spectrum of motions, if the Wilson cycle did operate in Precambrian times even with a relatively complete dataset, only a fraction of the openings and closings would be observable.

As examples of how the use of APW swathes affects the interpretation of the data in terms of tectonic displacements, Fig. 26-1 illustrates highly refined palaeomagnetic records of North America and Eurasia for the past 300 Ma from Irving (1979) and abbreviated palaeomagnetic records from India (Klootwijk, 1976) and Asia (McElhinny, 1973). Superimposed upon these paths is a hypothetical 15 degree APW swathe similar to that which might be constructed using data of Precambrian quality. Immediately apparent are these observations: (1) the most recent opening of the North Atlantic is just apparent (*A* in Fig. 26-1a); however, given the relatively sparse Precambrian record, it is quite conceivable that the opening of the North Atlantic could be missed entirely; (2) the complex nature of the Eurasian APWP during Cretaceous times (*B* in Fig. 26-1a) would probably not be visible; (3) the Siluro-Devonian closure of the Iapetus Ocean (*C* in Fig. 26-1a) might be quite difficult to demonstrate from the swathes which would represent the data. However, (4) the collision of India with Asia is quite diagnostic and probably would not be missed even were it represented by data of Precambrian quality (*D* in Fig. 26-1b). Observations (1) and (3) result largely from the proximity of the poles of opening/closing for the last Atlantic Wilson cycle, while (4) results from the fact that the absolute motion of India during Mesozoic and Cenozoic times had a dominant northward component. To summarize, the following points are important in interpreting Precambrian palaeomagnetic data:

(a) The absence of apparent polar motion does not preclude large real lithospheric motion.

(b) The presence of large apparent polar motion does not imply large real lithospheric motion.

(c) Apparent polar motion within the limits of the swathe is undetectable.

APPARENT POLAR WANDER PATHS

The logical starting point for the present analysis is an early Palaeozoic reconstruction of Gondwana. The fit proposed by Smith and Hallam (1970) has often been employed; other reconstructions are also possible (Barron et al. 1978; Norton and Sclater, 1979). We next pose the question "How were the various Gondwana fragments configured during and before the prebreakup Pan-African orogenesis between 750 and 450 Ma ago?" As will be shown, the palaeomagnetic data suggest that at least two independent fragments (called East and West Gondwana here) were probably involved in the

Pan-African events. After this discussion, the next step in the analysis is an interpretation of what fragments may have moved independently within East and West Gondwana to form pre-Pan-African mobile belts within these older separate units. This iterative process continues until all the applicable data are exhausted. Clearly, it makes sense only to discuss orogenic events for which there exist palaeomagnetic data older than the age of orogenic activity. As will become apparent, this is a considerable limitation with increasing age.

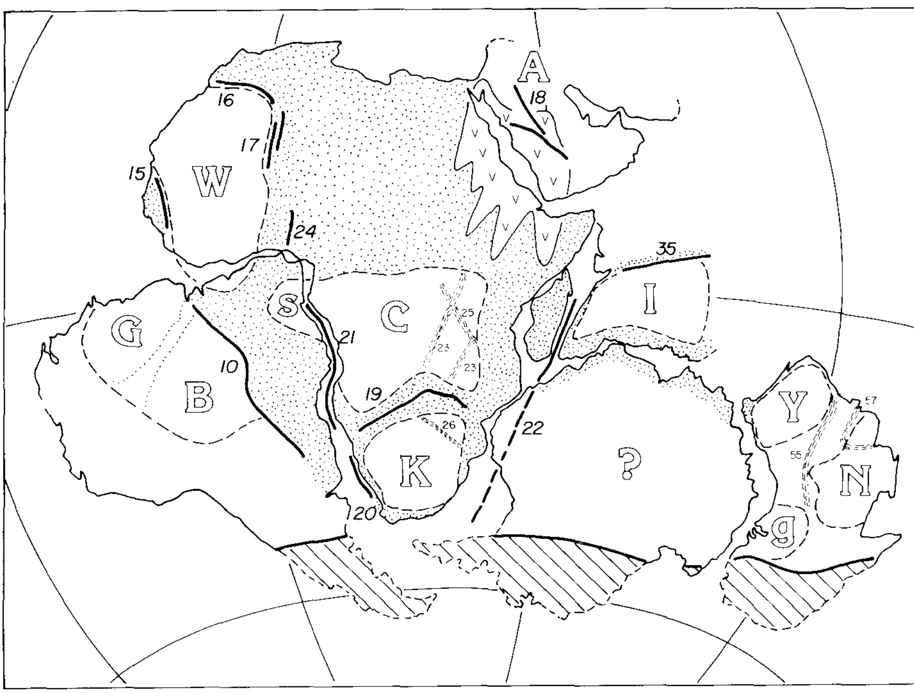

Fig. 26-2. Smith-Hallam (1970) schematic reconstruction of Gondwana, showing areas of Pan-African tectonic and tectono-thermal activity (stippled regions). Shield areas stable in late Proterozoic times: G = Guyana; B = Brazilian; s = São Francisco; W = West African; C = Congo; K = Kalahari; A = Arabian; I = Indian; N = Central Australia province (Kimberley, Pine Creek, McArthur, Mt. Isa, Arunta, Musgrave terranes, Amadeus basin); g = Southeast Australia province (Gawler, Mt. Painter, Willyama terranes, Adelaide geosyncline); Y = Western Australia province (Yilgarn, Pilbara terranes). Bold lines within reactivated zones: possible Pan-African sutures, numbers refer to Burke et al. (1977). *10* = boundary of Brazilian domain; *15* = Mauritanides; *16* = Bou Azzer; *17* = Pharusian; *18* = Jabal al Wask zone; *19* = Damara—Zambesi belt; *20* = Gariep belt; *22* = Mozambique belt; *24* = Dahomeyan belt; *35* = Aravallis. Other numbers refer to possible older sutures in Burke et al. (1977). Hijaz accretionary arc province (v pattern) from Kröner (1979b).

The Pan-African domains

The Pan-African belt system traverses nearly the entire African continent including Madagascar and is characterized by isotopic ages roughly within the range 750 to 450 Ma. Detailed geological reviews of the Pan-African system are presented elsewhere (Caby et al., this volume, Chapter 16; Gass, this volume, Chapter 15; Kröner, 1979b). The belts within Africa can be categorized (Clifford, 1970; Kröner, 1979b) into three broadly defined groups, all with characteristic Pan-African mineral isotopic ages: (1) orogenically deformed Upper Proterozoic geosynclinal sediments; (2) zones of deformed and rejuvenated older basement rocks; and (3) wide zones (largely in central and north Africa) where Pan-African isotopic ages predominate, but where the penetrative deformation found in (1) and (2) is absent. In the latter zones, the Pan-African events appear to be mainly thermal, while in the former zones tectonothermal effects predominate. In most Gondwana reconstructions, it can be seen that thermal and tectonothermal belts of Pan-African age form an interconnected network within the supercontinent (Fig. 26-2).

The worldwide map of possible sutures (Burke et al., 1977) is a convenient starting point for identification of potential Pan-African plate tectonics within Gondwana. Superimposed upon the map of reactivated Pan-African belts are possible sutures (suture numbers in Fig. 26-2 refer to Burke et al., 1977). By inspecting the geometrical configuration of cratons in the figure, it is apparent that these potential sutures cannot all be both real and contemporaneous. Given the fact that the activity in the Pan-African domains spans some 300 Ma, it is likely that the sutures which are indeed real were not formed at precisely the same time. As will be shown, a case can be made for the absence of large-scale relative motion in the formation of some of the potential sutures.

In Fig. 26-3 Ordovician palaeomagnetic results are plotted from Gondwana, representing the time at which Pan-African tectonism had ceased. The relatively good agreement of these poles suggests that, at this time, previous large-scale differential APW between the various Gondwana fragments had ceased, and Gondwana could be considered to be an intact unit until its breakup in the Mesozoic (see e.g. Schmidt and Morris, 1977). Most of the pre-Ordovician data relevant to Pan-African tectonics come from Africa and Australia. There are only a few late Precambrian to Cambrian poles from India, some latest Precambrian–Cambrian data of limited reliability from South America and no relevant late Precambrian data from Antarctica. The analysis is therefore centered upon a discussion of relative apparent polar motion between Africa and Australia; relative motions of the other fragments are correspondingly less certain.

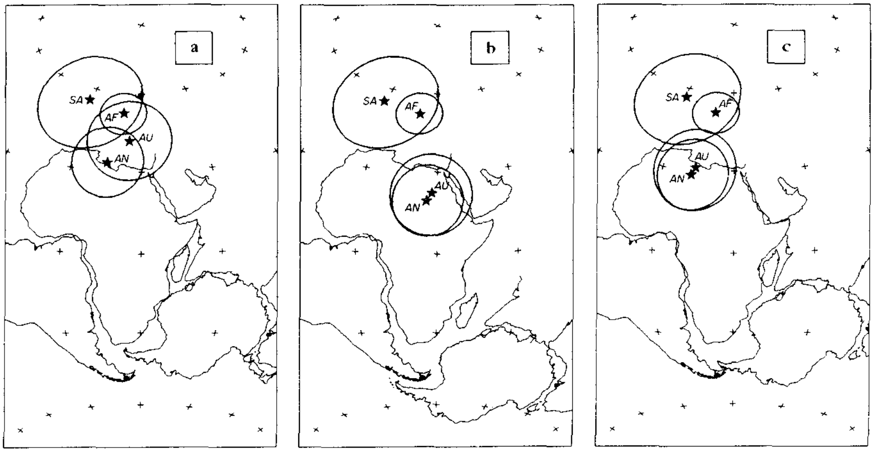

Fig. 26-3. Ordovician palaeomagnetic pole positions for South America (*SA*), Africa (*AF*), Australia (*AU*) and Antartica (*AN*), plotted with respect to reconstructions of: (a) Smith and Hallam (1970), (b) Barron et al. (1978) and (c) Norton and Sclater (1979). Individual poles are plotted in Fig. 26-4. GJ prefixes refer to Geophysical Journal palaeomagnetic data compilations (GJ list/pole).
Data Sources:
Africa: Table Mt. series, GJ4/32; Hasi-Mesaud sediments, GJ 15/141; Tassili sediments, Ileana (1971); Groupe de la Falaise d'Atar, Morris and Carmichael, 1978; Groupe des Plateaux d'Oujeft, Morris and Carmichael (1978); Blaubeker Formation. Kröner et al., 1980.
Antarctica: Mirnyy charnokites, GJ14/408; Sor Rondane intrusives, Zijderveld (1968).
Australia: Stairway sandstone, GJ14/393; Jinduckin Formation GJ14/395; Tumblagooda sandstone. GJ14/405.
South America: Salta sediments, GJ14/406; Urucum Formation, GJ15/132; Bolivia sediments, GJ12/140; N. Tilcara sediments, GJ 14/148; S. Tilcara sediments, GJ14/417; (Tilcara sediments may be Cambrian).

APW relative to Africa, 1200 to 450 Ma: West Gondwana

Late Precambrian to early Palaeozoic data for Africa are plotted with different symbols for different cratons in Fig. 26-4. Starting from the Ordovician group of poles directly north of present-day Africa and going backwards in time, upper and middle Cambrian poles group to the southwest of the West African Craton. The most reliable poles which are just older (i.e. Cambrian–Late Precambrian) than this group are *N2* and *DM* from the Kalahari and Congo cratons, respectively. The fact that the early and late Cambrian data are widely separated results in a polarity dilemma; the connecting APWP segment could extend eastwards and then south across Africa as shown in Fig. 26-4; alternatively, by reversing the polarity of *DM* and *N2*, the path would turn westwards (Kröner et al., 1980). Although it has a slightly longer total path length, the path shown in Fig. 26-4 incorporates

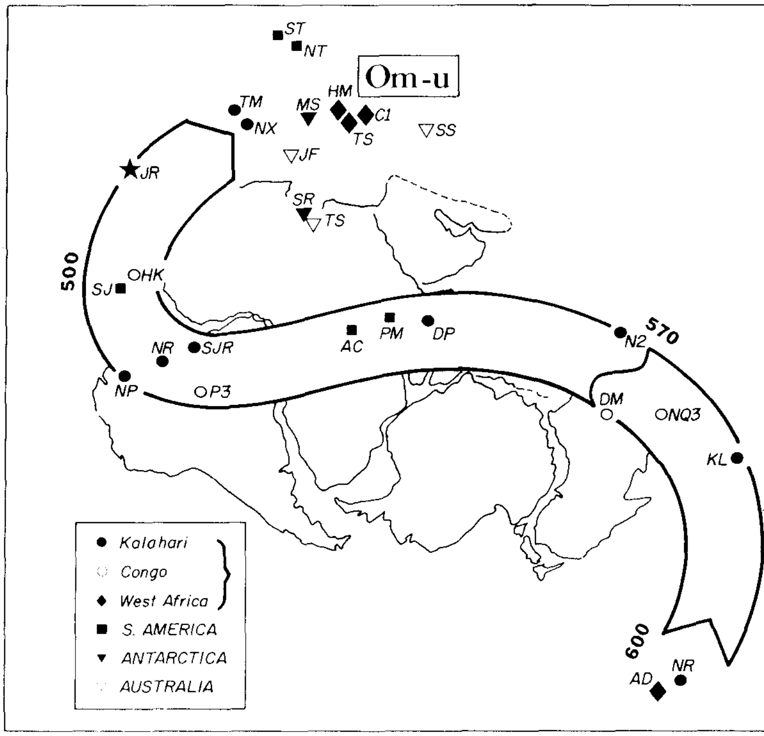

Fig. 26-4. Late Precambrian to early Palaeozoic APWP for West Gondwana. Smith-Hallam (1970) reconstruction, Africa fixed. Equal area projection, swathe width 15 degrees. Isotopic age code: K = K-Ar; R = Rb-Sr; U = U-Pb.
Reference to Ordovician poles as in Fig. 26-3 caption: TM = Table Mountain series; TS = Tassili sediments; C1 = Groupe de la Falaise d'Atar; NX = Blaubeker Formation; HM = Hasi-Mesaud sediments; MS = Mirnyy charnockites; SR = Sor Rondane intrusives; SS = Stairway sandstone; JF = Jinduckin Formation; TS = Tumblagooda sandstone; ST = S. Tilcara; NT = N. Tilcara; SL = Salta sediments; UR = Urucum Formation; BO = Bolivia sediments.

Pre-Ordovician data (Africa):			
JR	Cambrian—Ordovician	Jordan redbeds	Burek (1969)
HK	500	Hook intrusives	GJ9/132
NP	Cambrian	Nama Gp., overprint	Kröner et al. (1980)
NR	474, 479 K	Ntonya ring structure	GJ9/137
SJR	Precambrian—Cambrian	Sijarira Group	GJ12/149
P3	Cambrian ?	Plateau series	GJ14/562
DP	500—550	Doornpoort Fmn.	G14/529
N2	Precambrian—Cambrian	Nama Group, upper	Kröner et al. (1980)
DM	Precambrian—Cambrian	Mulden Group, lower	McWilliams and Kröner (1981)
NQ3	Precambrian—Cambrian ?	Nosib Gp., overprint	McWilliams and Kröner (1981)
KL	Precambrian-Cambrian	Klipheuvel Formation	Creer (1973)
NR	630 R	Ntonya ring structure	GJ9/137
AD	613 U	Adma diorite	Morel (1978)
Pre-Ordovician data, South America:			
SJ	Cambrian—Ordovician	Salta, Jujuy redbeds	GJ12/144
AC	Cambrian	Abra de Cajas	GJ14/419
PM	Cambrian	Purmamarca sediments	GJ14/421

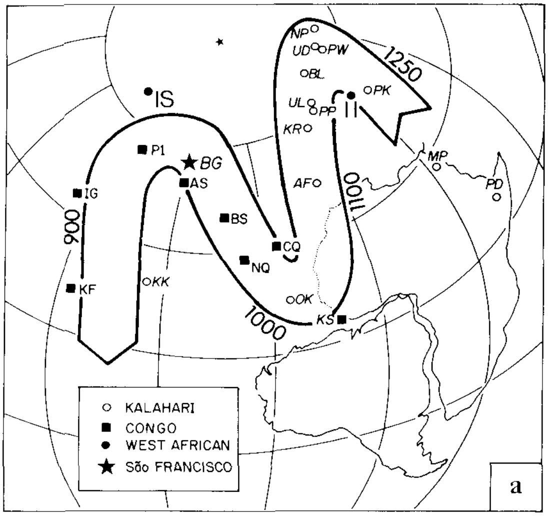

Fig. 26-5. a. ~ 1250 to ~ 900 Ma APWP West Gondwana. Equal area projection, swathe width 15 degrees at equator, Africa fixed. Age and source conventions as in Fig. 26-3 and 26-4 captions.

KF	> GL	Kigonero Flags	GJ14/544
KK	874 ± 41 K	Klein Karas dykes	GJ14/536
IG		Ikorongo Group	GJ14/537
PI	< AS	Plateau Series	GJ14/560
IS	< 1019 ± 32 R	Char Group	Morris and Carmichael (1978)
AS	> 940K	Abercorn sandstone	GJ14/559
BS	> KF	Bukoban sandstone	GJ14/543
BG	888 ± 11	Bambui Group	In: Vilas and Valencio (1978)
NQ	≅ CQ	Nosib quartzites	McWilliams and Kröner (1981)
CQ	1100—1050	Chela Group	Kröner (1975)
OK	1070 ± 20 R	O'okiep intrusives	GJ14/532
KS	1200—964 K	Kisii lavas	GJ14/522
AF	< 1020 ± 70 R	Auborus Formation	GJ14/535
KR	1200—1050	Koras Group	Briden et al. (1980)
PP	c. 1250 R	Premier kimberlite	Ito et al. (1978)
PK	c. 1250 R	Premier kimberlite	GJ10/197
UL	1115 R	Umkondo lavas	GJ8/154
BL	1265 R	Barby lavas	GJ14/533
UD	1140 R	Umkondo dolerites	GJ8/154
NP	c. 1250 R	National kimberlite	Ito et al. (1978).
PW	c. 1250 K	post-Waterberg dolerite	GJ8/155
II	> 1019 ± 32R	Char Group	Morris and Carmichael (1978)
MP	c. 1250 R	Montrose kimberlite	Ito et al. (1978)
PD	1310 ± 60 R	Pilansberg dykes	GJ/141

the available African data rather better than the previously suggested path of Kröner et al. (1980); in addition, the available data from South America can be incorporated without difficulty. Thus, while the polar gap and resultant polarity dilemma are not presently resolvable, the APWP segment shown in Fig. 26-4 is a reasonable working configuration which utilizes the

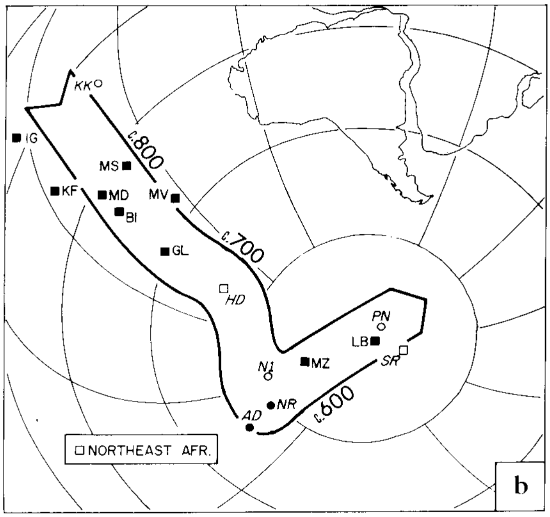

Fig. 26-5b. ~800 to ~600 Ma APWP for West Gondwana, details as in Fig. 26-5a. References for *IG*, *KK* and *KF* in caption to Fig. 26-5a. References for (AD) Adma diorite and (NR) Ntonya ring structure in Fig. 26-4 caption.

PN	653 ± 70 K	Pre-Nama dykes	GJ14/530
SR	> 540 K	Saboloka ring structure	Briden (1973)
LB		Lower Buanji Group	GJ14/538
MZ	743 ± 30 K	Mbozi complex	GJ14/539
N1	> 553 ± 13, < 686 ± 32R	Nama Group, lower	Kröner et al. (1980)
HD	c. 660	Hamdah monzonite	Kellogg (1978)
GL	813 ± 30 K	Gagwe lavas	GJ14/540
MV		Manyovu redbeds	GJ14/545
BI	806 ± 30 K	Bukoban intrusives	GJ14/541
MS		Malagaresi sandstone	GJ14/542
MD	940—680 K	Mbala dolerites	GJ14/558

available data. The tectonic implications are unaffected by a choice of this configuration or that of Kröner et al. (1980), as shown in a later section.

The older connecting segment for Africa in the 1200 to 650 Ma interval is shown in Fig. 26-5, with the same symbols as in the previous figure. The salient points of this polar track are a north—south path segment from c.900 to 650 Ma, followed by a double bending or "kink" in the path between 1150 and 900 Ma. While parts of this path and the previous 600 to 450 Ma segment are either sparsely populated or defined only by data from a single craton, the most important conclusion is that whenever data of approximately equal age from more than one craton are available, they are in reasonable agreement. There are four places along the 1200 to 450 Ma path where such a comparison between cratons is possible: in Ordovician times at the termination of Pan-African activity, at approximately the Precambrian—Cambrian boundary, at approximately 700 to 650 Ma, and at about 1100 to 1000 Ma, well before Pan-African orogenesis began. This good agreement (by Precambrian standards) suggests that one of the following hypotheses is tenable: (1) little or no relative cratonic displacement occurred between the three major African cratons during the entire 250—300 Ma Pan-African orogenic interval; (2) major intercratonic movements did occur, but did so in a geometrical configuration which did not produce significant differential AP motion; or (3) major intercratonic movements did occur which did produce significant (i.e. potentially detectable) differential AP motion, but in each case the motions were in two rapid opposing stages which resulted in little or no net observed differential AP motion. In the latter case, if the two-stage process had been stopped before the second (closing) phase occurred, it might be expected that differential apparent polar motion might be detected from the divergent APWPs which could result. In the second case, nothing would be detected because the palaeomagnetic and Euler poles were conveniently arranged so as to produce little or no AP motion. Considering the wide geographical and temporal distribution of the four areas of APWP agreement and the heterogeneous structural trends of the Pan-African belts, special situations such as the latter two cases seem difficult to imagine. The simplest explanation is that little relative motion has occurred between cratons and that all the data are incorporated in an APW track similar to that of Figs. 26-4 and 26-5.

Also plotted in Fig. 26-5 is a single pole [*BG*] from the Bambui Group (São Francisco craton, Brazil), corrected for the opening of the South Atlantic. Its position is not very different from the older APWP defined from Africa, suggesting that at about 100 Ma ago the São Francisco craton occupied approximately the same position with respect to the African cratons as it did in pre-Mesozoic times after the Pan-African events. No data are available from elsewhere in South America with ages between 1100 Ma and latest Precambrian—early Cambrian, and therefore the APW history for most of South America is undefined for this interval.

APW relative to Australia, 1000 to 450 Ma: East Gondwana

Australian palaeomagnetic data spanning the time interval ~ 1000 to ~ 450 Ma are plotted in various configurations in Figs. 26-6 and 26-8. Three structural provinces of middle Precambrian and older basement rocks separated by younger sedimentary basins are defined here for convenient analysis of the late Precambrian and early Palaeozoic data. Definition of these three provinces is applicable only in the latter part of the Precambrian; they are referred to here as the western province (Yilgarn and Pilbara blocks), central province (Kimberley, Arunta, Musgrave and Mt. Isa blocks and Amadeus Basin) and the southeast province (Gawler, Willyama and Mt. Painter blocks, Adelaide geosyncline).

Figure 26-6a shows that the Cambro-Ordovician data from the central and southeastern provinces are in good agreement with the two provinces

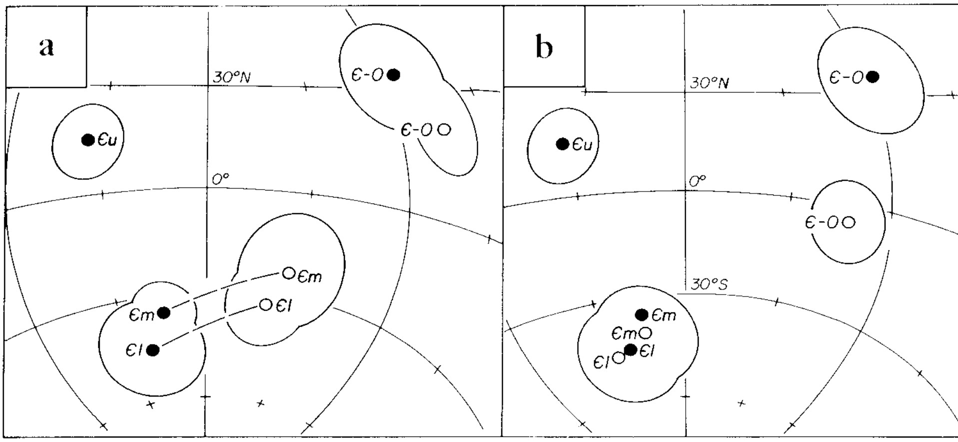

Fig. 26-6. Cambrian and Cambro-Ordovician data from southeast Australian province (open symbols) and central Australian province (solid symbols): (a) in situ; and (b) after rotation to close intervening younger basins. See text for rotation details. Reference key: [1] McWilliams and McElhinny (1980); [2] Klootwijk (1980I); [3] Klootwijk (1980II); [4] Klootwijk (1980III); [5] Kirshvink (1978); [6] B. J. J. Embleton, unpublished data. Pole symbols refer to late Precambrian—early Palaeozoic APWP, Fig. 26-8.

Data sources, southeast province:			
Lower Cambrian	*AD*	Aroona Dam sediments	GJ14/416
	HG	Hawker Group	[2]
	BC	Billy Ck.—Wirrealpa	[2]
	K1	Kangaroo Island	[4]
Middle Cambrian:		Lake Frome Group	GJ14/409
	LB	Lake Frome Group	[2]
Cambrian—Ordovician:	*W3*	Wooltana overprint	[1]
	M3	Merinjina overprint	[1]
		Bunyeroo Gorge overprint	[2]*
		Red Hill well overprint	[2]*
		Balcooracana overprint	[2]*
		Mt. Billy Ck. overprint	[2]*
	K2	Kangaroo Is. overprint	[4]
Data sources, central province			
Lower Cambrian	*AS*	Arumbera sandstone, upper	[5]
	TD	Todd R. dolomite, Eminta and Allua sandstones	[5]
Middle Cambrian:	*HR*	Hugh River shale	GJ14/409
	DC	Deception Formation	[3]
	IL	Illara sandstone	[3]
	TR	Tempe sandstone	[3]
	G1 and *2*	Giles Creek dolomite	[3]
Upper Cambrian:	*HF*	Hudson Formation	GJ14/413
		Pacoota sandstone	[6]
		Shannon Formation	[3]
Cambrian Ordovician:	*AY*	Areyonga overprint	[3]
	RR	Ross River overprint	[3]

*Averaged and plotted as "*DO*" in Fig. 26-8.

in their present relative position. However, middle and early Cambrian mean poles from the central and southeastern provinces are significantly different (Fig. 26-6a). This difference can be reconciled by postulating a middle to late Cambrian rifting episode which caused the fission of a smaller proto-Australian province into the three identifiable blocks (Fig. 26-7a; Harrington et al., 1973; McWilliams et al., in prep.). There are a multitude of possible Euler poles which will put the middle and early Cambrian palaeomagnetic data into statistical agreement.

A discussion of all the possible configurations of the three provinces in such a rifting scheme is beyond the scope of this review. In Figs. 26-6b and 26-7b a possible pre-rifting configuration is shown which is compatible with both the palaeomagnetic and geological evidence respectively. Keeping the central province fixed, the western and southeastern provinces are rotated 32 degrees clockwise and 38 degrees anticlockwise about Euler poles at 29°S, 125°E and 45°S, 145°E, respectively. These rotation poles are similar to those suggested by Harrington et al. (1973) and fit the western, central and southeastern provinces together in a smaller protocontinent. Besides explaining the discordant palaeomagnetic data, an interesting consequence is that the Adelaide geosyncline and the Amadeus Basin (which have pronounced similarities in stratigraphy and basin evolution) are brought much closer together (Fig. 26-7b).

Regardless of the exact details of province configuration, the palaeomagnetic data suggest that some relative motion between the 3 Australian provinces appears to have occurred in late Cambrian times. There is geological evidence to suggest that such a rifting event may have occurred, forming the Canning—Officer—Warburton basin system (see e.g. Harrington et al., 1973). In analyzing older data from these provinces, the same rotation must be applied to make intercratonic comparison possible. Adopting such a rotation explains some of the stratigraphic inconsistencies present in the unrotated data set (position of the Precambrian—Cambrian boundary, agreement of poles from glacial horizons, etc.; McWilliams, 1977). These inconsistencies arise if correlations between the Amadeus Basin and the Adelaide geosyncline are adopted (Preiss et al., 1978). Clearly either: (1) these correlations could be wrong (the discrepancy in pole position would then result from magnetization age differences); or (2) differential movement may have

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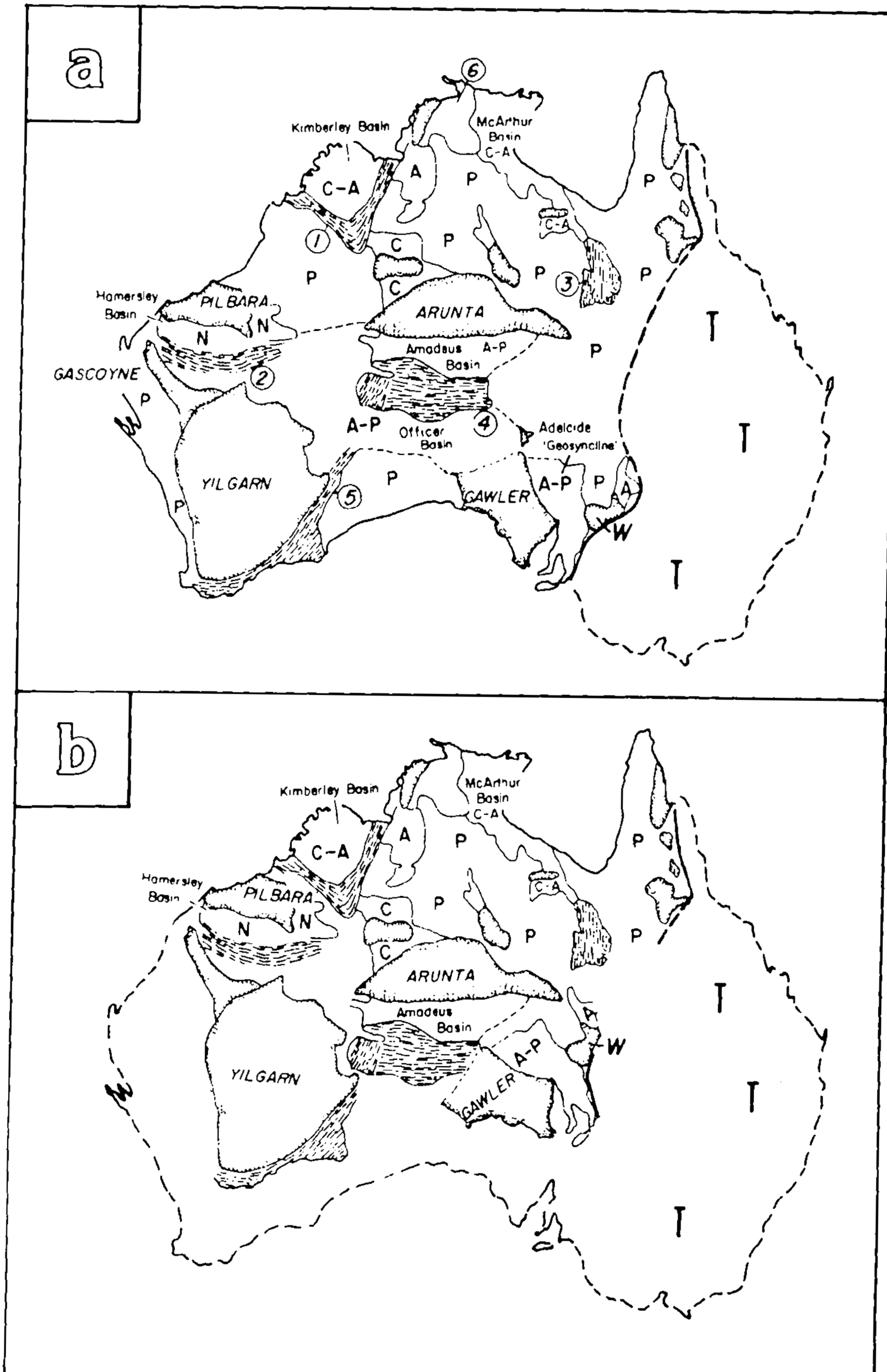
Fig. 26-7. a. Generalized structural map of Australia; b. Proto-Australia, with western and southeastern provinces rotated to close intervening younger sedimentary basins. Stable cratonic areas stippled, younger intervening mobile belts labelled *1* to *6*. *1* = Halls Creek—King Leopold mobile belt (~ 1800 Ma); *2* = Ophthalmian mobile belt (~ 1700 Ma); *3* = Mt. Isa geosyncline (~ 1800 to ~ 1400 Ma); *4* = Musgrave mobile belt (~ 1200 to ~ 1400 Ma); *5* = Albany—Fraser mobile belt (~ 1300 to ~ 1000 Ma); *6* = Pine Creek geosyncline. *T* = Tasman orogenic zone (Phanerozoic); *W* = Willyama block.
Age of sedimentary basins indicated as *N* (Nullaginian, ~ 2200 to ~ 1800 Ma), *C* (Carpentarian, ~ 1800 to ~ 1400 Ma), *A* (Adelaidean, ~ 1400 [?] to ~ 570 Ma). *P* (Phanerozoic).


a
Kimberley Basin
McArthur Basin
C-A
A
P
C
Hamersley Basin
PILBARA
N
ARUNTA
GASCOYNE
Amadeus Basin
A-P
Officer Basin
Adelaide 'Geosyncline'
YILGARN
GAWLER
W
T
1
2
3
4
5
6
b
Kimberley Basin
McArthur Basin
C-A
Hamersley Basin
PILBARA
N
ARUNTA
Amadeus Basin
A-P
YILGARN
GAWLER
W
T

occurred between provinces. Since the discrepancy in the early and middle Cambrian palaeomagnetic data is based upon reliable measurements with a good palaeontological basis, direct correlation of the younger data would appear to be valid and differential motion is a viable suggestion.

After compensating for early Palaeozoic intercratonic rifting, a single APWP can be constructed through all the available data (Fig. 26-8). This would suggest that prior to the rifting episode, no differential apparent polar motion of Pan-African age occurred within the proto-Australian craton. This is not surprising as there is relatively little isotopic and structural evidence for major Pan-African tectonism within Australia as compared to Africa, South America and India (Fig. 26-2). The data suggest that prior to late Cambrian times, a configuration such as shown in Fig. 26-7b may be applicable back to at least middle Proterozoic times. An older limit for this reconstruction is discussed in a following section.

The limited data available for India from c. 750 Ma ago to Cambrian times is in good agreement with the Australian data spanning the same time interval. This agreement is valid for either the Smith and Hallam (1970), Barron et al. (1978), or Norton and Sclater (1979) reconstructions, suggesting that Australia (in its pre-rifting configuration) and India moved as a single unit during late Precambrian times. The relationship of Antarctica to India and Australia in late Precambrian times is much less certain.

The assembly of Gondwana

The late Precambrian APWP segments for East and West Gondwana are superimposed in Fig. 26-9 relative to the Smith and Hallam (1970) reconstruction. It is apparent from the figure that the segments for East and West Gondwana are quite different and merge in early Palaeozoic times. This observation is also valid for the Barron et al. (1978) and Norton and Sclater (1979) reconstructions and suggests that East and West Gondwana were independent crustal units which collided in latest Precambrian—early Palaeozoic times to form Gondwana. Overall mean Ordovician data for East and West Gondwana are not significantly different, suggesting that the collision and suturing of the two crustal fragments was virtually complete by about this time.

Referring again to the distribution of Pan-African mobile belts (Fig. 26-2) it would appear that the Mozambique belt and its offshore counterparts were the approximate site of collision between East and West Gondwana. The African parts of the Mozambique belt apparently contain no suture (Kröner, 1977a). The collisional suture between the two Gondwana fragments must then lie to the east, possibly in coastal India and Antarctica and/or possibly in some of the submerged Indian Ocean plateaus. In any case, considering the present depth of exposure of the collisional zone, the suture itself would be cryptic and thereby quite difficult to observe.

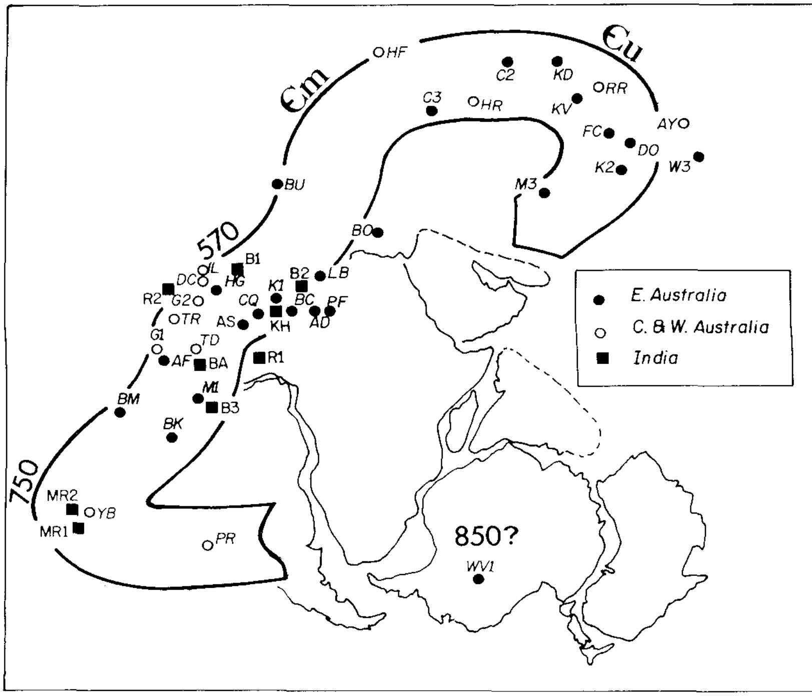


Fig. 26-8. Late Precambrian—early Palaeozoic APWP for East Gondwana. Smith and Hallam (1970) reconstruction, Africa fixed. Equal area projection, swathe width 15 degrees at equator. Palaeozoic poles not referenced below are listed in Fig. 26-6 caption. Reference key: [1] McWilliams (1977); [2] McWilliams and McElhinny (1980); [3] Karner Chamalaun, in reference [2]; [4] Athavale et al. (1970); [5] Klootwijk (1973); [6] McElhinny et al. (1978); [7] Wensink, (1972); [8] McElhinny (1969).

Australia:			
KV	Late Cambrian	King Island volcanics	[1]
KD	Late Cambrian	King Island dykes	[1]
C2	Late Cambrian	Cottons Breccia secondary	[1]
C3	Late Cambrian	Cottons Breccia secondary	[1]
BU	Cambrian ?	Bunyeroo Fmn., secondary	[2]
BO	Cambrian ?	Brachina Fmn., secondary	[3]
CQ	Cambrian ?	Copley quartzite secondary	[2]
AF	Late Precambrian	Angepena Formation	[2]
M1	Late Precambrian	Merinjina tillite	[2]
BM	Late Precambrian	Brachina Formation	[2]
BK	Late Precambrian	Brachina Formation	[3]
YB	750 R	Yilgarn B dykes	GJ16/149
PR	Late Precambrian	Pertatataka Formation	[1]
WV	c. 950—850	Wooltana volcanics	[2]
India:			
B1	Late Precambrian	Bhander sandstone	[4]
B2	Late Precambrian	Bhander sandstone	[5]
B3	Late Precambrian	Bhander sandstone	[6]
BA	Late Precambrian	Baghanwala sandstone	[7]
KH	Late Precambrian	Khewra sandstone	[8]
R1	Late Precambrian	Rewa sandstone	[4]
R2	Late Precambrian	Rewa sandstone	[6]
MR1	745 R	Malani rhyolite	[4]
MR2	745 R	Malani rhyolite	[5]

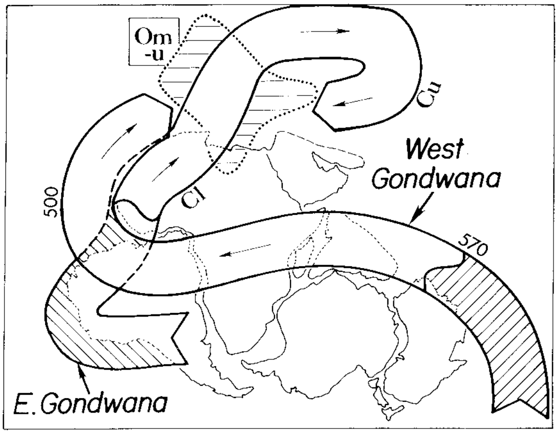


Fig. 26-9. Late Precambrian—early Palaeozoic APW paths for East and West Gondwana referenced to Smith and Hallam (1970) reconstruction. Precambrian parts of paths are ornamented with diagonal hachures. Irregular figure (horizontal hachures) is field containing Ordovician poles from all Gondwana continents (cf. Fig. 26-3).

To summarize the important features of the Pan-African evolution of Gondwana, the palaeomagnetic data would suggest that:

(1) The Kalahari, Congo and West African Cratons and possibly also the São Francisco Craton of South America may have maintained approximately their present relative positions within Gondwana before, during and after Pan-African orogenesis.

(2) After compensating for a minor amount of intercratonic movement within Australia, the western, central and eastern provinces appear to have moved together as a single unit (probably with Peninsular India and possibly with Antarctica) before and during Pan-African orogenesis.

(3) These two cratonic assemblages (East and West Gondwana) collided along the Mozambique belt in latest Precambrian times, resulting in the assembly of Gondwana in a configuration probably similar to one or another of the proposed reconstructions (e.g. Smith and Hallam, 1970; Barron et al., 1978; Norton and Sclater, 1979).

Inspection of the map of Pan-African mobile belts (Fig. 26-2) shows that if the above observations are true, belts *19*, *20*, *21* and *22* are not sites of cratonic collision preceded by large-scale convergence, while belt *22* (the Mozambique belt) did result from such a collision. The scale of relative movements in many of the other Pan-African belts, notably the Brazilian domain and the north African thermally reactivated domains, is unspecified. It is worth noting that a new palaeomagnetic result from plutonic rocks of the Arabian Shield (Kellogg, 1978), dated at about 660 Ma, is in excellent agreement with poles of similar age from the other African cratons (Fig. 26-4). Although based upon few samples, these preliminary results suggest that the Arabian Shield could also be considered as part of the overall African cratonic assemblage after about 660 Ma.

For the Pan-African domains we are thus left with the observation that of all the cases where palaeomagnetic data are of sufficient quality and quantity to make judgement feasible, only one of the Pan-African belts (the Mozambique belt) can be directly attributed to continental collision of previously widely separated cratons. The data suggest that several other Pan-African belts, notably the Damara belt and possibly the West Congo belt, and those separating the West African and Congo Cratons, did not form as a result of major intercratonic movements. Thus, it would appear that continental collision did play an important part in the evolution of parts of the Pan-African mobile belt system in Late Precambrian times. However, a major question to be resolved is whether the formation of the other parts of the Pan-African system are compatible with Mesozoic—Cenozoic-style plate tectonics, or whether a different mechanism is required (see Kröner, this volume, Chapter 3).

APW relative to East Gondwana, 1200 to 1800 Ma

High-quality Middle Proterozoic data for East Gondwana are few and are derived mainly from the Yilgarn and Gawler cratons of Australia. The 14 Australian palaeomagnetic poles for this interval are plotted in Fig. 26-10. The data have been corrected for the late Precambrian—early Palaeozoic intercratonic rifting episode proposed in an earlier section on the basis of reliable early Palaeozoic palaeomagnetic data. Three important features of such a middle Proterozoic reconstruction are notable. First, the very continuous ~1800 to ~1700 Ma old structural trends of Ophthalmian, King Leopold and Halls Creek mobile zones form an interconnecting network surrounding the Archaean Pilbara block and the Archaean—Lower Proterozoic Kimberley block. Second, the ~1400 to ~1200 Ma old Musgrave orogen is intersected by younger ~1300 to ~1000 Ma old structural trends of the Albany—Fraser mobile belt. Third, the Amadeus Basin and the Adelaide geosyncline are juxtaposed to form a single semi-arcuate, elongated feature with extremely similar mid-Late Precambrian sedimentary histories (see e.g. Preiss et al., 1978).

The palaeomagnetic data support the idea that a proto-Australia configuration such as shown in Fig. 26-7b may have been valid back as far as about 1700 Ma. The arguments for such a conclusion follow in a logical time order:

(1) The APWP segment of Fig. 26-10, together with the late Precambrian and Lower Palaeozoic data, illustrate that a case can be made for the Yilgarn and Gawler blocks remaining in approximately their late Precambrian proto-Australian configuration (Fig. 26-7b) from about 1700 to 1400 Ma. Thus,

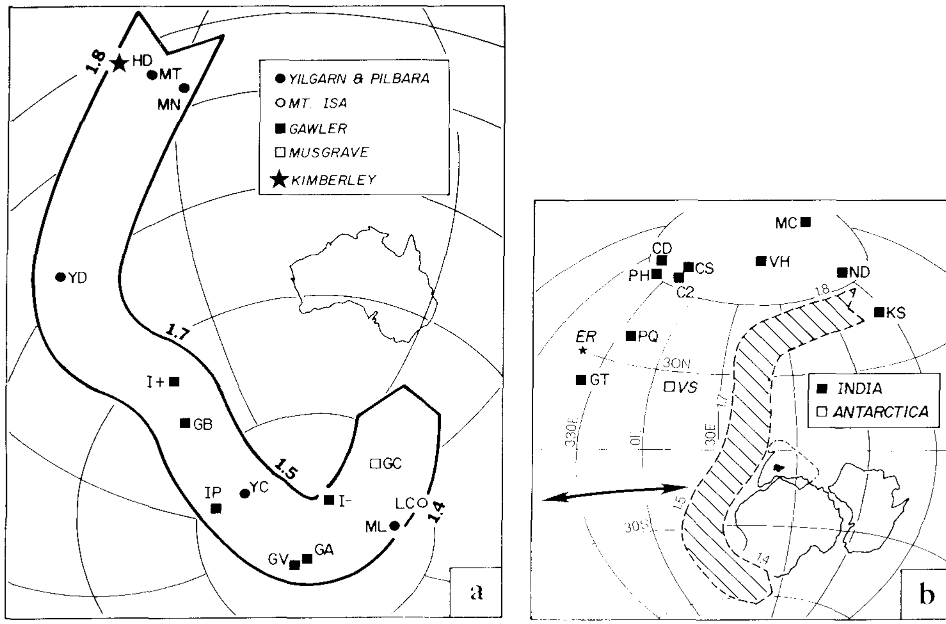


Fig. 26-10. a. Carpentarian (~ 1.8 to ~ 1.4 Ga) APWP for Australia. Equal area projection, Australia fixed. Swathe width 15 degrees at equator.

GC	1250—1140	Giles complex	GJ13/87
LC	1500 R	Lunch Creek lopolith	GJ14/430
ML	1500 R	Morawa lavas	GJ15/155
I-	1800—1500	Iron Monarch (-ve)	GJ9/146
GA	1500 ± 200 R	Gawler A dykes	GJ15/146
YC	1500 R	Yilgarn C dykes	GJ15/150
IP	1800—1500	Iron Prince deposit	GJ9/147
GB	1700 R	Gawler B dykes	GJ15/147
I +	1800—1500	Iron Monarch (+ ve)	GJ9/145
YD	c. 1700 R	Yilgarn D dykes	GJ15/151
HD	1800 R	Hart dolerite	GJ15/156
MT	c. 1800	Mt. Tom Price deposit	GJ10/174
MN	c. 1800	Mt. Newman deposit	GJ10/175

b. APWP of Fig. 26-10a (shaded) with Indian and Antarctic data superimposed. Smith-Hallam reconstruction; swathe width 15 degrees at equator. Reference key: [1] Athavale et al. (1972); [2] Embleton and Arriens (1973).

India:			
KS	1140	Kaimur sandstone	GJ8/150
MC	850—650	Mundwara complex	GJ7/58
VH	1200—900	Veldurthi hematite	GJ8/158
CS	1160	Cuddapah shale	[1]
C2	1160	Cuddapah sandstone	[1]
CD	1200—1100	Chifloor dyke	[1]
ND	< 1690—1560	Newer dolerites	GJ14/517
PQ	c. 1815	Pokhra quartzite	[1]
PH	c. 1815	Parkhuri hematite	[1]
GT	1830	Gwalior traps	GJ14/513
Antarctica:			
VS	1030 ± 220	Vestfold Hills dykes	[2]
Australia:			
ER	1760	Edith River volcanics	GJ1/145

the younger intervening Musgrave and Albany—Fraser belts could have resulted from small-amplitude intercratonic movements but probably did not result from collision of previously widely separated and independent units. East—west movements (relative to present geography) would be particularly difficult to detect, considering the applicable craton-pole geometry (Fig. 26-10).

(2) After 1700 Ma, the positions of the Pilbara and Kimberley blocks are somewhat constrained by the timing of activity in the Ophthalmian and Halls Creek—King Leopold mobile zones. There exist no major mobile zones younger than the Ophthalmian belt between the Pilbara and Yilgarn cratons. Therefore, these cratons can probably be considered as a single unit since at least 1700 Ma ago. Palaeomagnetic data from the Kimberley and Pilbara blocks are in good agreement at about 1800 Ma, suggesting that these two cratons have not moved greatly with respect to one another since that time. Again, east—west relative movements at ~ 1600—1400 Ma would be difficult to detect palaeomagnetically, but there is no geological evidence to suggest such activity.

At the younger end of the APWP segment, data from the Musgrave (pole *GC*) and Mt. Isa (pole *LC*) structural provinces suggests that these units can also be considered as part of the proto-Australian assemblage in the middle Proterozoic. By utilizing such a "leap-frog" approach (Gawler—Yilgarn, Yilgarn—Pilbara, Pilbara—Kimberley, then incorporating the data from the Mt. Isa block), a nearly complete peripheral loop is constructed around proto-Australia which constrains the scale of movements of the interior blocks (such as the Arunta block) to be small compared to present-day oceans. Again, it should be emphasized that small relative motions cannot be totally excluded, except in very favourable circumstances. The important conclusion from this part of the dataset is that the various Lower Proterozoic and Archaean fragments within proto-Australia appear to have maintained approximately similar relative positions from about 1700 Ma to the time of continental fission in the latest Precambrian—Early Palaeozoic and that these fragments are probably not cratonic blocks which were previously widely separated, totally independent units which nucleated in middle Proterozoic times.

The problem of where India and Antarctica fit in the framework of East Gondwana in middle Proterozoic times is much more difficult to assess. The limited data now available from these continents are plotted along with the data from proto-Australia in Fig. 26-10, with respect to possible East Gondwana reconstructions. The only prudent conclusion is that the data provide no positive evidence to suggest that, prior to about 1100 Ma ago, India and the proto-Australian assemblage were configured as a single unit. There is agreement between the c. 1800 Ma Indian pole positions (*GT*: Gwalior traps; *PH*: Parkhuri hematites; *PQ*: Pokhra quartzites) and the 1760 Ma pole from the Edith River volcanics of northern Australia (*ER*).

However, the latter pole is derived from NRM directions only; no demagnetization techniques were employed in the original study and thus the results cannot be used with much confidence. The relatively good internal agreement of the c. 1800 Ma data from India, together with the fact that these poles are considerably displaced from the c. 1800 Ma segment of the Australian path, suggests that the Indian and Australian paths are most probably not the same, and that the two proto-continents were independent units in middle Proterozoic times. The single pole from the Vestfold Hills dykes (*VS*) is not in agreement with any of the Indian or Australian data in any of the reconstructions; this could mean that East Antarctica too was an independent unit in mid-Late Proterozoic times, although a definitive conclusion is presently unwarranted, considering the available data.

APW relative to Australia, 2400 to 1800 Ma: The Yilgarn and Pilbara cratons

In the previous section, middle Proterozoic data from the structural units within proto-Australia were compared to determine whether large-scale relative displacements had occurred in middle or late Proterozoic times. In this section, Archaean to early Proterozoic data are compared to test whether major relative displacements occurred in early to middle Proterozoic times. Data are available only from the Yilgarn and Pilbara cratons (Fig. 26-11). A glance at the figure shows that a case can be made for the construction of a single APWP segment encompassing all of the Archaean and Lower Proterozoic data, as has been done by a number of authors (see e.g. McElhinny and Embleton, 1976; McElhinny and McWilliams, 1977; Embleton, 1978). The obvious conclusion of such a construction is that the Yilgarn and Pilbara cratons have retained approximately their present relative positions before, during and after the formation of the intervening Ophthalmian mobile belt and that a collisional model with large-scale displacements might be ruled out.

Careful inspection of the figure shows the difficulty in inferring a static model from the available data and illustrates the potential problems in assessing palaeomagnetically the scale of relative motion. As was discussed earlier, when the palaeomagnetic pole and the rotation pole are in close proximity, little or no apparent polar motion accompanies real lithospheric displacement. The available data are in two chronological groups, c. 2500 to 2300 Ma and c. 1800 Ma (Fig. 26-11). As the Ophthalmian mobile zone was active at about 1800 to 1700 Ma ago, only the older data (2500 to 2300 Ma) are applicable in a test for relative motions. Yet these data lie generally within Australia or to the southeast, in a configuration which would allow considerable undetected displacement of the Pilbara craton with respect to the Yilgarn craton. The younger data (*TP*, *MN*, *MI* from Pilbara; *KA*, *MG3*, *YF* from Yilgarn) are not useful in constraining any possible displacements,

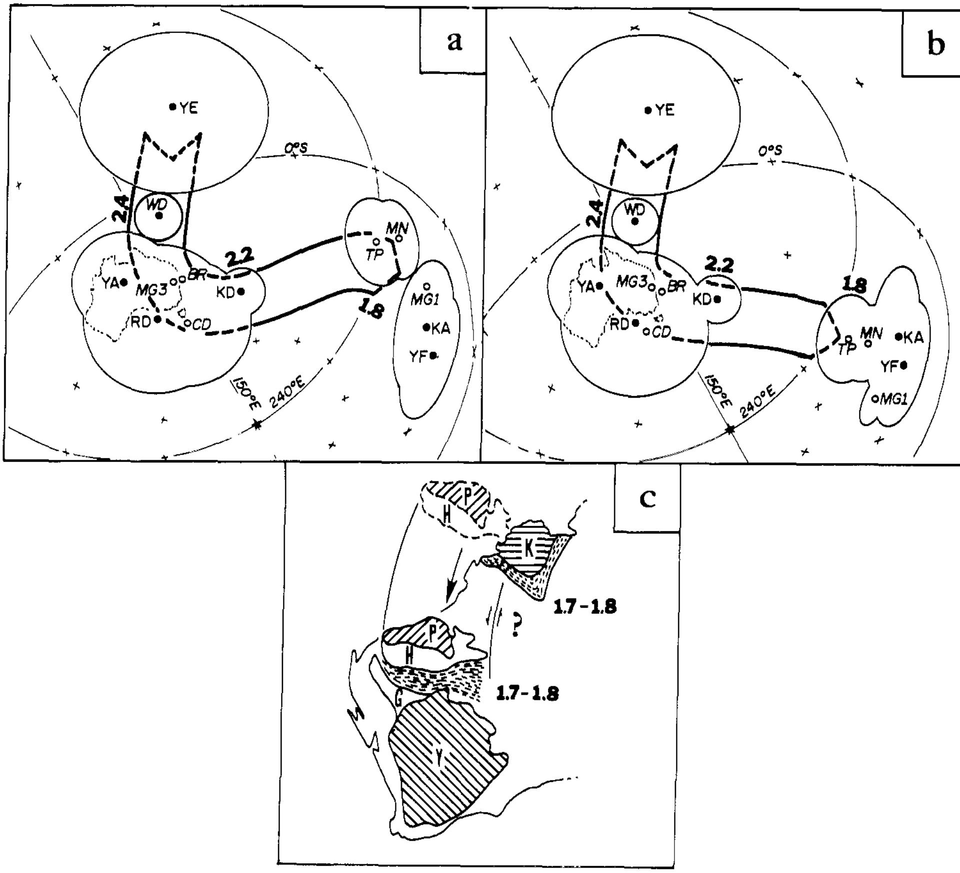


Fig. 26-11a. ~ 2.4 to ~ 1.8 Ga APWP for Yilgarn and Pilbara cratons. Equal area projection, Australia fixed. Swathe width 15 degrees at equator. Solid symbols, Yilgarn craton; open symbols, Pilbara craton. Error ellipses P = 0.05.

YF	c. 1700 R	Yilgarn F dykes	GJ15/153
KA		Koolyanobbing A deposit	Porath and Chamalaun (1968)
MG1	<MG3	Mt. Goldsworthy crust	GJ10/172
MN	c. 1800	Mt. Newman deposit	GJ10/175
TP	c. 1800	Mt. Tom Price deposit	GJ10/174
KD	2750–2200	Koolyanobbing Dowds Hill deposit	GJ10/173
CD	c. 2300	Cajaput dyke	Embleton (1978)
RD	c. 2500 R	Ravensthorpe dykes	GJ15/154
MG3	3000–2000	Mt. Goldsworthy lode	GJ10/170
BR	2329 ± 89 R	Black Range dyke	Embleton (1978)
YA	c. 2500 R	Yilgarn A dykes	GJ15/148
WD	2420 ± 30 R	Widgiemooltha dykes	GJ10/200
YE	c. 2500 R	Yilgarn E dykes	GJ15/152

b. As (a) but with Pilbara craton rotated 25 degrees clockwise about 30 S,150 E to illustrate how insensitive the older data are to local rotations.

c. To illustrate the effect that the rotation in (b) has upon the Pilbara craton: *Y* = Yilgarn; *G* = Gascoyne; *H* = Hamersley basin; *P* = Pilbara; *K* = Kimberley. Mobile belts ~ 1.8 to ~ 1.7 Ga: Ophthalmian and Halls Creek—King Leopold zones (see Fig. 26-7).

as their magnetization was acquired at approximately the same time as activity in the intervening mobile belt occurred. The pre-1800 Ma history of the Kimberley block is similarly unconstrained, as is shown schematically in Fig. 26-11.

Thus it is possible to make a case for a collisional origin for the Ophthalmian—King Leopold—Halls Creek mobile belt system. To keep a balanced perspective, it must be remembered that a multitude of rotation poles are possible, and to propose that the true rotation pole lies near Australia and thus produced no relative apparent polar motion requires an a priori special condition which may not be warranted. The previous discussion is meant to illustrate both sides of the argument; clearly the interpretations of McElhinny and Embleton (1976), McElhinny and McWilliams (1977) and Embleton (1978) remain valid.

APW relative to West Gondwana, 2300 to 1800 Ma

As with the Australian dataset, palaeomagnetic data from Africa decrease in quality and quantity with increasing age. The available African data for the interval ~ 2300 to ~ 1800 Ma are shown in Fig. 26-12, together with sparse South American data from the Guyana Shield corrected for the Mesozoic opening of the South Atlantic. The figure is similar to fig. 3 of Piper et al. (1973) and McElhinny and McWilliams (1977), with the addition of the Guyana data and new African poles from the Kaapvaal craton. For this interval it is possible to construct a simple APWP segment based upon the Kaapvaal data. When available, data from the West African and Congo cratons lie on or very near this reference path with the three African cratons in their present relative positions. Their positions along the Kaapvaal path are in accord with their age constraints; again, it must be remembered that some of these constraints are rather poorly defined, compared to younger data. Similarly, poles from the Guyana Shield fall on or near the reference path at approximately their correct age, with the Guyana Shield in its late Palaeozoic configuration with respect to Africa. No younger Precambrian data are available for the Guyana Shield.

The data show that it is possible to construct a single APWP through the available data without violating any of the age constraints. This agreement of early Proterozoic data, when coupled with the good agreement of poles from the Kalahari, Congo, West African and São Francisco cratons in middle to late Proterozoic times, makes a reasonable case for the absence of large-scale, independent relative movements between the older cratonic units of West Gondwana. To maintain a balanced perspective, the same caveats regarding data density, palaeomagnetic/Euler pole geometry and the inherent noise level in the data warrant cautious evaluation. Alternatively, the internal consistency of the data and the fact that a more complex explanation is not required argue for the simpler interpretation.

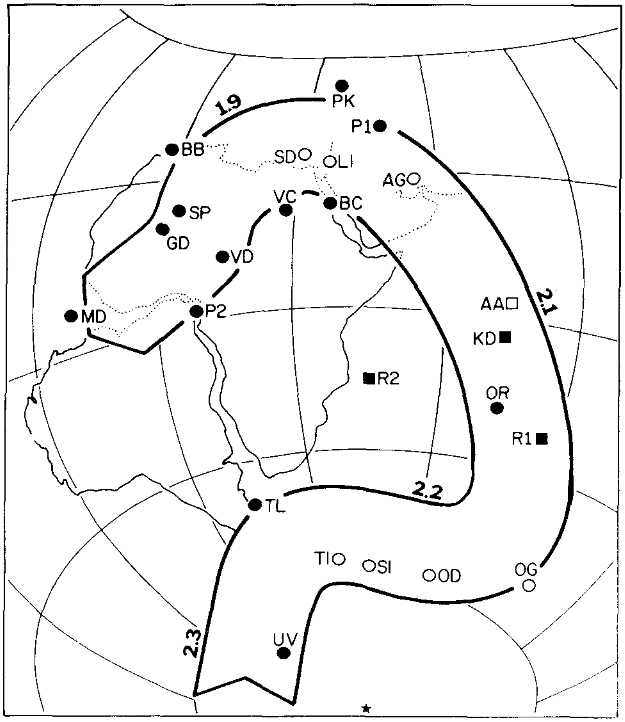


Fig. 26-12. ~ 2.3 to ~ 1.9 Ga APWP for West Gondwana. Equal area projection, Africa fixed. Swathe width 15 degrees at equator. Solid circles = Kaapvaal craton (predecessor of Kalahari craton, Fig. 26-2); open circles = West African craton; solid squares = Guyana craton; open square = Congo craton.

MD	1880 R	Mashonaland dolerite	GJ8/151
GD	c. 1900 R	Great Dyke overprint	GJ15/201
P2	c. 1850	Palabora complex	GJ15/209
VD	1650 ? K	Van Dyk kimberlite	GJ8/156
SP	ca. 1850 R	Sebanga Poort dyke	GJ15/203
BB	< c. 1850 R	Bubi, Crystal Spr. swarm	GJ15/202
VC	c. 1970 K	Vredefort ring complex	GJ12/163
SD	c. 1950 R	NW Sahara dolerites	in: Piper (1976)
LI	c. 1950 R	Losberg intrusion	in: Piper (1976)
BC	1950 ± 50 R	Bushveld complex	GJ1/142
PK	1750 ± 100 ? K	Premier kimberlite	GJ10/197
PI	c. 2000	Palabora complex	GJ15/208
AG	c. 1950 R	Aftout gabbro	in: Piper (1976)
AA	2079 R	Angola anorthosites	GJ14/426
KD	2090—2070	Kabaledo dolerites	GJ13/74
R1	2090—2070	Roraima dolerites	GJ10/160
R2	2090—2070	Roraima dolerites	GJ10/161
OR	> 1850R	Orange River lavas	GJ14/531
OG	2200 U	Obuasi greenstones	GJ14/510
OD	c. 2200	Obuasi dolerites	GJ14/511
SI	c. 2160 R	Syntectonic intrusives	in: Piper (1976)
TI	c. 2200	Tarkawaiian intrusives	GJ14/512
TL	c. 2250 R	Transvaal lavas	GJ15/207
UV	2300 ± 100 K	Upper Ventersdorp lavas	GJ9/157

DISCUSSION

This palaeomagnetic review of the southern continents, the accompanying reviews of North America and Europe, and previous summaries (Irving and McGlynn, this volume, Chapter 23; Poorter, this volume, Chapter 24; McElhinny and McWilliams, 1977; McElhinny and Embleton, 1976) show that considerable apparent polar wander occurred relative to the various continental nuclei in Proterozoic and latest Archaean times. On an axial geocentric dipole model, this observation leads to the conclusion that fragments of the continental crust have been in motion relative to the spin axis since at least 2600 Ma ago. Coupled with the available geologic evidence for the existence and activity of Archaean and Proterozoic continental margins (see e.g. Hoffmann, 1973; Windley and Smith, 1976; Erikson, 1979) and of oceanic crust (Anhaeusser et al., 1969; Wheeler and Gabrielse, 1972; Glikson and Lambert, 1973, 1976) the demonstrated existence of continental drift would imply that both continental and oceanic crust existed as separate entities at this time. Most Proterozoic APWPs are not constant in direction and rate of apparent polar wander, but rather are characterized by intervals of regular APW motion (tracks), bounded by abrupt changes in direction and rate (hairpins; Irving and Park, 1972). The existence of continental drift, the distinction of continental and oceanic crust and the observed irregular changes in apparent polar motion lend considerable support to the hypothesis that, in Proterozoic times, oceanic crust was being created and destroyed at active plate margins and that the continental crust moved in response to motion of the oceanic crust, perhaps as passive passengers on larger lithospheric plates. The mechanisms involved in the creation and destruction of oceanic lithosphere in older times are not yet clear, but it seems inescapable that a variation of Mesozoic—Cenozoic-style plate tectonics is the most plausible answer. In interpreting the palaeomagnetic data, the question at hand is: "given that some type of plate tectonics was operating in Proterozoic times in the oceanic domains, did the continental lithosphere respond to relative and/or absolute plate motions in the same way as did (and does) in Mesozoic—Cenozoic-style plate tectonics?" The motion of a modern continent reflects the net torque applied to it by forces which act along its margins (slab pull, ridge push of the attached oceanic plate) and along its bottom (mantle flow, mantle drag and associated forces). Many of these same forces were probably operative in Precambrian times, conceivably in different proportions. However, the response of the continental crust to applied torques (and hence its style of deformation) is not solely a function of the applied torques themselves, but is also a function of mechanical properties, especially crustal rigidity. Put simply, a change of the mechanical properties of the continental crust with time would lead to different responses (i.e. styles of deformation) given constant external driving forces. As discussed in previous sections, in all but a few cases the available Proterozoic palaeomagnetic data from the cratonic nuclei within West Gondwana and Australia can be contained on common APWP segments. The aberrant cases are latest Proterozoic in age. This could conceivably result from a real change in tectonic style, or simply from the fact that these younger data are of better quality in terms of determination of pole position and magnetization age. The suggestion that a common APWP explains the available data leads to an important question: What about the time intervals where there are no data, or where the available data are from only one structural unit? As a good example, the ~ 900 to ~ 650 Ma segment of the African path is defined by data from the Congo craton only. There are other similar examples in the datasets from Gondwana and elsewhere. Clearly, significant relative motions of the cratons for which there are no data are possible in the undocumented interval. Any proposed displacements are nearly always subject to an important condition: since the unconstrained intervals are bounded by time intervals for which there is good palaeomagnetic agreement, the displacements (large or small) must be such that the cratons *always* return to their previous relative configuration, within the limits of the data. Coupled with the geometrical considerations applicable in a particular reconstruction, this constraint is important in applying Mesozoic–Cenozoic-style plate tectonics to the Precambrian continental domains.

Turning to the alternative viewpoint, doctrinaire uniformitarianism dictates that to propose anything other than plate tectonics to explain Precambrian tectonics is *ad hoc.* A uniformitarian interpretation of the palaeomagnetic data would suggest that although major intercratonic movements did occur, the uncertainties inherent in applying Precambrian palaeomagnetic data preclude the observation of such movements. Further, the two constraints that the presently available data provide (either only small-scale motions or large-scale bidirectional motions which leave cratons in their approximate starting positions) have well-documented and widely accepted Phanerozoic analogs (e.g. Red Sea and Wilson cycle respectively). Uniformitarians therefore find little difficulty in suggesting that the Precambrian palaeomagnetic data are in accord with plate-tectonic models. This is essentially the interpretation of Burke et al. (1976), although their palaeomagnetic summary is undocumented and therefore the APWPs of their review remain cryptic.

But is uniformitarianism a useful or even applicable all-encompassing concept for Precambrian tectonics? Clearly not for early segregation and crustal formation in Archaean times. Most workers agree (whether they accept or reject Precambrian plate tectonics) that heat flow in Archaean times was significantly greater than at present, perhaps by a factor of two to three. Burke and Kidd (1978) argue that this increased heat flow was accompanied by vigorous plate activity to dissipate excess heat. Given that the mechanical response of continental crust to possible oceanic plate forces at its margins and to possible mantle forces at its base is a direct function of continental heat flow, it might be expected that even if plate tectonics was operating in the oceanic domains, response in the continental domains would change with time. This could then result in the continental tectonic style changing from a dominantly ductile, regional style to a dominantly rigid, more localized type of behaviour. Proterozoic tectonics may reflect the transitional period, with the occurrence of rigid or ductile behaviour depending on the temperature and strain rate in a given tectonic setting. An important point is that both types of behaviour would be possible at the same instant, but in different places.

Palaeomagnetic data from the southern continents show that many of the major Proterozoic orogenic belts contained within what much later became Gondwana are not necessarily the result of collision of previously widely separated, independent crustal fragments. Movement of the Precambrian continents relative to the spin axis is indicated by the data. Coupled with the evidence for the operation of some form of plate tectonics in the Precambrian oceanic domains, continental collision cannot be ruled out unless all the continental crust was arranged into a single supercontinent (Piper, 1976). The available palaeomagnetic data from the southern continents, North America (Irving and McGlynn, this volume, Chapter 23) and Europe (Poorter, this volume, Chapter 24) argue against such a long-lived supercontinent. While marginal tectonics may have taken a form similar to that envisaged for Mesozoic—Cenozoic-style plate tectonics, response of the Proterozoic continental crust to boundary and bottom torques was probably quite different. However, with the present state of knowledge, it is as premature to adopt a uniformitarian dogma regarding Proterozoic plate tectonics as it is to reject plate tectonics entirely. If the apparently widespread occurrence of intercratonic orogenesis without large-scale independent relative motion can be made compatible with modern plate tectonics, an acceptable solution may be in sight (see Kröner, this volume, Chapter 3). In this context, it is important to remember that such non-collisional Proterozoic belts might have their modern analogs hidden deep in the continental interior, well inboard from active continental margins. However, direct uniformitarian application of Wilson cycle tectonics to explain all Proterozoic orogenic zones is probably incorrect.

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F. Precambrian metallogeny and plate tectonics

Chapter 27

PRECAMBRIAN ORE DEPOSITS AND PLATE TECTONICS

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ABSTRACT

Relationships of Precambrian ore formations to various tectonic settings are determined largely from geological successions. Mineralized Archaean greenstone belts are interpreted as the upper structural levels of rocks emplaced at rift zones and "oceanic" spreading centres with minor island-arc volcanic rocks and local sediments derived from gneissic microcontinents. Predominantly silicic arc volcanism resulted from ocean-floor subduction, and microcontinent distribution was perhaps similar to that in present-day Indonesia and Melanesia. Arc—arc and arc—microcontinental collisions resulted in recumbent folding and thrusting of greenstone belts with generation of granodiorite and tonalite. Granodiorites were analogous to collision-related granites of Himalayan type. Sub-greenstone-belt oceanic crust was not preserved.

Archaean mineralization included placer gold-pyrite deposits at microcontinental margins, subduction-related volcanogenic deposits (Au, Cu) of Abitibi-type in island arcs and related Au-rich exhalative iron and carbonate deposits, sparsely preserved porphyry mineralization (Cu, Mo) and sulphide Ni-Cu deposits associated with komatiitic rocks formed in the upper levels of the rift or ocean rise-erupted, greenstone-belt volcanics. Early collision Cr-ore bodies were possibly emplaced during metamorphism resulting from arc collisions. Other possible collision-related mineralization includes Sb ores and pegmatites with Sn, W, Ta and Be.

Development of Proterozoic continents is indicated by shallow-marine sediments and widespread basins and was initially accompanied by greenstone-belt formation elsewhere. Proterozoic lineaments/megafractures indicate the presence of a major supercontinent with arc—continent collisional tectonics similar to those developed between Asia and bordering Taiwan, New Guinea and Timor in the late Cenozoic. Within the supercontinent, elongated narrow oceans opened and closed perhaps with little relative displacement from pre-drift positions. Ophiolitic and island-arc rocks are poorly preserved. High-grade metamorphic zones with linear features possibly developed through oblique collisions.

Proterozoic intracratonic rift zone and aulacogen mineralization included the Kupferschiefer-type stratiform ores of Cu, Co in the Copperbelt, Zn, Pb, Cu ores at Gamsberg, Namaqualand, South Africa and U deposits. Basic/ultramafic intrusives, e.g. Bushveld Complex, with Cr, Cu, Ni, Pt and carbonatites (Nb, P, U, Ce) were intruded in megafractures or lineaments. Hot-spot activity on Rondania, Brazil, had associated Sn mineralization. Ophiolitic deposits included podiform chromite in dunites while subduction-related ore bodies comprised porphyry Cu, Mo, and Kuroko-type Zn, Cu, Pb ores. Proterozoic ophiolitic and subduction-related deposits are comparable to Phanerozoic examples. Collision-related ores included the Rössing uranium-bearing granites and syn- to post-collision granitic rocks elsewhere with Sn, W, Nb, Ta, Li and Be.

INTRODUCTION

Despite voluminous literature on Precambrian geology and on ore bodies of Precambrian age, relatively few authors have attempted to relate the latter to the tectonic setting during mineralization. There are two approaches: either the tectonic setting can be inferred from comparison of Precambrian ore bodies themselves with those of late Phanerozoic deposits formed in relatively well-understood tectonic settings (e.g. Hutchinson, 1973), or the tectonic setting can be inferred from the overall geology and the mineralization related to the tectonic setting of the host rocks. Here we follow the latter approach, largely because far more information relevant to tectonic setting can be deduced from Precambrian successions than from the ore deposits themselves. If some of our inferences and our broadly uniformitarian views seem speculative, it should be remembered that the plate-tectonic concept became acceptable, even for Cenozoic times, only some ten years ago. This was despite the fact that the process is presently visible and has been there for interpretation since the days of William Smith, or perhaps rather of Alfred Wegener, who pioneered the continental drift hypothesis. Determining tectonic settings of Precambrian mineralization as far back as the Archaean, with its fragmentary preservation of rocks and structures, must therefore involve a certain amount of speculation based on available data.

TECTONIC SETTINGS IN THE ARCHAEAN

A number of authors have advanced hypotheses of Archaean crustal evolution which contrasted with that becoming accepted for the Cenozoic and much of Phanerozoic time. These hypotheses which included the vertical subsiding basin model for greenstone belts (Glikson, 1971), a rift-valley—marginal-basin model (Windley, 1973), basalt underplating (Fyfe, 1974) and sub-crustal accretion (Holland and Lambert, 1975), and the ensialic basin model of Groves et al. (1978), are discussed in various chapters in this volume.

In the late 1970's, hypotheses favouring a modified plate-tectonic evolution for the Archaean have become increasingly popular. Glikson (1976b) explained the Archaean greenstone belts as island arcs, as suggested by Goodwin (1971), with a distribution comparable to that in the Fiji area of Melanesia today, and Arth and Hanson (1975) emphasized the similarity in composition noted by some earlier workers between greenstone-belt basalts and island-arc tholeiite basalts of Cenozoic volcanic arcs. However, Hawkesworth and O'Nions (1977) suggested on the basis of major and trace elements that both the komatiite-tholeiite and much of the overlying calc-alkaline volcanics of greenstone belts in Southern Africa developed in a rift environment, while the uppermost silicic calc-alkaline sequences were derived from a basaltic or eclogitic source at depth. Greenstone belts have also been interpreted as "marginal basin" deposits formed in back-arc environments on continental (Tarney et al., 1976) or oceanic (Windley and Smith, 1978) crust and the high-grade gneiss terrains as Andean-type batholiths and their metamorphosed equivalents formed beneath subduction-related volcanic arcs; deformation of the greenstone-belt sequences has been related to closure of the marginal basins. In contrast to these views, Fyfe (1978) has argued from geochemical mass-balance considerations that the volume of sialic crust has decreased with time and that in the Archaean a type of hot-spot tectonics operated (*see also Lambert, this volume, Chapter 18, ed.*) However, Fyfe's (1978) hypothesis requires an assumption concerning subduction of sediments which is by no means universally accepted (*e.g. Moorbath and Taylor, this volume, Chapter 29, ed.*)

In our opinion the closest analogy between Archaean and Phanerozoic crustal processes is that suggested by Burke et al. (1976) who emphasized the significance of crustal shortening within greenstone belts. They explained the greenstone-granodiorite belts as island-arc and adjacent ocean-basin rocks and the gneissic rocks as roots of volcanic arcs and as syntectonic plutons formed by partial melting during collision in tectonic settings in part analogous to that of Tibet.

In order to provide a "plate"-tectonic framework for discussing Archaean mineralization, we attempt here to combine the evidence for collision provided by tectonic repetition and recumbent folding in greenstone belts with that for the postulated rift-related generation of most of the volcanic rocks. We suggest that the lower komatiite/tholeiite and much of the overlying calc-alkaline rocks were emplaced in rift zones, initially forming within older gneiss terrains and subsequently in marine "oceanic" spreading centres as the gneissic microcontinents separated (*see also Goodwin, this volume, Chapter 5, and Kröner, this volume, Chapter 3, ed.*) The greenstone-belt rocks at their margins thus overlie gneiss (Bickle et al., 1975; Beckinsale et al., in press), and elsewhere either lie on unexposed oceanic crust or themselves form the oceanic crust. Rifting and creation of the Archaean greenstone belt crust required crustal shortening elsewhere, accommodated by subduction of the greenstone-belt rocks and generation of the uppermost silicic calc-alkaline rocks. These were erupted to form volcanic arcs overlying the older more extensive tholeiitic and calc-alkaline sequences; some of the intrusive tonalites in greenstone belts were possibly deeper-level equivalents of the magmatic arc rocks.

Collision of the volcanic arcs and microcontinents resulted in recumbent folding and thrusting with consequent tectonic repetition of the greenstone-belt successions, as recorded in Botswana (Coward et al., 1976) and Greenland, with crustal thickening and local involvement of older gneisses of the "microcontinents". Large volumes of granodiorite and tonalite were generated in the collision from partial melting of the lower levels of the greenstone-belt rocks, of underlying basic and ultrabasic rocks not now exposed, and of older gneisses. These granodiorites were probably analogous in tectonic setting, but not in composition, to collision-related granite sheets within the subducting continental plate in the Himalayas, rather than with the inferred deep-level plutons of Tibet as suggested by Burke et al. (1976). The predominantly low initial $^{87}Sr/^{86}Sr$ ratios reflect the only slightly greater age ($<$100 Ma) of the mantle-derived greenstone-belt source rocks (Moorbath 1976, 1978).

The granodiorites rose as extensive syntectonic sheets into, and commonly engulfed large areas of, the deformed greenstone belts as in Greenland (Myers, 1976), at the same time undergoing regional metamorphism (Collerson et al., 1976) similar to Palaeogene metamorphism in the Himalayas. The enormous volumes of granodioritic material in Archaean terrains compared with those of syn-collision granites in Phanerozoic collision belts may reflect either a faster rate of plate convergence or partial melting at a relatively high structural level due to a higher geothermal gradient.

The high-grade (granulite facies) gneissic terrains lying between the greenstone-tonalite belts can most easily be explained as deeper structural levels of the collision zones, consisting very largely of collision-generated gneissic tonalite. The widespread basic and ultrabasic layered sheets within the greenstone belts and locally preserved in gneissic terrains predate granodiorite emplacement and were possibly intruded as sills within the greenstone belts prior to collision. Alternatively, evidence for major pre-consolidation deformation of anorthosites in Greenland (Myers, 1976) suggests that some may have been emplaced before the granodiorites in the early stages of collision.

TECTONIC SETTINGS IN THE PROTEROZOIC

Development of continents by tectonic accretion of Archaean greenstone-gneiss terrains is indicated by the presence of belts of shallow-marine orthoquartzites and carbonates from around 2300 Ma onwards, and in South Africa by deposition of the widespread, in part non-marine, Witwatersrand succession 2600 Ma ago. The presence of greenstone belts 2200 Ma old, appreciably younger than the Witwatersrand succession, suggests that consolidation of these continental areas had taken place in some areas while Archaean-type tectonic processes continued elsewhere (Burke et al., 1976).

Palaeomagnetic indications (Piper, 1974) and the distribution of Proterozoic lineaments (Davies and Windley, 1976; Glikson, 1976b) suggest the presence of a major "supercontinent" from about 2200 to about 1000 Ma ago. This has led many authors to argue that subduction and collision processes were not operating during the Proterozoic and that deformation, high-grade metamorphism and pluton emplacement took place in ensialic rift and shear zones (Sutton and Watson, 1974; Davies and Windley, 1976; Glikson, 1976b). However, Dewey and Burke (1973) have argued that a mid-Proterozoic ocean closed along the Circum—Ungava line in North America with consequent continental collision about 1800 Ma ago and, more convincingly, for a similar collision orogeny for the Grenville terrain in the late Proterozoic about 1000 Ma ago. (*For alternative interpretations of these belts see Baragar and Scoates, this volume, Chapter 12; Dimroth, this volume, Chapter 13, and Baer, this volume, Chapter 14, ed.*)

We follow the concept of Dewey and Burke (1973) for the collision-related nature of at least part of the Circum—Ungava and Grenville Provinces but suggest that, in general, in collision belts the zones of high-grade metamorphism, deformation and syn- to late-tectonic potash granites developed on the subducting continental plate and thus were analogous to the early Tertiary events in the rocks of the subducting Indian continent in the Himalayas, rather than to those on the overriding plate in Tibet. Evidence for orogeny and "reworking" of basement on the original overriding plate can be explained by syn-collision reversal of subduction polarity described by Roeder (1973) in Cenozoic orogens. A major difference from Archaean orogenies was the generation in the Proterozoic of syn-continental collision granites, with mostly high initial $^{87}Sr/^{86}Sr$ ratios, as opposed to the voluminous Archaean collision-related tonalitic magmas with low initial $^{87}Sr/^{86}Sr$ ratios.

Evidence for a supercontinent can be reconciled with that for orogenic activity within it by postulated opening of narrow oceans which subsequently closed, with little relative displacement of the collided continental margins from their original pre-rift positions. This could explain the presence of mid-Proterozoic high-grade gneiss terrains such as that at Broken Hill in Australia, the metamorphism being analogous to that of the Himalayas, and the absence of ophiolitic rocks reflecting tight suturing and tectonic juxtaposition of continental rocks of the overriding and subducting plates.

Evidence for late Proterozoic island-arc rocks, mostly in tectonic contact with ophiolitic or continental rocks, is found in some Pan-African belts (Clifford, 1970), in Saudi Arabia and northwestern Europe, suggesting that by 1000 Ma or earlier, subduction and collision were not unusual within the supposed supercontinent (Burke and Dewey, 1971). It is improbable that during mid-Proterozoic time alone plate-tectonic processes were not operating, and the presence of a possible volcanic-arc assemblage about 1350 Ma old, bordering the Damara Belt in southern Africa (Watters, 1976), suggests the presence of subducting ocean floor. It is also probable that ocean-floor subduction, island-arc formation and arc—continent collision took place mostly around the margin of "Pangaea", comparable to that bordering Asia in Taiwan, or Australia in New Guinea and Timor in the late Cenozoic. The scarcity of mid-Proterozoic ophiolite and recognizable volcanic-arc assemblages may reflect extreme post-collision elevation, resulting in erosion of obducted ophiolites and exposure of deep-level "roots" of the arc in areas subsequently deformed in late Proterozoic and Palaeozoic arc—continent collision.

The zones of high-grade metamorphism with abundant linear features common in the Proterozoic, for example the Limpopo Belt, are commonly interpreted in terms of intracontinental transform faults. An alternative explanation is that these belts developed in zones of oblique collision, with late-collision transform faults analogous to those in Taiwan cutting high-grade metamorphic rocks formed at depth during earlier stages of collision.

ARCHAEAN MINERALIZATION

Microcontinental margin ore bodies

Placer gold-pyrite deposits

Deposits of this type, resembling the much larger ore bodies of intracratonic basins described later, occur in the greenstone belts of South Africa and Rhodesia (Anhaeusser, 1976a). They occur in conglomerate-quartzite beds suggesting a non-volcanic continental source. Unlike the more extensive intracratonic deposits, they may be interpreted as having accumulated within greenstone belts but on the margins of small continental areas composed of older deformed greenstone-granite belts which formed the source areas. However, no older belts have been identified with certainty in these areas.

According to Anhaeusser, none of these South African auriferous placer deposits have yielded much gold. In the Uitkyk Formation in the Pietersburg greenstone belt, gold occurs with pyrite in polymictic conglomerates. In Rhodesia, placer gold deposits occur mainly in sediments of the Shamvaian Group which, at the Coronation Mine, consist of a rubble-bed of quartz pebbles, clay seams and films of clay around brecciated quartz. Banket ore at the Eldorado Mine is similar to that in the Witwatersrand but the gold is possibly epigenetic and introduced into shear zones (Stagman, 1961). The original gold distribution, however, was probably influenced by sedimentological controls similar to those in the Witwatersrand (Anhaeusser, 1976a).

Ore bodies of intracratonic basins

Quartz-pebble gold-uranium deposits

Detailed sedimentological work carried out on the Witwatersrand deposits of South Africa, dated at 2700 Ma (Rundle and Snelling, 1977), suggests an intracontinental or marine basin up to 350 km long by 200 km wide, fed by streams draining Archaean rocks to the northwest. The goldfields were formed as fluvial fans that built up at several points along the periphery of the basin, the gold and uranium being partly detrital and partly biochemical deposits formed by interaction with carbon-forming algal mats (Pretorius, 1975; Minter, 1976). Although there is some evidence for syndeposition faulting along the NW rim of the basin (Pretorius, 1975), there are few features typical of an aulacogen and the tectonic environment for deposition of the overlying Ventersdorp lavas is also uncertain.

Rift- and subduction-related ore bodies of the greenstone belts

In terms of Phanerozoic settings the type of subduction-related mineralization depends on a variety of factors including the stage of evolution of the arc (Mitchell and Bell, 1973), whether continental crust is present in the arc (Horikoshi, 1976; Garson and Mitchell, 1977) and the position of the arc and back-arc magmatic belts relative to the continental margin (Mitchell, 1976; Mitchell and Beckinsale, in prep.). The identification of Phanerozoic ore-deposition environments, and especially those of volcanogenic massive sulphide deposition, can also be determined from the trace element geochemistry of associated igneous host rocks (Pearce and Gale, 1977).

In the Archaean greenstone belts tectonic settings become increasingly difficult to determine. The Abitibi greenstone belt of the Canadian Shield, the largest Archaean greenstone belt in the world, probably provides the best evidence for the identification of volcanogenic sedimentary mineralization in either an island-arc or marginal-basin type setting, but much supplementary evidence comes from other greenstone belts in southern Africa, Rhodesia, Australia and Guyana.

Studies of the geochemistry of these greenstone-belt volcanic rocks suggest that they are broadly similar to Phanerozoic island-arc volcanic suites (Glikson, 1971; Pearce and Cann, 1973; Gunn, 1976; Winchester and Floyd, 1976). The average Archaean andesite is a typically low-K, low-Sr andesite of unmistakable island-arc type (Gunn, 1976). In Canadian belts a progressive increase in Al, K, Sr and Ba upwards in the pile relates to increasing calc-alkali features characteristic of a time sequence similar to that in recent island arcs (White et al., 1971) but with some differences (Glikson, 1976c) such as the presence in the Archaean belts of komatiites, a relative abundance of Ni, Cr and Co and a relative rarity of alkaline to shoshonitic varieties. However, komatiites are known in younger fold belts such as in Newfoundland (Gale, 1973) and in a few Phanerozoic environments (Brooks and Hart, 1974) while Ni, Cr and Co abundances are similar to those in Mesozoic marginal basins (Tarney et al., 1976).

Recently, however, detailed petrographic and geochemical studies have been carried out by MacGeehan and Maclean (1980) in the Matagami District of Quebec which suggest that the apparent "calc-alkaline" affinity of some of the large Archaean volcanic centres is due to widespread hydrothermal alteration of a bimodal basalt-rhyolite suite of tholeiitic affinity, emplaced in a tensional tectonic setting. This alteration is the result of localized sub-seafloor geothermal activity at volcanic centres related to the genesis of massive sulphide ores of Noranda-type.

Tarney et al. (1976) suggest that Archaean greenstone belts have developed in a back-arc marginal-basin position comparable to the "Rocas Verdes" complex in southern Chile. Pearce and Gale (1977) also favour a marginal-basin environment for arc deposition. Although marginal basins are evidently related to compressional subduction processes, it is emphasized that they formed by emplacement of new oceanic crust under tensional conditions. The evidence by MacGeehan and MacLean for a tensional environment for the Matagami volcanic rocks, and those of several other districts in the Abitibi belt, therefore supports the concept of Tarney et al. (1976) and also of Windley (this volume, Chapter 1). We consider, however, that there must have been a somewhat greater development of oceanic crust in these back-arc marginal basins (Fig. 27-1), or rifts, than hitherto envisaged.

Volcanogenic stratabound massive sulphides (Cu, Zn, Au, Ag)

Within the Abitibi belt eleven elliptical volcanic complexes (Fig. 27-2a, b) are situated close to the forelands of the belt, with a median part consisting of uniform tholeiitic basalts, fine-grained clastics and later granitic batholiths. This pattern may reflect the original linear distribution of a series of strato volcanoes (Goodwin and Ridler, 1970). Lower volcanic piles consist dominantly of basalt flows and gabbroic intrusions while andesitic rocks increase in quantity upwards and felsic rhyolites to dacites predominate in the upper levels. Komatiitic rocks occur as ultramafic flows. Sediments are largely turbiditic suggesting rapid accumulation in a deep basin; shallow- to deep-water transitions represent remnants of the original basins and trenches (Goodwin, 1973).

Lithological associations common to Cu-bearing sulphide deposits are felsic volcanic rocks reflecting central vent eruptions as in Phanerozoic Kuroko-type deposits, nearby sulphide facies iron formation, e.g. Noranda, Malartic, Matagami and Chibougammau reflecting reducing environments, and scarce coarse clastics off-shore reflecting deeper-water sites of accumulation. Au occurences are directly associated with carbonate facies iron formations either at felsic volcanic-intrusive centres or their margins. Anhaeusser's (1976a) schematic diagram of an Archaean volcano-sedimentary complex (Fig. 27-3) illustrating the possible relationship of gold and sulphide mineralization to various parts of the complex is similar to Colley's model (1976) of interrelations between various types of Kuroko deposit, while the peripheral exhalative mineralization has analogues in the Ordovician island-arc centres of central Norway (Halls et al., 1977).

Although Hutchinson (1973) and now Macgeehan and MacLean (1980) noted significant differences in tectonic setting between Kuroko-type

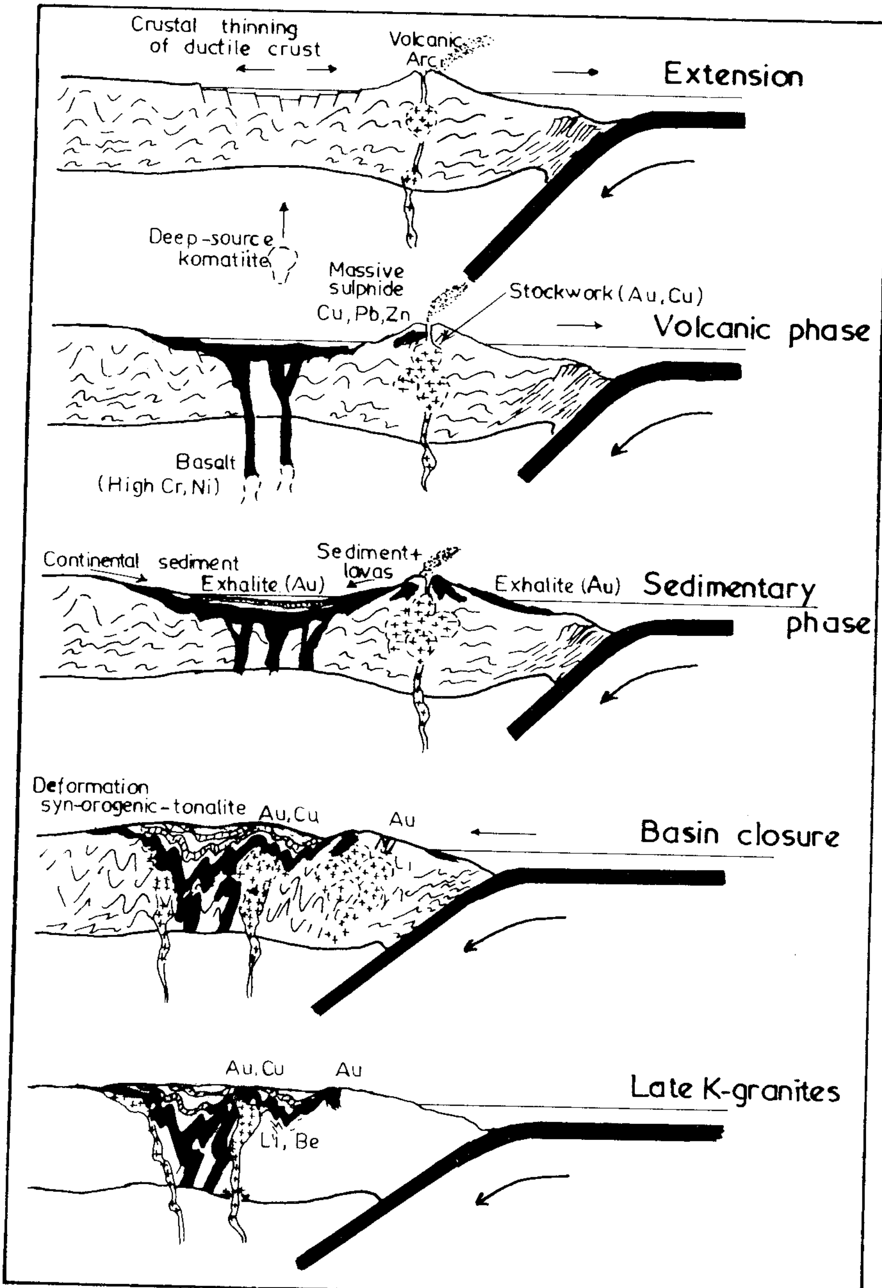

Fig. 27-1. Suggested development of an Archaean greenstone belt in a back-arc marginal basin similar to that suggested for the Mesozoic "Rocas Verdes" complex in S. Chile, after Tarney et al., 1976. Reproduced by permission of J. Wiley, London. Note that synorogenic tonalite can be interpreted as collision-related magma.

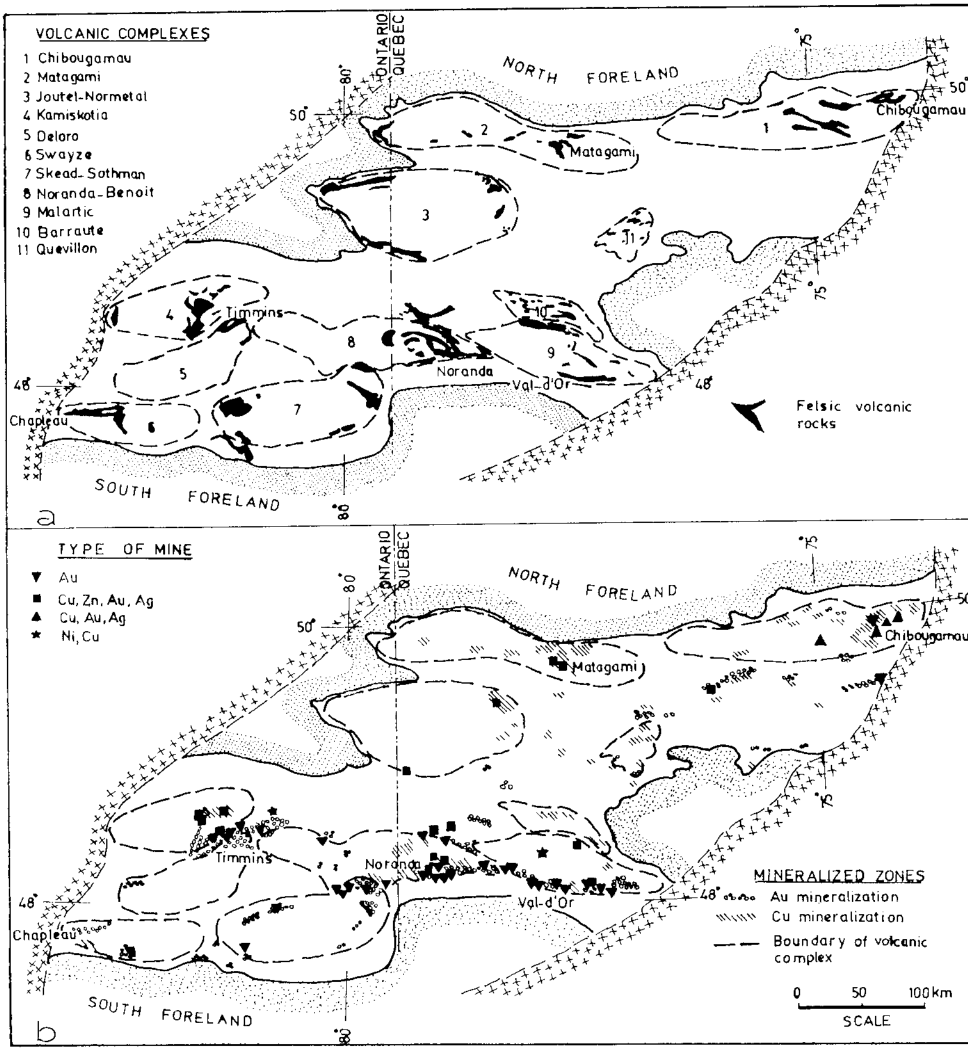

Fig. 27-2. a. Distribution of volcanic complexes in Abitibi Orogen (Goodwin and Ridler, 1970). Present elliptical shape of presumed originally circular volcanic complexes is attributed to compression folding. b. Metallogenic relations in Abitibi Orogen after Goodwin and Ridler, 1970. Reproduced by permission of the Geological Survey of Canada.

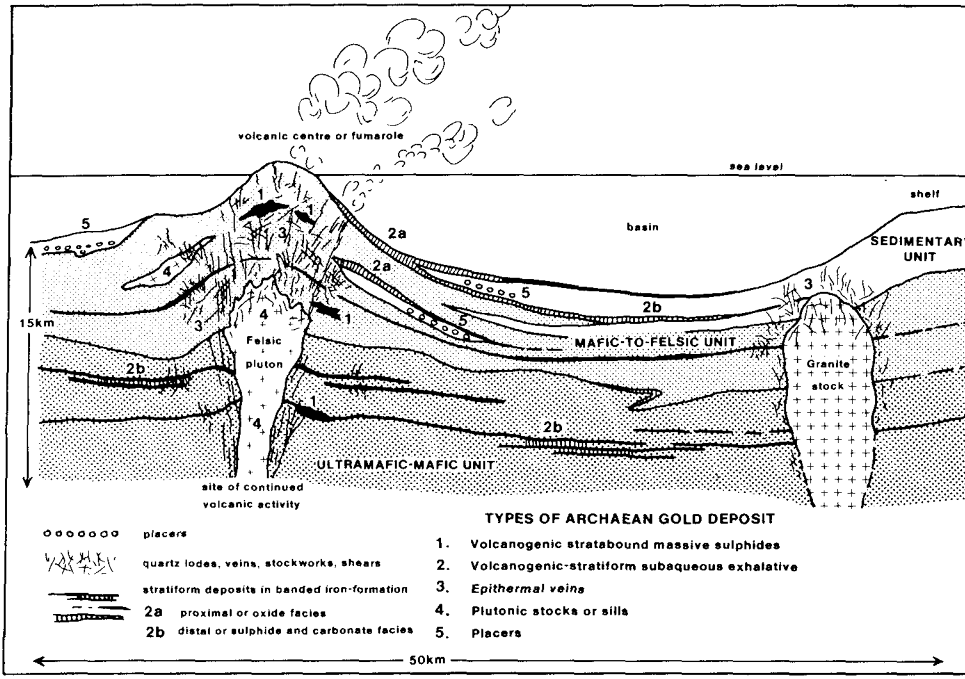

Fig. 27-3. Schematic diagram of an Archaean volcano-sedimentary complex showing the possible relation of gold and sulphide mineralization to various parts of the volcanogenic model (Anhaeusser, 1976a, after Goodwin and Ridler, 1970, and Hutchinson et al., 1971); reproduced by permission of Minerals, Science and Engineering).

mineralization and volcanogenic-sedimentary sulphide mineralization in Archaean greenstone belts in Canada, increasing evidence indicates many Archaean deposits are of similar origin and tectonic environment. Examples have been described from Vauze, Noranda (Spence, 1975) and Mattabi (Franklin et al., 1975) and recently at Agnico-Eagle in Quebec (Barnett and Hutchinson, 1978) where gold is present in massive pyritic ore overlying a distinctive welded rhyolite tuff. The mineralized zone of intercalated cherts, tuffs and graphitic pyrite-nodular schists with interdigitating welded rhyolite breccias grades into a chloritic alteration zone comparable with the feeder stockwork zone in Kuroko deposits (Horikoshi, 1969) and in the Iberian pyrite deposits (Strauss et al., 1977).

Volcanic exhalative iron formation (gold and carbonate-gold deposits)

Iron formations with or without economic gold mineralization are well-known from the southern African greenstone belts and from the Archaean basins of the Canadian Shield where they are sometimes referred to as Algoma-type deposits and account for 25% of Canada's iron reserves. In both regions the deposits are associated with volcanic rocks, mostly with the upper pyroclastic parts of the "mafic to felsic" association typical of the higher stratigraphic levels of the greenstone succession, and have undergone similar deformation and metamorphism.

Iron formations have been described in most detail from the Canadian Shield (Goodwin, 1973) where they occur in a number of basins now preserved as isoclinally folded remnants in synclinal ores. Goodwin (1973) has described how the oxide ores of haematite and magnetite occur preferentially around the inferred original margins of the basins, where conglomerates are abundant, while relatively scarce carbonate deposits of siderite occur between the oxides and the more widespread deep-water sulphide deposits, with pyrite and pyrrhotite associated with graphitic shales in the more central parts of the basin.

Rather similar deposits have been described from the Archaean of Zimbabwe where they are most abundant in the Sebakwian Group, the lowest stratigraphic unit of the greenstone belts. Here gold is concentrated in the sulphide facies of the jaspilitic iron formation, mostly with arsenopyrite which, together with pyrite and pyrrhotite, occurs in thin layers (Fripp, 1976).

The close association of the deposits with volcanic rocks has suggested to most authors a submarine volcanic exhalative origin involving hot brines venting from fumaroles in the sea floor. Sato (1976) has suggested that the brines were initially less dense than sea water, increased in density to that of sea water, and then decreased as mixing with sea water continued. Ridler (1976) has argued that gold-bearing carbonates in the Abitibi Belt are also submarine "exhalate", chemical sedimentary deposits.

Although many iron formations occur in association with mafic and particularly felsic volcanic rocks, broadly similar, predominantly sulphidic iron deposits also occur in a predominantly sedimentary succession. An example is provided by the deposits of the Fig Tree Group in South Africa (Anhaeusser, 1976b). The role of volcanism in genesis of these deposits is uncertain (Anhaeusser and Button, 1974) but the presence of quartzites and limestones in the succession suggests a shallow-water environment, perhaps analogous to the continental margins inferred for the Proterozoic deposits.

Goodwin (1973) argued that the silicic volcanic rocks associated with many of the iron formations in the Canadian Shield were volcanic arcs formed above subducting lithosphere. Whatever their origin, these rocks with the associated iron formations clearly accumulated near the boundaries between microcontinents and greenstone basins.

Gold solubility studies and modelling attempts (Fripp, 1976) relate to the possibility that ancient volcanic rocks in the greenstone belts contained unusually high levels of gold (Anhaeusser et al., 1975) but this has not been verified chemically. However, mafic to ultramafic komatiites do show a fairly uniform secondary hydration (7.5 to 10.3% H_2O), indicated chemically and by the presence of actinolite and antigorite (Anhaeusser et al., 1975). This may have been accompanied by leaching-out of gold (Weissberg, 1970; Fyfe and Henley, 1973). Evidence for association of gold with ultramafic rocks is found in Zimbabwe (MacGregor, 1951) and in the Timmins area of Canada (Pyke, 1975). Mineralogical support for the concept is provided by identifications by R. Saager (pers. commun., 1979) of gold in olivine crystals in the ores of the Louis Moore mine in the Sutherland greenstone belt, South Africa.

Porphyry copper deposits

Porphyry copper deposits were once believed to be restricted to the Cenozoic and Mesozoic, but have now been described from Palaeozoic in volcanic plutonic belts located on continental or microcontinental margins in the USSR (Laznicka, 1976), and Upper Proterozoic examples have recently been reported (see section "Proterozoic mineralization"). The scarcity of ancient porphyry deposits is most likely a result of erosion accompanying continental collision following their emplacement in a magmatic arc. However, it is still controversial whether Precambrian calc-alkaline island-arc-type magmas and associated porphyry coppers are subduction-related in a similar manner to Phanerozoic porphyry coppers (Sillitoe, 1975) as we favour, or are produced by sag-subduction or sagduction, involving crustal sag and resultant melting and interaction of accumulated volcanic material, etc. with mantle material (Gorman et al., 1978; Goodwin, 1978).

The disseminated Cu-Mo-Au deposit at the Pamour Mine in the Timmins area of NE Ontario is a possible example of an Archaean porphyry deposit with approximately 10 million tonnes of ore grading about 0.8% Cu, 0.05% Mo and 0.85 g per tonne Au (Davies and Luhta, 1978). Similar lenticular bodies occurring at roughly the same stratigraphical horizon within the thick sequence of early Archaean (Keewatin) mafic lavas throughout the area had previously been interpreted as early Archaean porphyry intrusions (Hurst, 1936; and Griffis, 1962). At the Pamour Mine similarities between the disseminated copper deposit and Mesozoic—Cenozoic porphyry copper deposits include feldspathic and sericitic alteration, alteration zoning, extensive alkali metasomatism, the sulphide assemblage chalcopyrite-bornite-molybdenite, disseminated sulphides, the presence of anhydrite and carbonates and a pyritic zone and gold-bearing quartz veins in adjacent altered mafic volcanic rocks. The main mineralogical difference is the presence of albite metacrysts (crystallized phenocrysts?) instead of the normal K-feldspar. However, albite has been noted in some porphyry deposits in Alaska and the Yukon (Hollister et al., 1975). Some data are perhaps capable of alternative interpretation, indicating that the deposit may have evolved in a different manner to that of most porphyry copper deposits. The porphyroidal albite crystals in the felsic schist, for example, are believed to be of metasomatic origin, accompanied by mineralization which occurred either during or after the final stages of deformation. However, Pyke and Middleton (1971) have suggested that the copper minerals were introduced into the host rocks prior to deformation, and that reconcentration occurred after folding.

Pre- to early syn-collision ore bodies of mafic-ultramafic association

Nickel-copper deposits of the Limpopo Belt

In Botswana nickel-copper deposits in the central zone of the Limpopo Belt occur within tightly folded Archaean supracrustals, correlated with greenstone belts to the north (Coward et al., 1976). Deposits at Pikwe and Selebi form major ore bodies, while those at Lentswe and Dikoloti are sub-economic (Marsh, 1978).

Mineralization is associated with metamorphosed, isoclinally folded ultrabasic sheets within an amphibolite facies metamorphic sequence and occurs specifically in metatroctolites at Pikwe and Selebi and in serpentinized metaperidotite and amphibolitized metapyroxenite at Lentswe (Marsh, 1978). The main sulphide mineral is pyrrhotite with minor pentlandite and chalcopyrite.

The Pikwe and Selebi deposits have been described by Wakefield (1976). The ore can be divided into three zones comprising massive sulphide in gneiss, hanging-wall sulphides above amphibolite and mineralized amphibolite. The presence of sulphide blebs in a small ultrabasic pod within amphibolite indicates that the ore was emplaced as an immiscible sulphide magma. The ore minerals, considered to be typical of those found in most Ni-Cu sulphide deposits, consist of pyrrhotite (84%), pentlandite (8%) and chalcopyrite (8%) with minor pyrite and magnetite. It is probable that the shape of the host amphibolite, forming a cigar-like body, is at least partly an original igneous feature, although modified by deformation.

There is little evidence for the tectonic setting during emplacement of the ultrabasic magmas and associated syngenetic sulphides, but they clearly predate the earliest recognizable deformation and metamorphism.

Chromite deposits in ultramafic intrusive complexes

Although most chromite in southern Africa is produced from the Proterozoic Bushveld intrusives, significant deposits also occur in Archaean layered complexes.

In the Selukwe greenstone belt of Zimbabwe of c. 3420 Ma age (Moorbath et al., 1976), chromite of general Alpine-type (Cotterill, 1979) occurs within an inverted sequence of banded ultrabasic rocks of the Sebakwian Group and is structurally underlain by conglomerates and basic volcanics. The Selukwe Belt is interpreted as an allochthonous thrust sheet (Coward et al., 1976). Smaller chromite deposits are scattered in the greenstone belts of southern Zimbabwe and the northern margin of the Limpopo Belt.

In the Red Lodge ultramafic complexes of Montana, U.S.A., chromitite-bearing ultramafic complexes, occurring amongst quartzofeldspathic gneisses, migmatites and amphibolites, have suffered intense deformation and metamorphism approaching granulite facies. Petrochemical considerations indicate that the chromitites, possibly from part of a satellite of the Stillwater Complex, were originally continental stratiform masses (Skinner et al., 1978).

The tectonic setting during intrusion of the Zimbabwe stratiform bodies remains largely unknown although considered to be rift- or fissure-related by Cotterill (1979). However, in western Ireland similar layered basic-ultrabasic bodies, subjected to upper amphibolite facies metamorphism and with associated migmatites, were emplaced during metamorphism of the latest Cambrian Grampian orogeny, recently re-interpreted as a result of continent—arc collision (Mitchell, 1978b). Similar ultrabasic bodies are not known from Cenozoic collision belts, but emplacement at a deep structural level of layered bodies of this type possibly may be typical of continental collision events of any age.

Western Australian nickel, copper

Naldrett and Turner (1977) interpreted the tectonic setting of the Ni-Cu deposits associated with the Upper Greenstone succession at Kalgoorlie in the Yilgarn Block, Western Australia, in terms of Red Sea-type graben structures which developed after the formation and folding of the Lower Greenstones: these were filled with mafic and ultramafic komatitic lavas, acid volcanics and sediments, derived from the volcanics and the adjacent rift walls (Fig. 27-4). Subsequent reactivation of the grabens, following granitic intrusions, resulted in vertical displacement along major faults in the area. It was suggested that convective plumes in the Archaean mantle produced rapid diapiric uprise of deep mantle material, causing extreme partial melting, and magnesian melts were intruded over existing crust within grabens. This model may be supported by the fact that the most magnesian liquids (up to 20% MgO) known in Phanerozoic oceans occur along the West Greenland and Canadian shores of Baffin Bay (Clarke, 1970) where graben formation in continental crust initiated Tertiary drift. However, high-magnesian lava feeder dykes in the Mesozoic Tortuga ophiolite complex, Chile, are also exceptionally primitive with 11—17% MgO and may represent parental magmas of this more evolved marginal-basin complex (Elthon, 1978). The graben structures or limited oceanic basins of the Kalgoorlie area, speculatively, could equally well be marginal basins related to subduction and the linear belts of Upper Greenstone volcanics may have resulted by stages of migration of subduction zones.

Groves (1979) has subdivided the Archaean Ni-Fe-Cu sulphide deposits in Western Australia into: (1) deposits related closely to komatiitic ultramafic volcanic flow sequences; (2) deposits occurring within intrusive

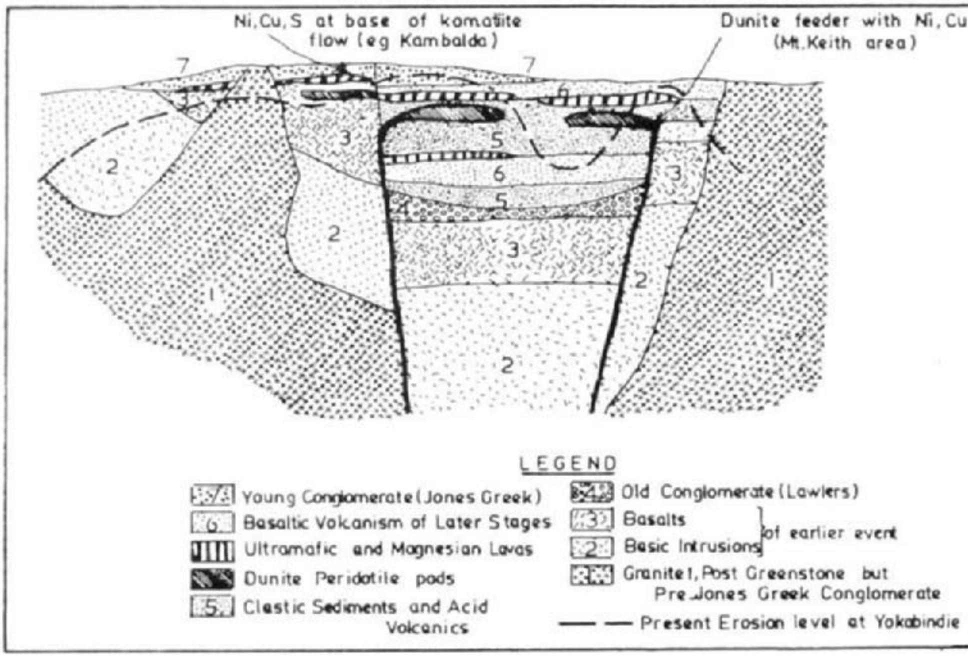

Fig. 27-4. Diagrammatic section illustrating the graben model for the Lawyers—Mount Keith area of western Australia showing types of Ni-Cu ore mineralization after Naldrett and Turner (1977); reproduced by permission of Precambrian Research.

komatiitic dunite pods; (3) stratiform deposits associated with layered intrusive bodies of tholeiitic or komatiitic affinity; and (4) deposits associated with sulphidic metasedimentary interflow units in volcanic sequences. Groups (1) and (2) are economically the most important and represent an important new class of komatiite-associated mineralization (Naldrett and Arndt, 1976; Binns et al., 1976). The main deposits occur within the Eastern Goldfields Province of the Yilgarn Block and Groves (1979) comments that the restriction of economic Archaean Ni-Cu deposits, particularly those of komatiitic affinity, to the granitoid-greenstone terrains of relatively young age appears to be a world-wide feature. Examples are the Canadian and Rhodesian deposits (c. 2800 ± 100 Ma) whereas the older greenstones of the Rhodesian Craton, Barberton and Greenland lack significant Ni-Cu deposits.

The volcanic-associated Ni-Cu deposits (Fig. 27-5) form 2—3 million tonne deposits of about 3% Ni and 0.3% Cu or clusters of deposits, e.g. Kambalda area, totalling over 20 million tonnes (Naldrett and Turner, 1977). The associated sequences of komatiitic flows (c. 40% MgO) are at or near the base of the larger flows. Massive ores are mainly pyrrhotite-pentlandite-pyrite-chalcopyrite-magnetite/chromite (Groves et al., 1977),

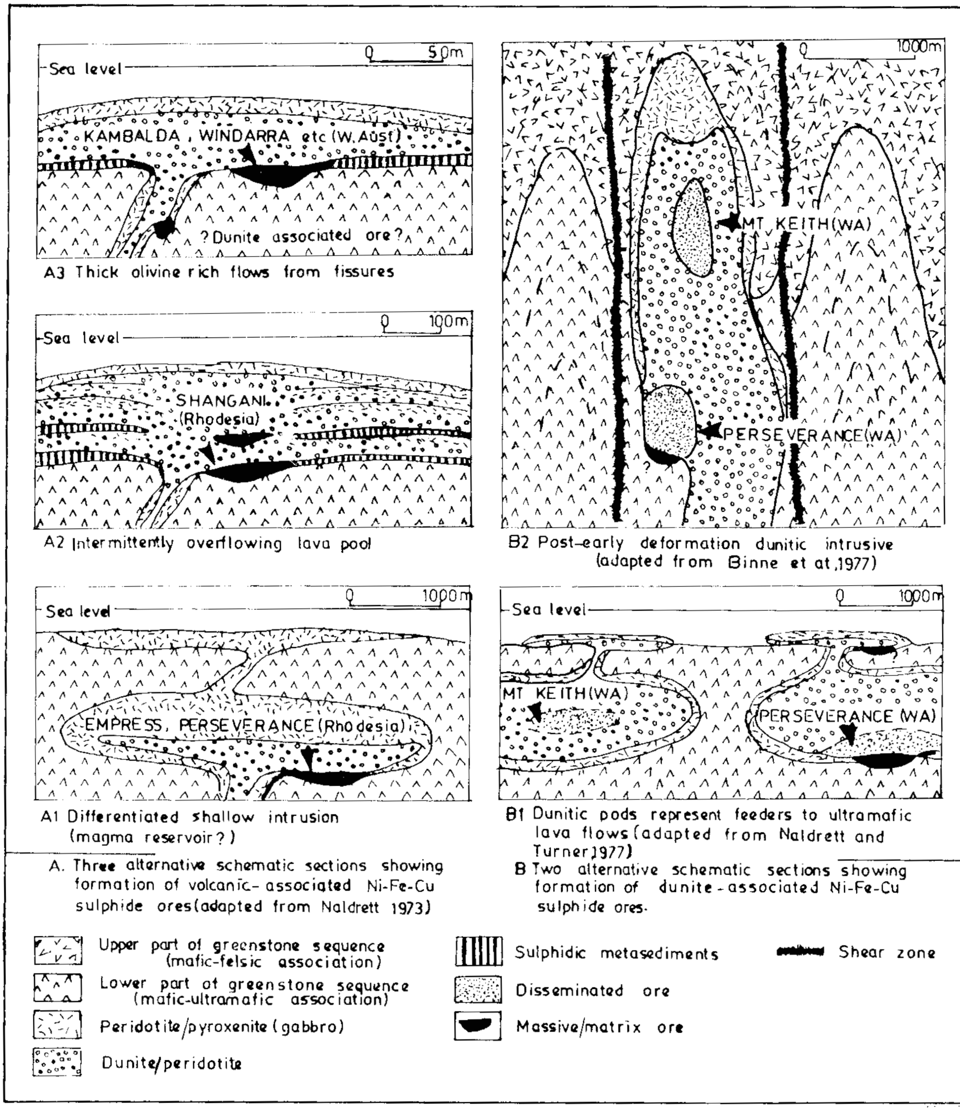

Fig. 27-5. Schematic diagram after Lambert and Groves (1981) of various origins possible for mineralized ultramafic flows and pods (Groves, 1979); reproduced by permission of Elsevier, Amsterdam.

generally overlain by matrix ores and more disseminated ores. Interpillow sulphides are locally present in underlying metabasalts. There are significantly high tenors of precious metal abundance patterns (Naldrett et al., 1979). Naldrett and Turner (1977) suggest that the Archaean mantle had sulphide zonation with the greatest enrichment at about 200 km in the source region of komatiite magmas.

The dunite-associated Ni-Cu deposits (Fig. 27-5) comprise large low-grade zones of disseminated sulphide in dunite pods, underlying and feeding the ultramafic flows with the volcanic associated Ni-Cu mineralization. Later reactivated fault zones possibly became feeder channels for sulphide-bearing ultramafic lenses (see Fig. 27-5). In the Mount Keith area there are 250 million tonnes with 0.6% Ni and 0.05% Cu, while the Perseverance remobilized ore body contains about 45 million tonnes with 2.0% Ni in dunite. The mineralization in unaltered dunite cores consists of disseminated pentlandite and chromite and locally there are massive ores on one margin, e.g. Agnew. Regional metamorphism has strongly influenced the final nature of these sulphide ores.

Ni-Cu deposits in stratiform intrusions are less important economically but are significantly more Cu-rich than the other deposits. Limited ore occurrences in intimate association with interflow sulphidic metasediments (Groves, 1979), e.g. Windara breccia ores, may be volcanic exhalative in origin or ore elements were redistributed due to metamorphism (Seccombe et al., 1977).

Collision-related ore bodies

Murchison Range antimony

The major antimony mines of the Murchison Range lie in the Kaapvaal craton north of the Barberton greenstone belt (Viljoen et al., 1978). The Range extends for 140 km, disappearing beneath younger rocks of the Transvaal Supergroup in the west and the Karoo in the east.

The Range consists in the south of magnesian metabasalts and talcose serpentinites and centrally of basic and acid volcanic rocks and metasediments, comparable, respectively, to the Lower Ultramafic Unit and Mafic to Felsic Unit of the Barberton Mountain Land. A layered mafic intrusion forms the northern part of the Range.

Antimony deposits, largely associated with carbonate rocks, occur along a narrow "Antimony Line" within the central part of the central basic and acid volcanic zone. The carbonates are commonly bordered by talc schists which pass transitionally into chlorite schists; they are often siliceous, resistant to weathering, and green or fuchsitic due to chlorite. Sb occurs in silicic pods or bands within the carbonate, commonly in association with minor As and Hg.

The linear nature of the Range and presence of ultrabasic, volcanic and carbonate-quartz-chlorite rocks comparable to those found in many Phanerozoic collisional sutures, cause us to speculate that these rocks are an allochthonous remnant of an Archaean microcontinental collision belt. It is uncertain whether the mineralization is syngenetic and analogous to Lower

Palaeozoic Sb-W-Hg deposits in Austria, possibly generated above a palaeo-Benioff zone (Höll, 1977), or epigenetic and related to magmatic, metamorphic or meteoric hydrothermal solutions.

Although no major deposits of antimony are known from Phanerozoic collision belts, stibnite mineralization in mid-Tertiary flysch within the India—Asia collision zone in Pakistan occurs on a transform fault (Sillitoe, 1978) and could have been derived from ultrabasic carbonate rocks at depth.

Pegmatites

Within Archaean terrains there are very few anatectic granites and leucogranites with high initial $^{87}Sr/^{86}Sr$ ratios comparable to those of the Himalayas and older orogens. However, pegmatites with important metal and gemstone deposits are known from a number of greenstone-granite terrains and constitute minor sources of tin, tungsten, beryl and tentalum, e.g. in Zimbabwe. Most are potassic pegmatites differing in composition from the predominantly tonalitic gneisses and appear to be syn- to late-tectonic in origin. In Archaean granulites in Kerala, southern India, unzoned chrysoberyl-biotite-perthite pegmatites carrying beryl and columbite are related to major transcurrent fractures probably developed during a late orogenic stage.

PROTEROZOIC MINERALIZATION

Ore bodies of intracratonic basins

The early Proterozoic contrasts with the Archaean, latest Proterozoic and Phanerozoic in its variety of major deposits, stratabound within predominantly sedimentary continental successions, and its scarcity of well-preserved subduction-related sulphide deposits. Some Proterozoic deposits can reasonably be related to intracontinental rift environments but the tectonic setting during deposition of others is uncertain and in some cases could be rift, back-arc, or post-collision foreland or hinterland basins. The significance of back-arc and post-collision basins was recently emphasized in a discussion on tectonic settings for post-Archaean alluvial gold deposits (Henley and Adams, 1979). Here we consider intracontinental deposits of uncertain tectonic setting as intracratonic basin deposits.

Quartz-pebble conglomerate uranium deposits

Uranium deposits in quartz-pebble conglomerates of early Proterozoic (2000 Ma) age are known from Canada (e.g. Elliott Lake), Brazil (Jacobina) and Western Australia. The uranium, commonly associated with pyrite, occurs above an angular unconformity at the base of a thick succession of shallow marine or continental sedimentary rocks. In the Elliott Lake deposits of Canada the uranium is considered to be largely detrital, derived from pegmatite and gneiss (Robertson, 1974), and transported and deposited as uraninite in an oxygen-free atmosphere.

Stratabound "sedimentary" base-metal ores

The stratabound copper-cobalt deposits of Zambia and Zaire, comprising the Central African Copper Belt, yield more than 20% of the western world's copper (Bowen and Gunatilaka, 1977). They occur above a regional unconformity at the base of a thick succession of Upper Proterozoic, regionally metamorphosed, mostly shallow-marine sedimentary rocks. The sedimentary facies are predominantly detrital in Zambia, but largely carbonates in Zaire (Brown, 1978). Mineralization, originally considered to be epigenetic and related to magmatic hydrothermal solutions, was subsequently widely interpreted as syn-sedimentary in origin (e.g. Garlick, 1961). More recently a syn-diagenetic origin has become increasingly accepted, and in the Zambian part of the Belt it has been suggested that pyrite was replaced by copper during upward movement of metal-chloride-bearing brines into structural traps where hydrocarbons had accumulated (Annels, 1979).

Evidence that the host-rock succession accumulated in an incipient rift zone which failed to develop into an ocean basin (Burke and Dewey, 1973) is based largely on similarities with the extensive stratabound copper deposits of the Permian Kupferschiefer in northern Europe. These similarities include the concentration of sulphides in dolomitic carbonaceous mudstones and sandstones above an angular unconformity, the presence of evaporites either stratigraphically above the ore-bearing beds or as matrix to sandstones, and evidence that the host rocks accumulated during a marine transgression. The Kupferschiefer accumulated during incipient rift development in the North Sea area, related either to a system of mantle hot-spots or to the immediately preceding end-Carboniferous continental collision of the Variscan orogeny. Other Phanerozoic copper deposits showing similarities to the African Copperbelt are the mid-Cretaceous copper deposits within rift-facies successions of the Atlantic margin of Africa.

Some authors consider the African Copperbelt deposits to be part of a major global late Proterozoic episode of copper mineralization which include the sub-economic deposits of the Adelaide Geosyncline in Australia and possibly the White Pine deposits of Michigan. However, evidence that rift facies are favourable environments for stratabound copper mineralization (e.g. Sawkins, 1976b; Raybould, 1978), together with lack of close age-control on the Proterozoic deposits, suggests that this type of mineralization may have developed in any incipient rift zone where palaeolatitudes were favourable for evaporite formation. Relative scarcity of such deposits may indicate that accumulation of sedimentary facies favourable for diagenetic mineralization requires depositional events, peculiar only to some rift zones, for example the sudden flooding of the basin during the marine transgression.

Other Proterozoic stratabound base-metal sulphides interpreted as having accumulated in a failed rift include the mid-Proterozoic Mt. Isa Ag, Pb, Zn deposits (Glikson et al., 1974), the McArthur River deposits (Raybould, 1978), and the Late Proterozoic stratabound copper mineralization of the Adelaide Geosyncline (Rowlands et al., 1978). All these deposits are more or less stratiform within mostly thick shallow-marine to non-marine successions, with features suggesting early diagenetic mineralization.

In each case the evidence for the presence of a syn-sedimentary and hence syn-mineralization rift is debatable. At McArthur River there is only limited evidence for the position of the inferred rift boundaries, in the Adelaide Geosyncline the metamorphism is more typical of a collisional orogeny, and in the case of Mt. Isa the succession lies near the probable Proterozoic eastern margin of the continent. However, in the absence of satisfactory alternative interpretations, formation of the mineralization in rift zones remains at least a possibility.

Stratiform "sedimentary" zinc, lead, barite ores

Within the Late Proterozoic Namaqua Mobile Belt in Cape Province, S. Africa, of amphibolite to granulite facies metamorphic grade, zoned ore bodies in the Gams Iron Formation have also been compared with the Kupferschiefer mineral zoning (Rozendaal, 1978). The ore deposits, from west to east with ore grades in parentheses are: Black Mountain (0.8% Cu, 2.9% Pb, 0.6% Zn); Broken Hill (0.36% Cu, 3.0% Pb, 2.2% Zn), Big Syn (1.23% Pb, 2.88% Zn) and the Gamsberg 95 million tonne deposit (0.5% Pb, 7.0% Zn).

At Gamsberg the upper mineralized zone is a banded quartz garnet-grunerite rock with 10—40% opaque minerals comprising iron, zinc and lead sulphides and accessory minerals. The lower horizon consists of mineralized cordierite rock overlying fine-grained quartzite and banded mineralized sillimanite schist with iron, zinc and copper sulphides. Barite locally forms intercalated lenticular bodies or layers in an upper horizon.

Despite the high grade of metamorphism in the area, the distinct zoning across the first-mentioned three ore deposits at Aggeneys through to Gamsberg is believed (Rozendaal, 1978) to illustrate lateral zoning of Kupferschiefer-type (Wedepohl, 1971) with Cu near the rim of the basin, Pb intermediate and Zn near the centre. The vertical zoning also shows marked sedimentary facies changes. Within the Gams Iron Formation as a whole an oxide facies, a carbonate facies and sulphide facies can be recognized in a transgressive sedimentary basin. Electron-probe analyses (Stumpfl, 1973) showing that the pyrite lacks "magmatic" Ni or Cr and the magnetite is a non-magmatic type with no Ti, V or Cr, also support a sedimentary origin for the deposit.

Although possible Kupferschiefer-type deposits such as the Copperbelt and the Gamsberg sedimentary ores are apparently not related to plate tectonics in any obvious way, basins of this type have formed well within continents during or following plate convergence. Osmaston (1977) has suggested on the basis of a novel re-interpretation of the geological and geophysical frame work of plate tectonics that plates are thicker and stiffer than popularly believed, so that plate stiffness makes epeirogenic processes at plate edges felt at great distances from them, causing flexures, intra-plate rifting, block-and-basin structure and basin-subsidence.

Sandstone-type uranium ores

Uranium in sedimentary rocks occurs within the thick succession of mid-Proterozoic rocks of the "Athapuscow Aulacogen" (Hoffman, 1973), east of the Great Slave Lake in Canada. The uranium is mostly in the form of interstitial uraninite, replacing sulphides, and occurs in non-marine sandstones and conglomerates at several stratigraphic horizons in the 12-km thick succession of continental and marine sediments and volcanic rocks. Morton (1974) has suggested that the uranium was precipitated by metamorphic hydrothermal fluids expelled either from the deepest levels of the succession or from the underlying Archaean basement. Hoffman (1973) has argued that the succession accumulated in a failed rift or aulacogen which formed one arm of an RRR triple junction, the other two arms having opened to form the Coronation geosyncline.

Ore bodies in intrusions within megafracture systems

The formation of megafracture systems in the early Proterozoic is regarded (Windley, 1977) as an initial effort to fragment the first-formed extensive continental masses. Magmas intruded into these fractures gave rise to transcontinental dyke systems, major basic-ultrabasic layered complexes, carbonatites and alkali intrusions.

Cr-Ni-Pt-Cu mineralization

In southern Africa examples of three types of intrusion related to a megafracture system are first the Great Dyke of Zimbabwe with a minimum age of 2530 ± 30 Ma (Allsopp, 1965), secondly the dolerites of Mashonaland, Umkondo and Waterberg of 1750—1950 Ma age (Vail, 1977) and thirdly the layered basic and ultrabasic Bushveld Complex, dated at 1950 ± 150 Ma (Nicolaysen et al., 1958). The larger intrusions such as the 480 km long Great Dyke and the Bushveld Complex are characterized by Cr-Ni-Pt-Cu mineralization of both disseminated and cumulate type.

The Great Dyke of Zimbabwe which averages nearly 6 km in width includes large lopolithic subcomplexes with layering dipping inwards at a shallower angle than the main dyke contacts (Worst, 1960). Bichan (1970) has described each complex as consisting of cyclic sequences of ultrabasic rocks overlain by a thick cap of gabbro. The ultrabasic cycles have basal chromite seams and sequences upward of perodotite, pyroxenite, anorthositic gabbro and norite, and finally of quartz gabbro. Bichan considered that the dyke formed in successive magmatic surges over the thermal updraft of a mantle convective cell, and a waning of the heat flow pattern resulted in slumping of the dyke into its graben. Recently, Bhattacharji (1978) has made stress analyses and model experiments showing that mantle upwelling may produce funnel-shaped structures resembling modern rift valleys in various stages of development, and has suggested that some Precambrian ultramafic layered intrusions may be zones of incomplete rifting in the shield over ancient hot-spots and are roots for modern or ancient rift zones.

The Bushveld Complex, one of the world's largest igneous bodies and repositories of magmatic ore deposits, occupies an elliptical area of about 66000 km^2 with a vertical thickness of up to 8 km. The complex has a central part underlain by acidic rocks and two lobate marginal belts composed of rocks ranging from dunite to norite, anorthosite and ferrodiorite. Windley (1977) comments that although much of West Africa was undergoing orogenic activity at 1850 ± 250 Ma (Clifford, 1972), the intrusion of the Bushveld Complex is indicative of widespread stable conditions in southern Africa at about this time.

Magmatic ore deposits in the ultramafic parts of the Bushveld Complex include vanadiferous and titaniferous magnetite layers and pipe-like bodies in the upper layered sequence (anorthosite), platinum and nickel in the Merensky Reef and in bronzitite pipes, and chromite in chromitite layers in the lower layered sequence of pyroxenites, norites and anorthosites.

An alignment of igneous centres along the trace of a great circle passing through the Bushveld Complex and the Great Dyke was first recognized by Cousins (1959). Additional complexes have been added and the line now apparently extends (Fig. 27-6) for a distance of 3,800 km between the Orange Free State and Ethiopia (Vail, 1977). A further 2000 km extension is possible (see inset to Fig. 27-6), if large basic intrusions in Sudan and three Egyptian carbonatites, invaded by batholithic granites of 800—1000 Ma ages (El Shazly et al., 1973) are included (Garson, in press). This mega-lineament is remarkably smooth without the zig-zag patterns of typical rift structures and may more readily be compared with major transcurrent or transform faults with which ultrabasic to alkaline complexes may be associated (Mitchell and Garson, 1976). Vail (1977) has commented that despite apparent different ages of emplacement from about 2600 Ma ago, the alignment of centres has persisted, suggesting that a remarkedly permanent and fundamental phenomenon has occurred, relatively unaffected by subsequent tectonic efforts. One theory that may account for the simultaneous origin of bodies at the southern end of the line is that they were the result of meteorite impact (Dietz, 1961). The Vredefort Dome and the Bushveld Complex may have structures compatible with this idea (Hargraves, 1970; Hamilton, 1970), suggesting that the magmatic events in the south could have been triggered off by impact which produced a deep-seated

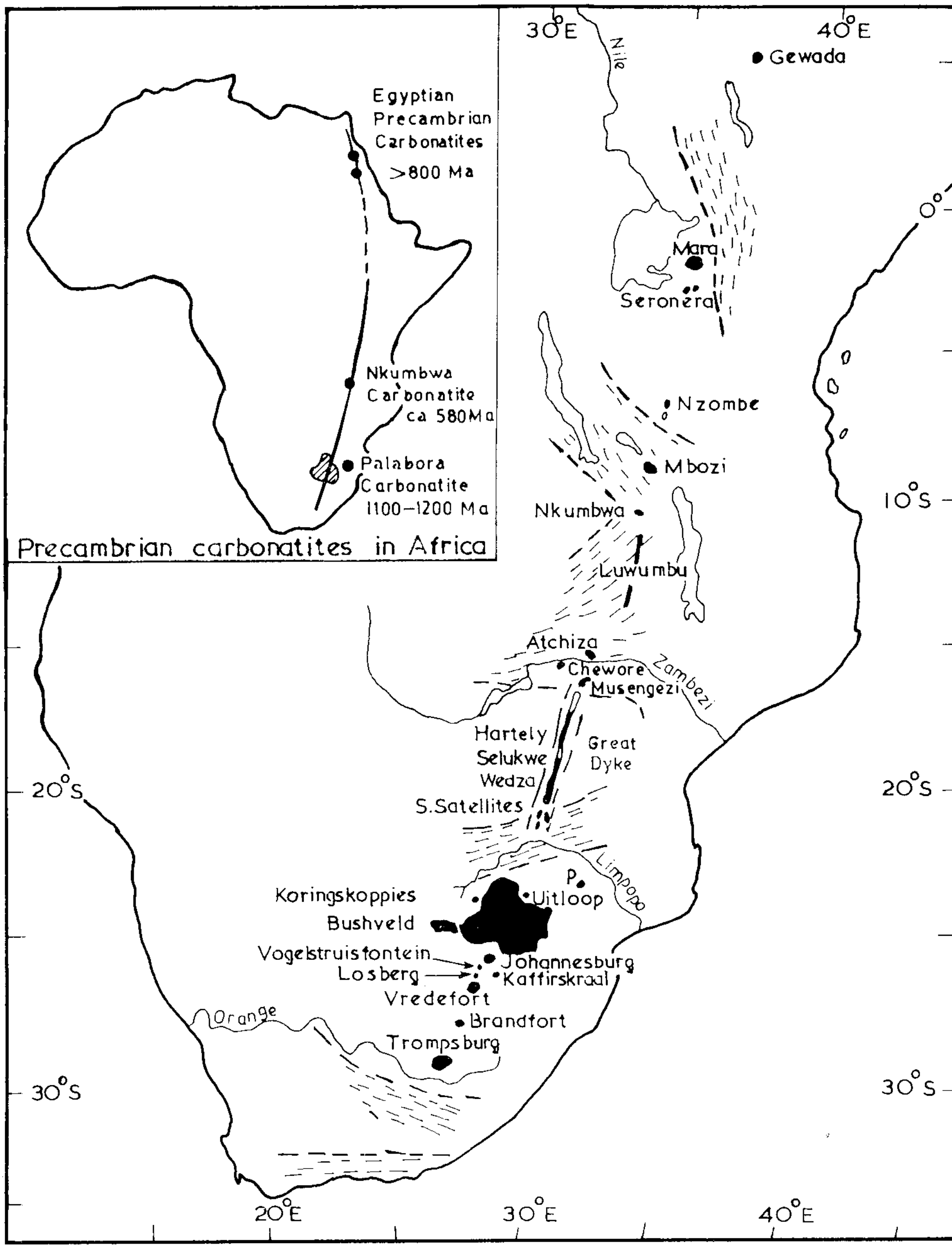

Fig. 27-6. Alignment of Proterozoic basic/ultramafic complexes along a megafracture in Africa (Vail, 1978); reproduced by permission of the Geological Society of South Africa. Inset shows position of Precambrian carbonatites.

crack in a relatively thick crust, accessible to later surges of magmatism. An impact origin has been suggested by Dietz (1964) for the comparable Sudbury Complex in Canada of slightly younger age (1720 Ma), but with similar Cu, Ni, Co, Au and Pt mineralization. This complex is aligned with nine younger carbonatite complexes in the Ottawa Graben, (Kumerapeli, 1976), and along the WNW extension of this direction (Erdosh, 1979) there are additional Proterozoic carbonatites.

Carbonatite mineralization (P, Nb, U, Cu, Ce, Zn)

No published accounts have been given of pre-Proterozoic carbonatites, unless supposed gold-bearing exhalative carbonatites of Archaean age at Timiskaming (Stricker, 1978) are confirmed. However A. Kröner has reported (pers. commun., 1980) the presence of a c. 3 Ga old carbonatite from the Nellore greenstone belt in Andra Pradesh, India. The Proterozoic carbonatites throughout the world cluster round certain chronological peaks of which few are composed exclusively of measurements from a single area. The main times of emplacement occur at intervals of about 250—350 Ma (2020 Ma; 1875—1695 Ma; 1540—1310 Ma; 1150—976 Ma; 700—565 Ma) which correlate with recognized Precambrian orogenies (Gittins et al., 1967), and are probably symptomatic of locally increased heat flow at these times.

In Africa carbonatites situated along or close to the mega-lineament described above (Fig. 27-6) include the barren Egyptian carbonatites associated with peridotite, and Nkumbwa in Zambia, dated at 680 ± 25 Ma (Snelling et al., 1964) with disseminated pyrochlore and monazite mineralization (Deans, 1966). Spitzkop, at the northeastern edge of the Bushveld Complex, and Palabora (see Fig. 27-6) are both dated at 1310 Ma (MacIntyre and Dawson, 1971). Palabora is a rich worked deposit of Cu (mainly chalcopyrite), U (uranothoriante), Zr (baddeleyite) and apatite mineralization (Foster, 1958). The copper mineralization is disseminated in the carbonatite while the apatite occurs abundantly in central areas and also in an outer zone of pyroxenite. Although Proterozoic carbonatites along this mega-lineament are scanty, it seems to have exercised a considerable control on later rifting and carbonatite locations.

As mentioned previously Precambrian carbonatites also occur on a tectonic line through the Sudbury Complex in Canada and a gravity/magnetic high, termed the Kapuskasing High, which trends NNE from the Seabrook carbonatite on this line, also has carbonatites ranging in age from 1012 Ma (Gittins et al., 1967). These are rich in phosphate, both as primary apatite in carbonatite and as secondary phosphorite (Erdosh, 1979); pyrochlore is also present in several of these carbonatites.

In Tamil Nadu, India, elongated intrusions of Precambrian carbonatite together with massifs of syenite and dunite and with quartz-barite veins, occur along the Bhavani system of en-echelon faults. Sevathur, the largest carbonatite, is rich in uranian pyrochlore, apatite and zircon (Udas and Krishnamurthy, 1970).

In summary, the Proterozoic carbonatites are similar in most features including composition and associated mineralization to Phanerozoic carbonatites. Even the remarkably elongated Siilinjärvi carbonatite in Finland (Puustinen, 1971) has an analogue in the Sangu Complex (Coetzee, 1963) along the NE rift of Lake Tanganyika, formed of three bodies totalling 26 km in length.

Intraplate tin deposits of possible hot-spot origin

Generation of tin-bearing granite complexes in Rondônia, Brazil (Kloosterman, 1967), has been related (Sawkins, 1976b) to a mantle hot-spot. About thirty ring-like granitic intrusions, ranging in size from 5 to 20 km across, form elongate possibly fault-controlled belts in high-grade metamorphic rocks in the Rondônia area (Priem et al., 1971). In the northern part of the area several granitic complexes intrude metasedimentary rocks of pre-Grenville age which infill a rift valley trending roughly E—W.

The intrusions are strongly differentiated with chemical compositions ranging from those of biotite granite to quartz monzonite and syenite. Their syenitic affinities and field relations were believed by Priem et al. (1971) to be indicative of anorogenic emplacement in a stable cratonic environment. Rb-Sr data provide a computed isochron age of 977 ± 20 Ma, with an initial $^{87}Sr/^{86}Sr$ ratio of 0.710 ± 0.008, indicating considerable assimilation of crustal rocks. The tin mineralization in greisen-like zones and topaz-quartz veins, cutting both complexes and country rocks, is characterized by a topaz/cassiterite association and locally wolframite and columbite-tantalite are present.

Sawkins (1976a) has suggested that the age of emplacement of the Rondônia granites coincided with a major widespread episode of hot-spot activity associated with the fragmentation of a postulated proto-Pangaea.

It is tempting to relate the Bushveld granite and associated mineralization to hot-spot activity which postdated the ultrabasic Bushveld Complex, particularly in view of its significant position on a megafracture (see Fig. 27-6) which has channelled igneous activity of various types. However, the granite-felsite-granophyre association of the Bushveld granite seems to have little in common with the alkaline character of other anorogenic intraplate granites (Sillitoe, 1974), and the distribution of the tin deposits is considered to have been determined by the intersection of the mega-fracture with a broad E—W belt of compression (Stear, 1977). The presence of tin in this granite, however, is important according to Watson (1973), since tin accumulation is related to slow-operating fractionation processes, and yet in this area these processes were operating 2000 Ma ago, presumably within a relatively thick continental crust.

Ore bodies of the ocean floor and rise

Ophiolitic deposits (Cr, Cu)

The Red Sea area of Pan-African ophiolites and associated volcanic arcs is one of the few areas where easily identifiable Precambrian ophiolitic mineralization can be seen. This area is the subject of a separate study (Gass, this volume, Chapter 15) and much of this mineralization has been described elsewhere (Garson and Shalaby, 1976; Al-Shanti and Mitchell, 1976; Bakor et al., 1976) and abstracts of a recent symposium on the development of this area are published (various authors, Precambrian Research, 1978, 6: A1—A40). It is therefore unnecessary to give full details here. Briefly, examples of ophiolitic mineralization comparable to that of type, if not in scale, in Phanerozoic times, include typical podiform chromite in serpentinized dunites in the Eastern Desert of Egypt and north Sudan and sulphide bodies associated with cherts in ophiolitic volcanics in Saudi Arabia. These deposits occur in arc-arc collision settings. Of particular interest are mafic/ultramafic intrusions with, in one case, cumulate Ni, Cu sulphides and in others cumulate ilmenite within transverse tectonic structures believed to be continental Precambrian analogues of oceanic transform faults (Garson and Krs, 1976).

At Bou Azzer in Morocco inferred ophiolites of similar Pan-African age, obducted on to the west African craton, contain chromite pods and asbestiform chrysotile in the upper pillowed lava layers (Leblanc, this volume, Chapter 17).

Shackleton (1977) has listed basic and ultrabasic rocks of possible ophiolitic character associated with the Poronge Group in the Ribeira Belt in Brazil, and he also mentions a belt of mafic to ultramafic rocks in central Brazil which stretches intermittently for a distance of 1500 km from the Itacaiune River to the vicinity of Chiba.

The Matchless amphibolite zone of altered ultrabasic rocks and tholeiites of supposed oceanic type (Finnemore, 1978), with associated iron, copper sulphides, in the Damara Belt of Namibia has been defined as a possible suture zone (Burke et al., 1977; Shackleton, 1977). This was disputed by Kröner (1977b) who considered that these rocks were stratabound basic volcanic intercalations in metasediments extruded in a subsiding intracontinental basin. Later, however, Kröner (1979), although again concluding that the Matchless belt had features incompatible with the Himalaya-type plate-collision model, admitted of the possibility that the Matchless rocks could have formed in an aulacogen in which deformation and metamorphism followed crustal detachment and diapiric rise of an upper mantle asthenolith. He considered that upper mantle rocks may have broken through at a late stage to form an incomplete ophiolitic suite. We believe that the ideas of Burke et al. (1977) are not completely irreconcilable with those of Kröner (1979), the main difference being partly of semantics and

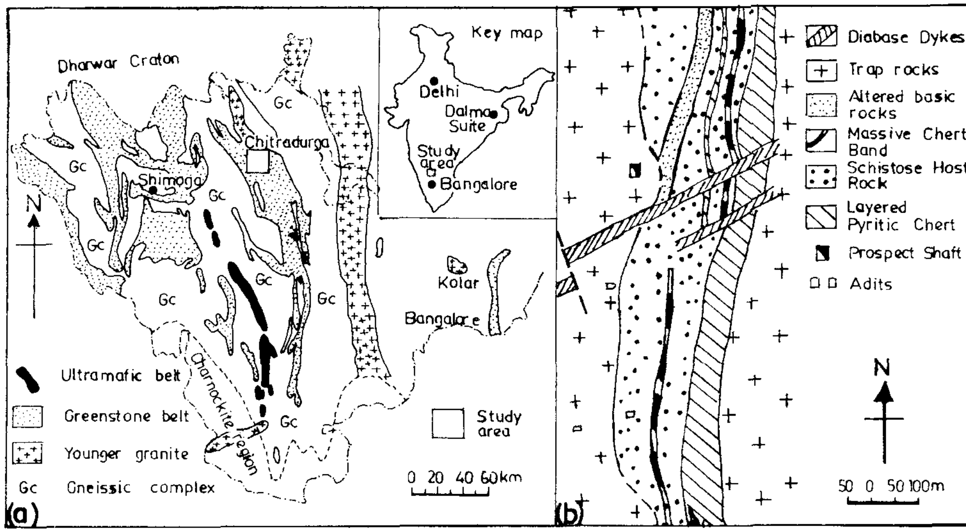

Fig. 27-7. a. Geological map of the Southern Indian Precambrian shield (Yellur and Nair, 1978; reproduced by permission of Precambrian Research).
b. Surface geological map of Beligudda Hill, Chitradurga, showing distribution of rocks after Vishwanathiah et al. (1969).

of scale, in that Kröner maintains that insufficient oceanic crust was generated in the very narrow ocean to allow more than incipient subduction during oceanic closure and orogenesis.

Data from the Chitradurga sulphide occurrence in southern India (Fig. 27-7a) fit well for ophiolite-type mineralization in this greenschist area of the Dharwar schist belt, believed to be of early Proterozoic age (c. 2500 Ma). A detailed study of the Chitradurga metabasalts shows that they were probably ocean-floor basalts erupted from an axial rift region (Yellur and Nair, 1978). The Chitradurga metabasalts occupy NNW-trending belts, some several hundred km long, in which lie the Chitradurga copper mine and several gold occurrences (Viswanathiah et al., 1969). A 350 km long narrow belt of serpentinized ultramafic rocks in the central Dharwar craton has been classified as of Alpine-type by Pichamuthu (1969); this is a possible Proterozoic suture zone.

The Chitradurga copper deposits occur within a mineralized zone some 35 km long and are worked at an enriched area at Belligudda Hill (Fig. 27-7b) within a 2500 m thick succession of party pillowed metabasalts, agglomerate and tuffs with interbedded pyritiferous chert and sericitic phyllites (Naqvi et al., 1977). Sulphide mineralization consisting of pyrite, pyrrhotite, chalcopyrite and sphalerite is predominant in chert bands but there is also local enrichment of pyrite and chalcopyrite in pillowed lavas. Virtually identical mineralization was seen by the writers in association with a less altered assemblage of pillowed lavas, tuffs, greywackes, minor argillites and red/green cherts in the Betts Cove ophiolitic complex of Newfoundland (Upadhyay et al., 1971). Studies of the Chitradurga ore mineralogy show that it strongly resembles the Cu-rich ophiolitic ores in Cyprus (Naqvi et el., 1977).

Subduction-related mineralization

Porphyry copper

Proterozoic calc-alkaline magmatic arcs now within continents are generally deeply eroded and porphyry ores, normally emplaced around and within the upper parts of stocks, consequently have largely been removed. Nevertheless there are several isolated examples which are closely comparable to Phanerozoic porphyry ores.

One of the best examples of late Proterozoic age, occurring in an early island-arc environment (Garson and Shalaby, 1976), is the Hammash-Umm Tundub porphyry in the Eastern Desert of Egypt (Ivanov et al., 1973) which was once mined for gold and copper. Although no economic mineralization was found during a UNDP investigation, drillholes penetrated pyrite-rich outer shells typical of those flanking zones of Phanerozoic porphyry copper mineralization. The ancient Hammash mines occur in Dokhan Volcanics of 570—650 Ma age comprising various types of calc-alkaline volcanic rocks with extensive areas of hydrothermal alteration including propylitization, kaolinization, silicification and pyrophyllitization (Ivanov et al., 1973).

Another area of porphyries of much greater age but with characteristic porphyry Cu-Mo mineralization has been described in Upper Volta and Niger in West Africa (Goossens, 1971). Two rhyodacite and albitophyre centres (Gaoua and Goren) are situated in Upper Volta and the third, Kourki, is a "granitoid" in southern Niger near the NE border of Upper Volta. The three centres all occur within the c. 2200 Ma old Birrimian system. Mineralization at the Upper Volta centres consists of veinlets of pyrite and chalcopyrite and disseminations of molybdenite associated with typical porphyry-type hydrothermal activity. At Kourki chalcopyrite and molybdenite occur in specks in quartz veins and disseminations in the altered rocks, and a shell of copper enrichment surrounds an enriched molybdenum core; hydrothermal activity is limited to albitization, epidotization and kaolinization. However, F. J. Sawkins (pers. commun., 1979) considers that this type of alteration is more typical of rift-related hydrothermal deposits. The reported ore tonnages are: at Gaoua — 35 million tonnes at 0.7% Cu; at Goren — 50 million tonnes at 0.4% Cu; and at Kourki — 400 million tonnes at 0.2% Cu. Grades are therefore probably sub-economic unless molybdenum values and associated gold values are sufficiently high.

The Birrimian system consists of phyllites, greywackes, slide conglomerates, tuffs and pillow lavas with intrusions of mafic rocks which are now largely epidiorite and serpentinites (Junner, 1940); these rocks pre-date the Eburnean orogeny of 1800—2100 Ma (Clifford, 1968). Volcanics include tholeiites and spilites and there are acid to calc-alkaline rocks including the intrusions with porphyry-type mineralization described above. Burke and Dewey (1971) consider, somewhat speculatively in our opinion, that the area between the Birrimian and Tarkwaian schist belts of West Africa include a number of microcontinents which collided, that continental crust was differentiated from the mantle after the formation of the schist belts, or that both processes operated. The presence of porphyries may suggest, however, that there were arc—arc collisions or arc—microcontinent collisions following subduction and convergence.

The Samba porphyry-type copper deposit, situated in the pre-Katangan basement of the Zambian Copperbelt (Wakefield, 1978), consists of pyrite, chalcopyrite and bornite in deformed granodioritic and quartz monzonitic porphyritic intrusions of pre-Katanga age. It is unclear if these rocks are intrusive into, or unconformably overlain by, Muva System sediments of approximately 1300 Ma age. At Samba, despite later metamorphism, alteration zones can be recognized including a central sericitic core. There are crackle breccia textures and vein and disseminated mineralization with an overall grade of 0.5% Cu. Wakefield (1978) suggests that the Katangan mineralization of the Copperbelt owes its location at least partly to the presence of cupriferous basement rocks including porphyries and that other hidden bodies may yet be discovered.

The age of the Samba porphyry is uncertain and scanty knowledge of the pre-Katangan basement does not provide clues as to the possible position of a suture. Burke and Dewey (1971) have suggested a convergence of the Mozambique belt in a similar manner to that of the Grenville Front. This might have produced subduction-related porphyries to the west but Kröner (1977b) considers there is no collision suture along the western margin of the Mozambique belt and suggests that the so-called cryptic sutures of Burke and Dewey (1971) and (Burke et al. 1977) are hallucinosutures, as observed by Shackleton (1976). In this connection it is tempting to suggest that the missing suture could be represented by the narrow elongated zone of ultramafic rocks in the Mozambique belt of southern and central Malawi, consisting of amphibolite, metagabbro, anorthosite and serpentinite with closely associated thin layers of quartzite and marble (Bloomfield and Garson, 1965). McWilliams (this volume, Chapter 26) considers the suture must lie east of the present African coast belt but this seems too distant to be related to production of porphyries in the Copperbelt basement.

In Canada, porphyry copper deposits are found in the 1800 Ma old porphyry intrusions in the Echo Bay Group andesites of northern Canada (Badham et al., 1972), and Windley (1977) queried whether these were related to an early Proterozoic subduction zone. In the northern Yukon Archer and Schmidt (1978) have described mineralized breccias of early Proterozoic age which have mineralogical similarities to breccia pipes associated with porphyry copper deposits.

Kuroko-type (Zn-Pb-Cu) stratiform sulphides

Stratiform massive sulphides, similar to the Japanese Kuroko ores, consisting of Zn, Pb and Cu ore in association with silicic volcanic rocks erupted in a shallow-marine environment, are known from the Upper Proterozoic arc—arc collision belt of the Red Sea area (Garson and Shalaby, 1976; Searle et al., 1978; Sabir, 1978). In contrast to the Archaean Abitibi-type stratiform sulphides the Proterozoic Kuroko ores carry Pb in small to large amounts.

In Egypt the metavolcanics, with which the Kuroko-type ores are associated, include thin andesites, rhyolites, dacites and pyroclastics overlying great thicknesses of pillowed basalts. Searle et al. (1978) have recognized five volcanic cycles comprising a basic or intermediate phase succeeded by an acidic phase. Volcanoclastic breccias, cherty tuffs and banded cherts exhibit sedimentary features such as graded bedding and slumping, indicating submarine volcanism and deposition. Sulphide ores, forming lens-shaped bodies at the contact of rhyolitic domes and felsive volcanic breccias (mill-rock?) comprise massive and disseminated sphalerite, chalcopyrite, galena and pyrite; stockwork ore is present in mine dumps. Locally there is enrichment in manganese, and silver grades range up to 200 g/tonne.

Similar stratabound sulphide deposits occur in Saudi Arabia (Sabir 1978; Moore, 1978). Three types were recognized by Sabir. The first comprises lens-shaped massive sulphide deposits at Jabal Sayid and Nugrah associated with siliceous, graphitic and/or calcareous beds, forming stratiform mineralization zones within felsic pyroclastic sequences. These deposits are mainly sphalerite and pyrite with variable amounts of chalcopyrite, galena, sulfosalts, oxides and tellurides. The second type includes pipe-like stockwork sulphide deposits in chloritized volcanoclastic rocks at Jabal Sayid and Jabal ash Shizam which contain mainly cupriferous mineralization and pyrite. At Jabal Sayid the cupriferous stockwork is a feeder to the sphaleritic massive sulphide area immediately above. A third type at Al Amar also resembles a stockwork; this is characterized by high Au and Ag contents with variable amounts of pyrite, chalcopyrite, sphalerite, galena and tellurides.

Similar Kuroko-type sulphide deposits in Eritrea were mined until recently. Barite which typically forms in Japanese Kuroko deposits is found at one locality associated with sphaleritic sulphide ore (Garson and Shalaby, 1976).

Magmatic uranium

The uranium mine at Rössing in Namibia, recently described by Berning et al. (1976) and Jacob (1978), is the largest and richest magmatic uranium deposit and is also the oldest significant deposit of this type. The deposit lies within the central part of the Damara belt, near the base of the thick Damara Supergroup of shallow-water metasediments which underwent regional metamorphism and isoclinal folding in the latest Proterozoic "Pan-African event". Uranium occurs within alaskitic pegmatitic granite which forms one of a number of late- to post-tectonic intrusions. The alaskitic rocks mostly form pods and dykes of fine to very coarse-grained or pegmatitic granite.

The Damara granites are considered to have formed by partial melting of underlying pre-Damara gneisses and deeply buried stratigraphic equivalents of the Damara Supergroup source rocks and the underlying basement.

Uranium, in the form of uraninite and secondary minerals, is present in concentrations of several hundred ppm in the alaskitic rocks at Rössing and occurs preferentially in alaskites underlying and in contact with carbonate beds of the Rössing Formation, suggesting that the marbles trapped the melt (Jacob, 1978).

The Damara belt is commonly interpreted as having evolved from an ensialic intracratonic sedimentary basin (e.g. Martin, 1978), but the presence of a continental collision zone is suspected (see ophiolitic deposits) south of the Rössing area. Regional metamorphism and deformation of the Damara Supergroup with emplacement of syn- to late-tectonic leucogranites can be interpreted in terms of crustal thickening of the subducting continental plate analogous to that of the Indian plate in the Himalayas. The mineralized Rössing alaskite can thus be considered as a late differentiate of a collision-related granite suite.

The Rössing deposit has been compared with anatectic pegmatitic and silicic granites within areas of high-grade regional metamorphism in Proterozoic rocks of Canada and late Proterozoic rocks of eastern U.S.A. (Rogers et al., 1978). Broadly similar uranium-bearing Devonian leucogranites of the western Massif Central of France (Leroy, 1978), and also the post-tectonic late Carboniferous tin- and tungsten-bearing granites of southwest England have been compared specifically with the collision-related Tertiary granites of the Higher Himalayas (Mitchell, 1974, 1978a). Apogranites in the Eastern Desert of Egypt (Bugrov et al., 1973), may likewise be late differentiates of a collision event. The Egyptian apogranites carry tin-tungsten mineralization largely in albitized and greisenized apical portions and stockworks. Accessory minerals are scheelite, columbite, tantalite, chalcopyrite, molybdenite, lithium micas, beryl and topaz.

CONCLUSIONS

(1) The broad similarities between modern plate-tectonic-related ore

deposits and Precambrian mineralizations and their settings indicate that uniformitarian concepts of plate tectonics are probably applicable well into the Archaean.

(2) A comparison between Phanerozoic and Precambrian mineralization and tectonic setting deduced from the geology leads to an acceptance that the concepts of Burke et al. (1976) were fundamentally correct. Mineralized Archaean greenstone-granodiorite belts were interpreted as island-arc and basin volcanic rocks, formed during subduction of oceanic crust which had developed by back-arc spreading and post-collision rifting.

(3) In the Archaean distribution of arcs and spreading centres and much of the associated mineralization were analagous to those in present-day Indonesia, Melanesia and the Caribbean, covering much of the earth's surface.

(4) Back-arc marginal basins and rifts were important environments for greenstone belt mineralization of Abitibi-type and related Au-rich exhalative iron and carbonate deposits, sulphide Ni, Cu ores associated with komatiites, stratiform Cr ores and Kuroko-type ores. Sparsely preserved porphyry mineralization (Cu, Mo) was probably related to subduction processes.

(5) In the late Archaean placer gold-uranium deposits formed in intra-cratonic basins, e.g. Witwatersrand; these basins were developed also in the Proterozoic with deposition of stratiform ores of Kupferschiefer type (Cu, Co) in the Copperbelt and at Gamsberg (Zn, Pb, Cu) in South Africa.

(6) Megafractures were important features of the early to middle Proterozoic and associated mineralization included Cr, Cu, Ni and Pt in large layered intrusions such as the Bushveld Complex, and in large dykes, e.g. Great Dyke of Zimbabwe. Megafractures also controlled the distribution of many Proterozoic carbonatites with Nb, P, U, Ce, Zr mineralization in Africa, Canada and India.

(7) Some tin deposits, e.g. at Rondônia, Brazil, can be related to Precambrian hot-spot activity and Sn was also associated with U, W, Nb, Ta, Li, and Be in syn- to post-collision granites and pegmatites.

(8) Mineralization, including chromite pods in ultramafic rocks and copper sulphides in pillowed lava, is most easily identifiable as of ophiolitic type within obducted rocks belonging to the Pan-African orogeny in the Red Sea area and Morocco, but there are also indications in the Chitradurga area of southern India of ophiolitic mineralization of much older age (c. 2500 Ma).

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Chapter 28

METALLOGENIC EVOLUTION AND PRECAMBRIAN TECTONICS

R. W. HUTCHINSON

ABSTRACT

Mineral deposits of widely differing types, and the geological settings in which they were formed, are products of particular broad tectonic environments. These controlled the nature and composition of igneous rocks associated with deposits, the morphology of their depositional basins — and therefore the nature and composition of associated sedimentary rocks — and also governed the metallic compositions of the ores themselves. Consequently the mineral deposits and their geological settings provide interpretive guides to tectonic environments.

Moreover, metallogenic studies reveal distinct evolutionary changes in mineral deposits and their formational environments through time. These changes, therefore, provide guides to tectonic evolution of the earth. Some types of mineral deposits are unique to Archaean or to Proterozoic rocks and are not found in Phanerozoic rocks. Other types persist through the entire span of time and occur in rocks of many ages from Archaean to Cenozoic. Still other types are restricted to, or are only extensively developed in, Phanerozoic or even in late Phanerozoic rocks, but are absent, or only poorly developed, in older rocks. These relationships indicate that tectonic processes have changed through time; that Precambrian tectonic processes were not identical to those of Phanerozoic plate tectonics, although some of the attendant mechanisms were indeed similar and have been repeated through time to produce similar deposit types.

Abundant geological evidence suggests evolutionary changes through earth history, not only in the nature of mineral deposits but in faunal development and in the composition of the atmosphere—hydrosphere. It therefore seems illogical to believe that tectonic processes have remained constant. Phanerozoic plate-tectonic processes probably represent a relatively late, evolutionary form of tectonism. The major problem, therefore, is not to seek their role in Precambrian time, but to decipher the nature of their Precambrian antecedents and the sequential, gradational changes in tectonism that occurred from earliest Archaean, through Proterozoic into Phanerozoic time.

Early Archaean greenstone-belt tectonism was similar to that at modern accreting plate boundaries but later greenstone-belt tectonic processes were like those in the early stages of subduction at consuming plate boundaries. Unlike plate-tectonic processes, however, Archaean tectonism involved mainly vertical movement and was dominated by subsidence. Rocks and ore deposits of the later Archaean were therefore superposed directly above the earlier ones. Proterozoic tectonism was dominated by rifting of continental crust formed during Archaean time. It may be likened to that of the later stages of subduction at consuming plate boundaries, to back-arc spreading and epeirogenic tectonism. Early Proterozoic rift fault displacement was mainly vertical, but major lateral displacement began in mid-Proterozoic time, perhaps initiated by global expansion. Lateral translation continued and, in late Proterozoic time, resulted in the major lateral interactions of oceanic and continental plates that characterize Phanerozoic plate tectonics.

INTRODUCTION

Studies of ore deposits have traditionally emphasized their economic importance or genetic processes, but their significance in the broader context of geological evolution has not been equally considered. Yet metallic ores are rocks of unique composition, and many exhibit distinct evolutionary changes in nature and geological setting from oldest Archaean to youngest Phanerozoic strata. Broad metallogenic comparisons between major deposit types of different ages reveal significant differences in the nature and composition of igneous rocks associated with the deposits, in the lithofacies of associated sedimentary rocks, in the depositional environment of their host rocks, and in the compositions of the ores themselves (Hutchinson, 1973, 1980). Such broad differences must be due to major changes in tectonic environments which alone are sufficiently broad to encompass and to influence all these geological characteristics. Consequently, the evolutionary changes in major mineral deposit types and their geological settings provide useful guides to earth's tectonic evolution.

It is not the purpose of this chapter to fully describe and document all the evolutionary changes that occur in the various major types of ore deposits. These have been outlined in detail for the massive base metal sulphide family of ore deposits (Hutchinson, 1973, 1980) and have also been summarized for iron-rich sedimentary rocks, nickel deposits, gold ores and uranium deposits (Hutchinson, in press). Only the main aspects and most important examples of these evolutionary changes are therefore outlined below. In this paper, time eras (e.g. Archaean) do not refer to definite, absolute geologic ages, because these differ in the Precambrian rocks of southern Africa and Canada. The terms are used, instead, in a relative age sense to describe the development of sequential geological environments from older to younger rocks.

ARCHAEAN MINERAL DEPOSITS

Important nickel, iron, gold and base metal deposits occur in the greenstone belts of all the world's Archaean cratons. The geologic settings and metallogenic characteristics of all these deposits (Hutchinson et al., 1971) are remarkably similar throughout the world. Moreover, these deposits are genetically all closely associated with one another and with the thick Archaean volcano-sedimentary greenstone-belt successions. This suggests a common link between all these deposits and the extensive Archaean volcanism that formed their host successions. Their volcanogenic affiliations are now widely recognized (Gross, 1966; Ridler, 1970; Naldrett, 1973; Hutchinson, 1973, 1976). Notwithstanding their common volcanic affiliations, these deposits and their associated rocks belong to two somewhat different petrochemical-petrogenetic groups of Archaean volcanic rocks. The older of these groups is distinctively ultramafic-komatiitic to tholeiitic. It contains important nickel sulphide, iron and gold deposits, but only minor base metals, as in the greenstone belts of the Rhodesian and Kaapvaal cratons of southern Africa, or those of the Yilgarn Block in Western Australia. The younger of the two groups is better differentiated and includes tholeiitic, calc-alkaline and silicic-alkalic rock types. It contains important iron, gold and base metal deposits but only minor nickel sulphide ores, as in the Abitibi greenstone belt of the Superior Structural Province in Canada. It is recognized that these two groups are not completely exclusive of one another and that in places their rock types and deposits occur together, or in close juxtaposition. Where this occurs, as in the Abitibi greenstone belt (Naldrett and Gasparini, 1971; Pyke, 1976; Jolly, 1977) the ultramafic-komatiitic group underlies the differentiated tholeiitic-calc-alkaline-silicic-alkalic group and appears to be the older. However, there does not seem to be a major break between the two, and the latter appears to be simply a later evolutionary phase of volcanism. These two broadly differing Archaean volcanic environments are, however, recognizable on a world-wide scale. Although both are important in Archaean rocks, their time-stratigraphic extent in younger rocks is markedly different, and this is the important aspect from the standpoint of Precambrian tectonics. The ultramafic-komatiitic to tholeiitic group of rocks with its mineral deposits is restricted to Archaean cratons and does not occur in younger terrains. The differentiated tholeiitic-calc-alkaline-silicic-alkalic group recurs repeatedly in younger strata.

Archaean nickel sulphide deposits are lenticular bodies of nearly pure iron-nickel sulphide or larger disseminations of these sulphides in extensively serpentinized, poorly differentiated, ultramafic-komatiitic to tholeiitic, extrusive or subvolcanic rocks. The best examples are the many deposits of the eastern Yilgarn Block, particularly those of Kambalda in Western Australia (Ross and Hopkins, 1975). They are found only in Archaean rocks, and although their specific genetic process is controversial (Lusk, 1976a, b; Groves et al., 1976) their origin is clearly linked to the unique ultramafic volcanism that occurred during the early stages of Archaean greenstone-belt development. This type is virtually unknown in post-Archaean rocks, although the deposit at Thompson, Manitoba (Zurbrigg, 1963) may be a younger ore body that was metamorphosed during the Hudsonian orogeny about 1.8 Ga ago (Davidson, 1972), and the yet-unmined deposits of northern Ungava in Quebec (Wilson et al., 1969) have some similarities to them and are of early Proterozoic age. In general, however, these deposits and their host rocks represent a tectonic environment that apparently became extinct at the end of, or perhaps even during, Archaean time.

Archaean iron formations are chemical sedimentary strata that are intercalated with rocks of both the ultramafic-komatiitic to tholeiitic group and the differentiated tholeiitic-calc-alkaline-silicic-alkalic group (Dubuc, 1966;

Boyum and Hartvikson, 1970; B.H.P. staff, 1975). They have been called Algoman-type by Gross (1966) and their iron supply was probably by extensive sea-floor hydrothermal leaching (Bischoff and Dickson, 1975; Hajash, 1975) from the subaqueous volcanic rocks with which they are everywhere interstratified. They are strongly reduced, except where affected by surficial weathering. Ferrous carbonate, silicate and sulphide facies are abundant and magnetite predominates over hematite in the oxide facies. Although also common in Archaean successions, their time-stratigraphic distribution in younger rocks is markedly different from that of the Archaean nickel sulphide deposits discussed above, for iron formations of this type also recur in rocks of virtually all later geologic ages, as in the Ordovician of New Brunswick (Smith and Skinner, 1958; Lea and Rancourt, 1958). These rocks and ores represent a tectonic environment that appeared first in Archaean time, but has been repeated frequently throughout earth's tectonic evolution.

Archaean gold deposits are lodes of widely variable morphology, from simple to en-echelon veins to vein stockworks and from concordant to discordant in attitude (Hutchinson, 1976). Like the Algoman-type iron formations, they occur in volcanic or volcanoclastic rocks of both the ultramafic-komatiitic to tholeiitic and differentiated tholeiitic-calc-alkaline-silicic-alkalic groups. They are present and comprise important gold mining districts in these rocks on virtually all the world's Archaean cratons as in the Porcupine, Kirkland Lake and other districts of the Canadian Superior Structural Province (Ridler, 1970; Pyke, 1976; Latulippe and Germain, 1979), the Yellowknife district of the Canadian Slave Structural Province (Boyle, 1955), the Morro Velho—Raposos area of Minas Gerais in Brazil (Tolbert, 1964; Ladeira, 1980), the Steynsdorp and Barberton districts of the Kaapvaal craton in South Africa (Viljoen et al., 1969; Anhaeusser, 1976), the many gold deposits of the Rhodesian craton (Fripp, 1976), the Ashanti gold fields of West Africa (Ntiamoah-Adjaquah, 1974), the Kalgoorlie district of the Yilgarn Block, and the deposits of the Pilbara Block in Australia (Naldrett and Wyatt, 1961; Tomich, 1974). Although their genesis is currently being reinterpreted (Boyle, 1955; Ridler, 1970; Hutchinson, 1976; Kerrich et al., 1977; Fryer et al., 1979; Kerrich and Fryer, 1979) its link with volcanism can hardly be denied, considering the worldwide distribution and remarkably consistent geological setting of these ores. This is reinforced by their time-stratigraphic distribution for, again like Algoman-type iron formations, these ores recur in similar volcano-sedimentary successions of younger age, as in the Ballarat—Bendigo districts of Victoria, Australia (Bowen and Whiting, 1975) and the Mother Lode of California (Knopf, 1929). Thus, they too reflect a tectonic environment that has been repeated many times during supracrustal tectonic evolution.

Archaean base metal deposits are stratabound, exhalative volcanogenic, massive sulphide lenses of a primitive, lead-deficient type (Hutchinson, 1973, 1980). They represent the opposite extreme, in a complete spectrum of Archaean mineral deposits, from the nickel sulphides because they occur only in rocks of the younger, differentiated tholeiitic-calc-alkaline-silicic-alkalic group. They probably formed (Hutchinson, 1973) by sea-floor fumarolic discharge of metalliferous hydrothermal brine that accompanied this volcanism. The many important deposits of the Superior Structural Province (Walker et al., 1975; Spence and de Rosen-Spence, 1975; Roberts, 1975; Franklin et al., 1975) are the best examples, but others occur in the Slave Province (Money and Hislop, 1976) and at Golden Grove in the Yilgarn Block of Western Australia (Reynolds et al., 1975). Again, as in the cases of Algoman-type iron formations and Archaean gold lodes, this particular type of massive base metal sulphide recurs repeatedly in similar volcanic rocks of younger geologic age, as in Proterozoic rocks at Jerome, Arizona (Anderson and Nash, 1972), in early Palaeozoic rocks as in northwestern Newfoundland (Tuach and Kennedy, 1978) and in Devonian rocks at West Shasta in California (Kinkel et al., 1956). The tectonic environment in which these rocks and ores are generated has reappeared periodically through time.

It is also important to consider the absence, or very poor development, of many important types of ore deposits in Archaean rocks. Magmatic nickel-copper sulphide and palaeoplacer gold ores are poorly developed, although the deposits of the Stillwater complex in Montana and some ores of the Pamour mine near Timmins, Ontario, may be, respectively, rare exceptions (Parans, 1967; Conn, 1979). Superior-type iron formations and several types of massive base metal sulphide deposits, including those of the exhalative sedimentary group and cupreous pyrite types (Hutchinson, 1980), are absent. Lead deposits of all types are essentially lacking. This is particularly well illustrated by the paucity of lead in the Archaean massive base metal sulphides where it occurs seldom in amounts that permit profitable extraction. Significant lead sulphide-bearing ores of other massive sulphide types, or of Mississippi Valley type, appear only in rocks of later geologic periods. Significant uranium deposits are also lacking in Archaean rocks, again appearing only in younger geological sequences. Porphyry copper deposits are poorly or incipiently developed in Archaean rocks. Although porphyry-like occurrences are well known (Blecha, 1974) very few are in production, and as a group these are insignificant in terms of world copper supply. Again, the great deposits of this type appear only in younger groups.

PROTEROZOIC MINERAL DEPOSITS

Early Proterozoic

In addition to the local recurrence of older, Archaean-like iron formations, gold lodes and massive base metal sulphides, early Proterozoic rocks contain new types of variants of nickel sulphide ore, iron formation, gold-uranium and massive sulphide deposits. These have their own distinctive geological settings, quite different from those of their Archaean antecedents.

Nickel sulphides occur with increased amounts of copper and platinoid elements, either disseminated in layers within, or as lenticular concentrations near the contacts of relatively fresh and unaltered, differentiated mafic intrusions. The best examples are the Merensky Reef of the Bushveld Complex in the Transvaal and the important deposits of Sudbury, Ontario (Cousins, 1969; Souch et al., 1969). Similar deposits also recur in rocks of later Proterozoic age as in the Duluth gabbro (Mainwaring and Naldrett, 1977) or the Great Lakes Nickel intrusion of western Lake Superior (Geul, 1970), in Devonian rocks at Noril'sk in the Soviet Union (Godlevski and Grinenko, 1963), and in late Palaeozoic rocks at Insizwa, South Africa (Scholtz, 1936). The tectonic environment governing the generation of these deposits and their host intrusions has clearly been periodically repeated in later time. Although their genesis remains controversial (Fleet, 1979; Naldrett, 1979), they are generally considered to be of magmatic origin. Mafic tholeiitic igneous activity was widespread in early Proterozoic time and formed thick basalt sequences like those in the Ventersdorp of the Transvaal (Brock and Pretorius, 1964a) or the Grão Para of Brazil (Tolbert et al., 1971), thick intrusive sills like the Nipissing Diabase—Sudbury Gabbro of Ontario (Thomson, 1956), and extensive continuous mafic dykes like the Jimberlana or Binneringie of Western Australia.

Early Proterozoic iron formations are of the Superior-type (Gross, 1966) which is markedly different from the earlier Algoman-type. These iron formations are much thicker, more extensive and occur with epiclastic and commonly algal carbonate rocks of shallow-water, cratonic basin or craton-margin lithofacies. They are much more highly oxidized than their Archaean counterparts; hematite predominates in the abundant oxide facies, and the reduced ferrous carbonate, sulphide and silicate facies are less common. Their time-stratigraphic distribution is unique. These iron formations occur in rocks of almost the same absolute age on virtually all the world's Precambrian shields and constitute an outstanding example of a worldwide metallogenic epoch. Examples include the Lake Superior and Labrador iron ranges of North America, the Cerro Bolivar deposits in Venezuela, the Serra dos Carajás and Quadrilátero Ferífero districts in Pará and Minas Gerais, Brazil, the Sishen and Thabazimbi districts of South Africa, the Hamersley iron ranges of Western Australia and the Krivoi Rog field in the Soviet Union (Tolbert et al., 1971; Trendall, 1973). Unlike their Algoman-type predecessors these iron formations do not recur significantly in younger rocks. They therefore reflect a widespread, but "once-only" depositional environment during supracrustal tectonic evolution that has not been repeated since early Proterozoic time. Like the volcanogenic Archaean nickel sulphide deposits they became extinct, disappearing from the geologic record after their initial appearance and abrupt proliferation about 2.2 Ga ago.

Probably the most distinctive of the new ore types characteristic of early Proterozoic rocks are the important gold-uranium palaeoplacers of the Witwatersrand, Elliot Lake and Jacobina districts in South Africa, Canada and Brazil (Brock and Pretorius, 1964a, b; Gross, 1968; Robertson, 1976). Other, similar, occurrences are known in the Moeda Formation of Minas Gerais, Brazil, the Nullagine rocks of Western Australia and the Tarkwaian sequence of Ghana. These ores are the oldest important uranium deposits and the most important for gold. They contain varying proportions of these two metals and occur as concordant, although elongated, channel-like bodies in distinctive, coarse quartz-pebble conglomerates with abundant pyritic matrix and clasts. These rocks and other associated quartzites are clearly of shallow-water, fluviatile-deltaic lithofacies. In the extensively studied Witwatersrand (Brock and Pretorius, 1964a, b) and in the Huronian of northern Ontario (Card and Hutchinson, 1972) they were deposited along active, downthrown, fault-controlled margins of a major, asymmetrical, yoked basin. Judging from the widespread distribution of these rocks and ores, this distinctive tectonic environment appeared on many continents and craton margins in early Proterozoic time. Source for the gold and uranium was weathering of surrounding uplifted Archaean greenstones containing gold lodes, as well as granites and pegmatites generated at the close of Archaean time. Although perhaps not deposited in tectonically identical basins, younger palaeoplacer and placer gold deposits were formed following uplift and granitic plutonism in orogenic belts of many later geologic periods as, for example, the low-level and high-level Tertiary gravels and stream placers of the California foothills (Lindgren, 1911) or the placers of Ballarat-Bendigo in Victoria, Australia (Bowen and Whiting, 1975).

Early Proterozoic massive base metal sulphide deposits are volcanogenic and resemble those of Archaean age. In general, however, they are richer in lead and silver and are associated with increasingly felsic, fragmental-pyroclastic and possibly partially subaerial volcanic rocks. The petrochemical affinities of these rocks are poorly known but the assemblage may be bimodal rather than differentiated; calc-alkaline types may be rare. Intercalated sedimentary rocks are increasingly epiclastic and of cratonic provenance; carbonate- and sulphate-rich strata are increasingly abundant, particularly in rocks younger than about 2.0 Ga (Hutchinson, 1973). Examples are the little-known deposits of the White-water Group within the Sudbury Basin (Martin, 1957), those of the Prescott area in Arizona (Gilmour and Still, 1968) and some of the deposits of the Flin Flon-Snow Lake district in Manitoba (Byers et al., 1965; Sangster, 1972). Many important deposits of this polymetallic massive sulphide type also recur in younger rocks, as in the Tasman belt of eastern Australia (Burton, 1975; Davis, 1975), the Ordovician of New Brunswick (Lea and Rancourt, 1958; Smith and Skinner, 1958) and the Tertiary of Japan (Ishihara, 1974).

As in the case of Archaean successions, many important mineral deposit types are lacking or are only poorly developed in early Proterozoic rocks. In particular, these include massive base metal sulphide deposits of the exhalative sedimentary group and cupreous pyrite type, all uranium ores other than the palaeoplacers and, again, the porphyry coppers.

Mid-Proterozoic to late Proterozoic

Like early Proterozoic successions, mid-Proterozoic strata are hosts to some earlier deposit types but also to distinctive, new, diverse types and their geological settings, two of which are particularly significant. Firstly, massive base metal sulphide deposits of the lead-zinc-silver-rich, exhalative sedimentary group (Hutchinson, 1980) appear and form numerous, large and important ore bodies. The trough-deposited, clastic sediment-hosted type (Morris et al., 1972; Johnson and Klingner, 1975) appears characteristic of mid-Proterozoic rocks, whereas the shelf-deposited, carbonate sediment-hosted type may be somewhat younger and best developed in late Proterozoic rocks (Lea and Dill, 1968). Both types also recur in rocks of later geologic periods (Derry et al., 1965; Anger et al., 1966; Pereira, 1967; Gwosdz et al., 1974; Carne, 1976). Moreover, there is a corresponding and significant absence in rocks of this age of all the copper-gold-rich exhalative volcanogenic types of massive base metal sulphides and of the various types of volcanic rocks that contain them, which were important in older rocks, and recur in younger ones. Matching this interesting "volcanogenic gap" is a lack of the volcanogenic gold lodes which were also important in older rocks and recur in younger ones. These remarkable relationships must reflect the advent of a new and distinctive tectonic environment in mid-Proterozoic time that did not favour recurrence of the older, exhalative volcanogenic massive sulphides but permitted generation of the exhalative sedimentary types.

Secondly, two new and important types of uranium deposits appear in mid- to late Proterozoic rocks. These are the rich unconformity-related lenses or veins such as Ranger, Jabaluca and Nabarlek in the Northern Territory of Australia (Hills and Thakur, 1975; Rowntree and Mosher, 1975) or Rabbit Lake, Key Lake and Cluff Lake of northern Saskatchewan (Knipping, 1974; Hoeve and Sibbald, 1978) and the granite- or pegmatite-affiliated deposits like those of the Grenville Structural Province in south-eastern Ontario (Satterly, 1957; Hewitt, 1967). The Rössing deposit in the Damara belt of Namibia (South West Africa) may also belong to this category (Berning et al., 1976) although the mineralized alaskite there is only c. 500 Ma old but represents the product of anatectic melting of mid-Proterozoic gneissic basement and late Proterozoic psammitic sediments (A. Kröner, pers. commun., 1980). The unconformity-related types also recur in younger rocks as at the Schwartzwalder and Pitch mines of Colorado (Sheridan et al., 1967; Young, 1979), but there are no important younger examples of the granite-pegmatite types.

Once again, as in the case of Archaean and early Proterozoic rocks, some mineral deposit types that are important in younger strata are lacking, or are only rarely developed in mid- to late Proterozoic rocks. These include sandstone-type uranium deposits, the cupreous pyrite type of exhalative volcanogenic massive base metal sulphide, Mississippi Valley type zinc-lead ores and porphyry copper deposits, although the Haib deposit in Namibia (Minnitt, 1979) is a rare mid-Proterozoic example of the last.

LATEST PRECAMBRIAN AND PHANEROZOIC MINERAL DEPOSITS

The most outstanding general characteristic of Phanerozoic mineral deposits is their extreme diversity. This results from the reappearance in Phanerozoic rocks of most, but not all, older Precambrian deposit types and geological settings, plus the appearance of additional new or changed varieties with their own distinctive settings. It is not possible here to discuss all these, but a few of the distinctive new types or variants are pertinent.

There are interesting changes in nickel deposits although these are relative, not absolute. Nickel sulphide deposits of the volcanogenic Archaean type are absent, and those of the Proterozoic magmatic type are rare. Conversely, lateritic nickel deposits, which are only rarely developed on Precambrian ultramafic rocks, are numerous and much more important on Phanerozoic ultramafic bodies (Cumberlidge and Chace, 1968; INAL staff, 1975). There are also changes in iron-rich sedimentary rocks. Volcanogenic Algoman-type iron formation recurs, but the early Proterozoic Superior-type does not. The distinctively different, much smaller, Clinton and Minette types of ironstones appear in Palaeozoic and Mesozoic strata, respectively (Hollingworth and Taylor, 1951; Simpson and Gray, 1968; Hutchinson, in press). Perhaps most significant, tectonically, is the initial appearance of the distinctive cupreous pyrite type of exhalative volcanogenic massive base metal sulphide in the pillowed, tholeiitic basalts that are integral parts of ophiolite sequences. The oldest well-documented examples are those of early Palaeozoic age in western Newfoundland (Duke and Hutchinson, 1974; Bachinski, 1976), apparently formed in oceanic crust of the Proto-Atlantic (Church and Stevens, 1971). The most important deposits, however, are those of Cyprus, Turkey and Oman (Bear, 1963; Griffits et al., 1972; Carney and Welland, 1974; Gealey, 1977) in the Mesozoic Tethyan belt. Deposits of this type are unknown in Precambrian rocks and Glikson (1979) has recently discussed the similar absence of their host rocks. Mississippi Valley and Alpine-type zinc-lead deposits, which are rare in Precambrian strata, become widespread and extremely important, particularly in Cambrian to Triassic rocks (Maucher and Schneider, 1967; Sangster, 1976). Porphyry copper deposits, also rare and poorly developed in Precambrian successions, are numerous and important in still younger, late Palaeozoic to Quaternary rocks (Titley and Hicks, 1966; Sillitoe, 1973; Lowell, 1974).

Finally, significant new types of uranium deposits appear in Phanerozoic strata. These include uraniferous black carbonaceous marine shales in Palaeozoic rocks like those of the Rostad area of Sweden (Armands, 1972a, b) and the important continental sandstone-type deposits in Mesozoic-Tertiary rocks, for example those of the Colorado Plateau or the intermontane basins of Wyoming (Harshman, 1968; Fischer, 1974;) in the western United States.

TECTONIC SIGNIFICANCE

A consistent pattern of mineral deposit distribution is evident through all these major time divisions. In each successively younger period some older deposit types recur, a few disappear, new ones appear, but some, important in still younger periods, are as yet not present or insignificantly developed. The resulting occasional extinctions, prolonged continuations, sudden proliferations and diversifications of types are strongly reminiscent of faunal evolution, to which metallogenic evolution therefore appears closely analogous. It is, however, the tectonic significance of these patterns which is the main point of this paper. Insofar as the mineral deposit types and their distinctive geological settings reflect particular tectonic environments, these time distribution patterns clearly indicate evolutionary changes in earth's tectonic environments and processes through time. Some early, primitive tectonic environments disappeared, others were somehow repeated periodically, but also underwent changes or were joined by new ones. It follows that the relatively well-understood, plate-tectonic processes, which have dominated crustal evolution during Phanerozoic time, are a late evolutionary stage of prolonged tectonic evolution on the earth, hence were not operative during at least part, and perhaps most, of Precambrian time. Nevertheless, it also follows that plate tectonic mechanisms somehow resulted in recurrence of some, but not all, tectonic environments resembling those of older time periods. The challenge in applying plate-tectonic concepts to Precambrian tectonism lies not in their direct applicability, which is doubtful, but in using them as guides to reconstruct the prior tectonic evolutionary stages that preceded them. Representing the latest stage in tectonic evolution they constitute an essential, relatively well-understood starting point for this reconstruction. Moreover, the patterns of disappearing, recurring and diversifying tectonic environments through time are important in the reconstruction, for although some plate-tectonic mechanisms must be new diversifications, others are replicas of older ones and these, in true sense of uniformitarianism, provide keys to their predecessors.

TECTONIC INTERPRETATIONS

Phanerozoic plate tectonics and mineral deposits

The plate-tectonic environments of the major mineral deposit types

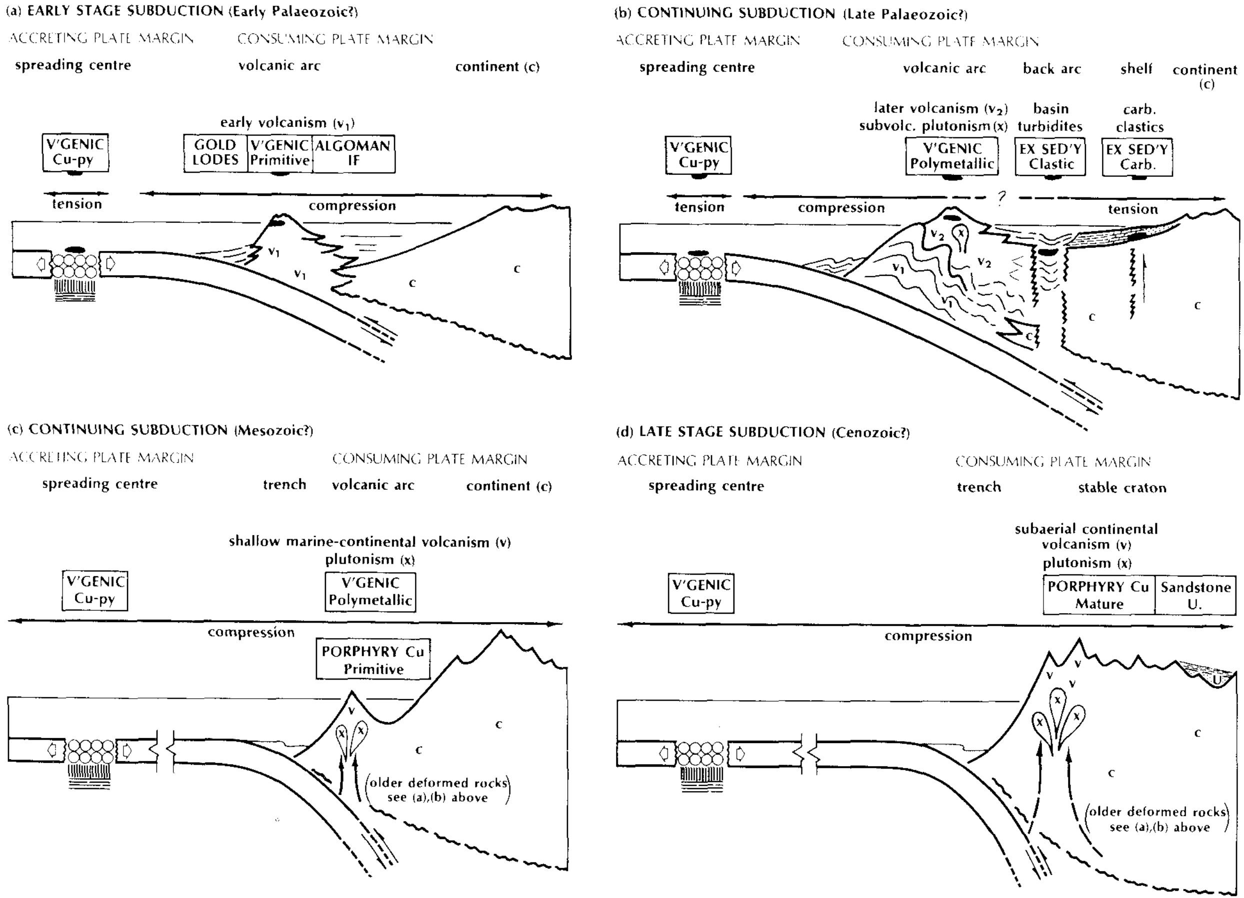

Fig. 28-1. Possible evolutionary stages in consuming plate boundaries. For explanation see text.

that occur in Phanerozoic rocks, both recurrent older types and newly appearing ones, are reasonably well understood. The cupreous pyrite type of massive base metal sulphide is formed in the mid-ocean rift-ridge environment (Fig. 28-1a, left) in ophiolites generated at spreading centers along accreting plate boundaries (Degens and Ross, 1969; Hutchinson, 1973; Francheteau et al., 1979). These ores and rocks are later transported to continental margins by sea-floor spreading. Here some are incorporated into the continent and preserved by obduction, others are subducted, reworked and destroyed. The absence of rocks and ores of this type in Precambrian successions (Hutchinson, 1973; Glikson, 1979) is strong evidence that these plate-tectonic processes were not active in Precambrian time. Nevertheless, although the two are not identical, there are some similarities between these ophiolites and the older Archaean extrusive rocks of ultramafic-komatiitic affinity. This similarity may be pertinent to early Archaean tectonism in the greenstone belts and if so, it is significant for the conclusion that plate-tectonic processes generate new oceanic crust from and on the mantle, with no underlying or nearby granitic crust.

The primitive copper-zinc-rich type of massive base metal sulphide, Algoman-type iron formations and lode gold deposits, all similar to their more numerous and larger late Archaean antecedents, are formed in the differentiated rocks of the volcanic arc environment, probably during early stages of subduction of oceanic crust along consuming plate boundaries (Fig. 28-1a, right). This plate-tectonic configuration is therefore the one that recreates the late Archaean-like tectonic environment. In using it as a guide to reconstruct the latter it is significant that subduction and reworking of earlier ophiolitic, mid-ocean rift-generated crust gives rise to this environment. During somewhat later stages of continuing subduction along consuming plate margins more felsic, late- or post-arc volcanism occurs (Fig. 28-1b, right) in shallower, craton-margin basins formed by back-arc spreading and tensional tectonics along the continental side of the orogen. This is the tectonic environment in which the polymetallic type of volcanogenic massive base metal sulphide deposit is formed (Hutchinson, 1973) and may be compared to the current Japanese margin of the Pacific ocean. The two lead-zinc-rich exhalative sedimentary types may also be generated here, either in turbiditic clastics deposited within deep rift-controlled troughs, or in shallow-water, carbonate-clastic sediments deposited on the flanking cratonic shelfs (Hutchinson, 1980). Mississippi Valley-type zinc-lead ores are closely related genetically to the latter. They are also formed in shelf-facies carbonate rocks of this environment, probably by diagenetic reactions between unconsolidated carbonate sediment and circulating metalliferous hydrothermal fluids that did not reach the sea floor (Hutchinson, 1980). Porphyry copper deposits are associated with subvolcanic plutonism and are also generated over subduction zones along consuming plate boundaries (Mitchell and Garson, 1972). It has also been suggested (Hutchinson, 1980) that their generation occurs during still later stages of subduction along consuming plate boundaries. During these stages continued plate collision first closes the back-arc basin (Fig. 28-1c), forming margins like those of the West Pacific with primitive porphyry coppers (Sutherland Brown, 1976; Gustafson and Titley, 1978), and subsequently causes volcanism to become subaerial and continental (Fig. 28-1d), rather than subaqueous and marine, as in the earlier stages. This results in continental margins like the Andean with a somewhat different, mature type of porphyry copper deposit (Titley and Hicks, 1966) but few massive base metal sulphide deposits. Evolutionary changes in consuming plate margins and their mineral deposits have been ascribed to progressive changes in the dip of the underlying subduction zones (Fig. 28-1) from shallow to steep as subduction continues (Uyeda and Nishiwaki, 1980). These interpretations offer an explanation for the differing space and time distribution of the various types of massive sulphide deposits, Mississippi Valley-type ores and for the different types of porphyry coppers. The various Phanerozoic plate-tectonic configurations that cause later recurrence of the Precambrian deposit types also provide uniformitarian guides to their earlier, Precambrian tectonic environments and processes.

Finally, uraniferous, black, carbonaceous marine shales of early Palaeozoic age were formed by syngenetic precipitation of uranium from seawater in reducing marine environments. In contrast, the late Palaeozoic—Mesozoic—Tertiary sandstone type uranium deposits were formed in continental coarse clastic sediments (Fig. 28-1d), often associated with continental volcanism, and following periods of uplift, with resulting weathering of granitic terrain. The clastic host rocks commonly rest unconformably on the uplifted surface. Uranium source may have been weathering of either continental volcanic rocks, granitic terrain, or both. Uranium transport was in oxidized circulating meteoric groundwaters, and precipitation was commonly due to their reduction by organic debris buried in the rapidly deposited clastic sediments (Rackley, 1976). This particular geologic setting cannot have existed until the evolutionary appearance of terrestrial organisms in Devonian time which explains the time and space distribution of these two types of Phanerozoic uranium deposits. Both types have some similarities to Proterozoic uranium deposits, and their tectonic environments again provide guides to those of their older antecedents.

Precambrian tectonics

It remains to attempt an interpretive reconstruction of Archaean and Proterozoic tectonics using as guides comparisons between Precambrian and Phanerozoic mineral deposit types and settings and also the known plate-tectonic environments of the latter. On these bases it appears firstly that Archaean tectonic processes must have been generally similar to those

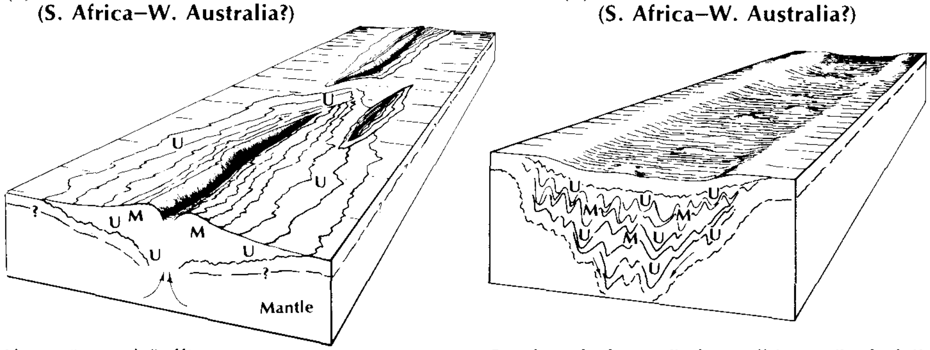

Fig. 28-2. Possible Precambrian tectonic evolution: early to mid-Archaean.

at accreting plate boundaries and also to those of the earliest subductive stage along consuming boundaries. Proterozoic tectonic processes, on the other hand, appear more closely akin to tectonism in the later subductive stages along consuming boundaries and to intracontinental epeirogenic tectonism.

The comparison between Phanerozoic ophiolites and Archaean ultramafic komatiitic-tholeiitic successions suggests that earliest tectonism in the Archaean greenstone belts was dominated by oceanic rifting. This occurred either through a very thin older granitic crust or, perhaps, as in the case of modern oceanic rifts, directly on the Archaean protomantle (Fig. 28-2a). It resulted in ultramafic komatiitic-tholeiitic volcanism forming "new" proto-oceanic crust, containing volcanogenic nickel sulphide deposits, Algoman-type iron formations and gold lodes. This early Archaean protomantle may have been poorly differentiated, rich in reduced sulphur gases with high f_S and low f_{O_2}. This would have promoted generation of Fe-Ni-S melts, or Ni may also have been hydrothermally leached from the subaqueous volcanic rocks, either process generating the associated volcanogenic nickel sulphide ores. Continuing mantle degassing, progressive generation of thickened sialic crust and later oxygenation of the atmosphere—hydrosphere—lithosphere would all have combined to prevent later repetition of these primitive tectonic conditions. This may explain the absence of these deposits and their characteristic geologic setting in younger rocks.

Modern accreting plate margins, although not identical, are, however, close analogues of this primitive Archaean tectonic environment.

Prolonged ultramafic volcanism resulted in gravitational instability along the oceanic rifts with major subsidence or "vertical subduction" (Fig. 28-2b; *see also Goodwin, this volume, Chapter 5, ed.*). Partial melting of the subsided proto-oceanic crust subsequently generated the differentiated tholeiitic-calc-alkaline-silicic-alkalic volcanic sequences with their Algoman-type iron formations, gold lodes and primitive type of copper-zinc-rich massive sulphide deposits. Although the magmatic processes were like those which generate similar differentiated volcanic rocks and deposits through reworking of oceanic crust by subduction along Phanerozoic consuming plate boundaries, the tectonic mechanism was different. Vertical subsidence of Archaean proto-oceanic crust, rather than subduction due to Phanerozoic lateral plate motions, caused vertical superposition of the differentiated Archaean volcanic rocks above their ultramafic-komatiitic progenitors (Fig. 28-3a; *see also Kröner, this volume, Chapter 3, ed.*) rather than generation of the differentiated suite along the subducted margin of the oceanic slab, as in modern plate tectonics. The vertical interactions of Archaean tectonism may have been governed by smaller mantle convection cells and more rapid convective circulation, resulting in hot-spot tectonics (Fyfe, 1978; *see also Lambert, this volume, Chapter 18, ed.*).

Although its sources and mechanisms are obscure, prolonged Archaean volcanism culminated in worldwide granodioritic plutonism of the late Archaean orogeny (Fig. 28-3b). This tectonism finds its younger analogue in the granitic plutonism that so commonly follows early volcanism in Phanerozoic orogenic belts along consuming plate boundaries, and rare-element pegmatites or contact metasomatic deposits are its common products. Younger examples are the Devonian Acadian orogeny of the Appalachians in eastern North America or the Cretaceous Nevadan orogeny of the Cordilleran belt in the western United States. This type of tectonism marked the end of Archaean time, uplifted and exposed the greenstone belts to erosion and appears to have generated, possibly for the first time, a worldwide stable, thick continental crust. Rifting of this continental crust (Kumarapeli and Saul, 1966; Kanasewich, 1968; Hutchinson, in press) became the dominant tectonic process of Proterozoic time and controlled the nature of Proterozoic magmatism as well as the evolution, morphology and sedimentation of Proterozoic basins.

In early Proterozoic time displacement on these rifts was mainly vertical (Fig. 28-3c), perhaps still reflecting the earlier, predominantly vertical, tectonics of Archaean time. This permitted recurrence of differentiated volcanic successions with Algoman-type iron formations, gold lodes and the primitive copper-zinc-rich volcanogenic massive base metal sulphides, particularly where older Archaean or Proterozoic mafic rocks were reworked by deep subsidence. Where felsic-granitic rocks were similarly reworked,

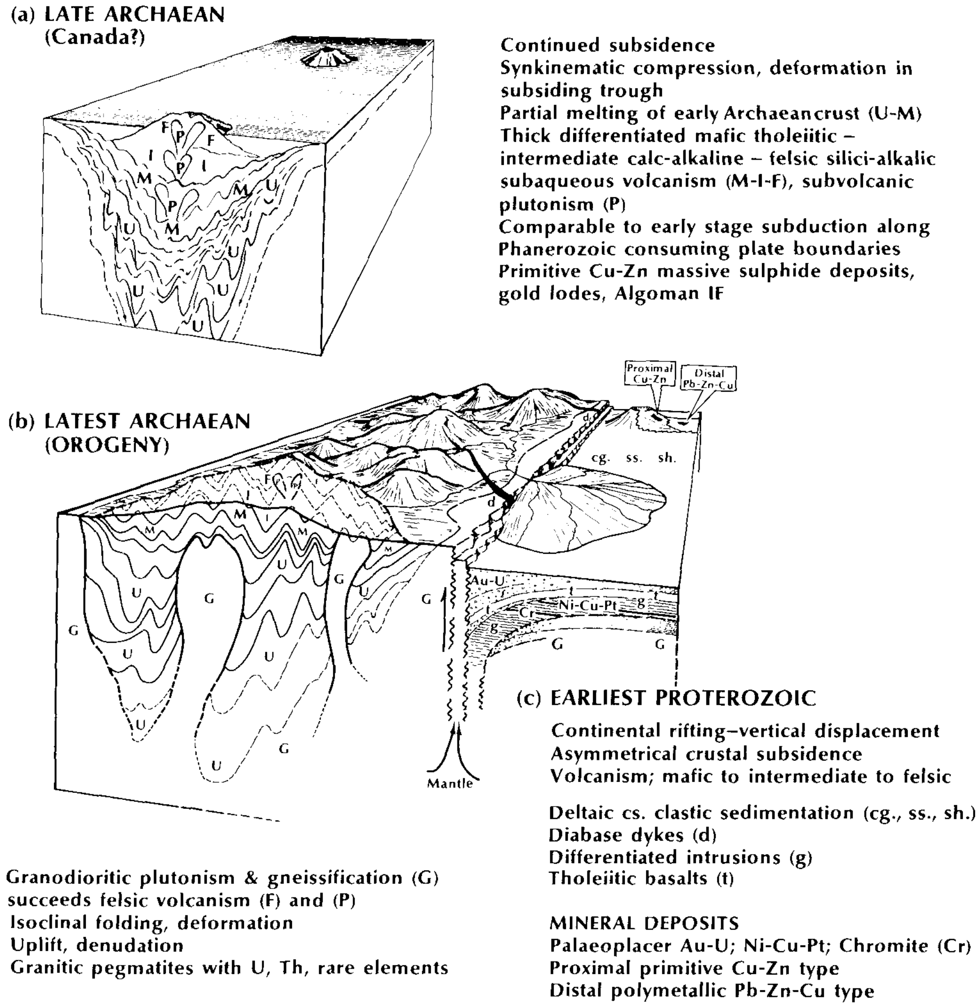

Fig. 28-3. Possible Precambrian tectonic evolution: late Archaean to earliest Proterozoic.

more siliceous, ensialic volcanism occurred and was accompanied by generation of the new polymetallic volcanogenic massive base metal sulphide type. Sulphate gangue minerals are common in these deposits but rare in the earlier, primitive Archaean type which suggests that early Proterozoic oxygenation of the atmosphere—hydrosphere (Garrels et al., 1973) was also a factor in the evolutionary appearance of the polymetallic type. Increased f_{O_2} may have decreased lead solubility (Hutchinson, 1973), resulting in increasing lead content in these areas.

Early Proterozoic vertical rifting also formed major, asymmetrical, yoked basins like the Witwatersrand or the Huronian of northern Ontario (Brock and Pretorius, 1964a, b; Card and Hutchinson, 1972). Here the gold-uranium palaeoplacer deposits were formed, these elements derived from weathering of the uplifted greenstone belts in fluviatile-deltaic sediments. These were deposited as thick elastic wedges, perhaps fringed by stromatolitic-algal offshore reefs, along the vertically subsiding margin of the tilted, rift-controlled block (Fig. 28-3c). Major, mantle-generated, mafic magmatism accompanied the rifting, forming major mafic intrusions like the Nipissing diabase of Ontario, widespread and continuous diabase dykes like those seen in most Archaean cratons, or thick extrusive successions like the Ventersdorp lavas of the Witwatersrand basin (Brock and Pretorius, 1964a) or the Grão Pará basalts of Serra dos Carajás in Pará, Brazil (Tolbert et al., 1971). In some of the layered, differentiated intrusions the magmatic nickel-copper-platinoid sulphide deposits were formed, probably where the intrusions were locally sulphurized.

Oxygenation of the oceans in early Proterozoic time, probably due to biogenic evolution (Garrels et al., 1973), resulted in relatively rapid, extensive deposition of the new Superior-type iron formations. This occurred on the shallow, hinged flanks of the yoked basins, possibly in pools of evaporite-like precipitation behind extensive algal banks (Fig. 28-4a). Here oxygen supply from these organisms was direct and abundant, causing ferric-iron precipitation from seawater which was relatively rich in ferrous iron, due to prior reducing conditions and to the extensive preceding mafic marine volcanism of Archaean—earliest Proterozoic time. In this way ferrous iron may have acted as a buffer to oxygenation of the oceans until their iron content was reduced to its current low levels. This might explain the narrow time-stratigraphic range and worldwide distribution of Superior-type iron formations and their absence in younger rocks.

In mid-Proterozoic time, however, major lateral separations began along the continental rifts (Fig. 28-4b) and the two lead-zinc-rich exhalative sedimentary massive sulphide types were formed, the clastic-hosted type in thick, turbiditic sediments deposited in opening and deepening rifts (Kanasewich, 1968) or in aulacogens, and the carbonate-hosted type on flanking shallow cratonic shelfs. Mafic magmatism again accompanied separation along the rifts and is illustrated by amphibolites and mafic igneous rocks that commonly accompany these ores. This tectonic environment has its younger plate-tectonic analogue in the tensional regime of back-arc spreading and basins. In mid-Proterozoic time it may have been initiated by a period of global expansion, resulting in the first appearance of deep-ocean basins separating major continental blocks. Global expansion and resulting extensional tectonics might explain the lack of compression-generated, differentiated, calc-alkaline volcanic rocks (Christiansen and Lipman, 1972; Miyashiro, 1975) in mid-Proterozoic successions and the corresponding absence of the lode gold deposits and volcanogenic massive sulphides which they commonly contain. It might also explain the absence of obducted ophiolites in rocks of this age (Glikson, 1979).

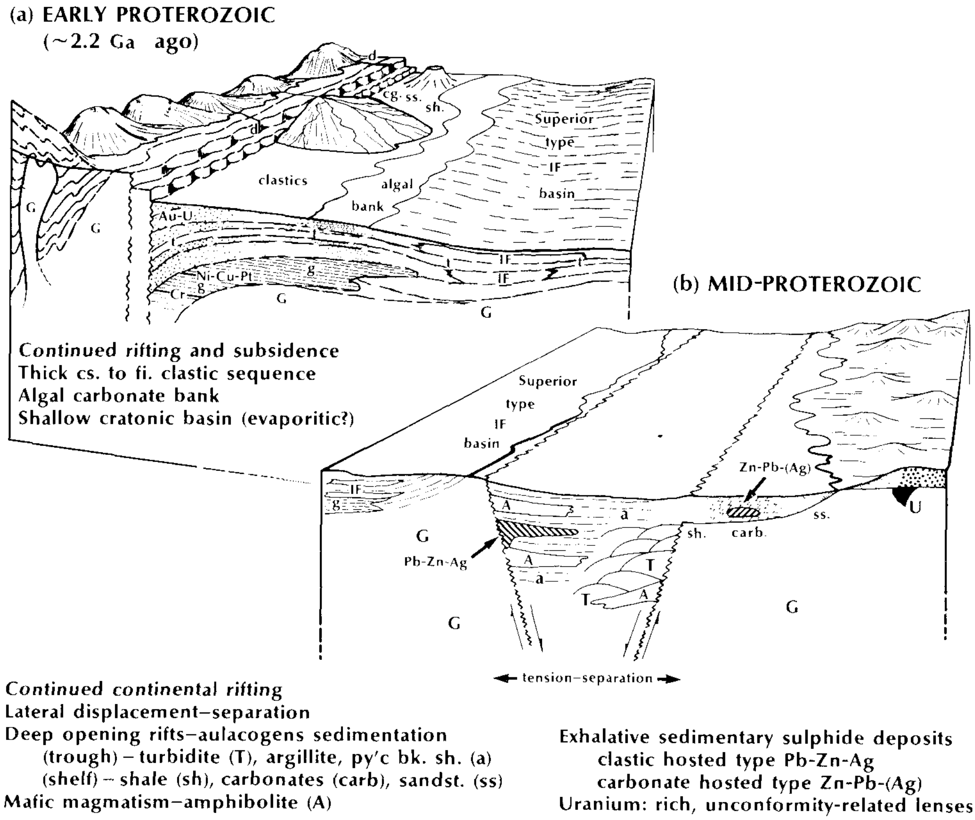

Fig. 28-4. Possible Precambrian tectonic evolution: early to mid-Proterozoic.

Solution transport of uranium became possible in mid-Proterozoic time, following prior oxygenation of the atmosphere—hydrosphere. This led to uranium extraction by oxygenated meteoric waters from granitic or earlier palaeoplacer source rocks that had been uplifted and exposed to weathering during mid-Proterozoic orogenesis. These circulating groundwaters transported soluble U^{6+} through porous, basal, coarse, continental clastic sediments deposited on the uplifted surface. Uranium precipitation occurred where they encountered any reducing agent such as graphitic or sulphidic rocks, or even by ground-water reduction at the palaeo-water table (Fig. 28-4b). Soluble U^{6+} was thereby reduced to insoluble U^{4+} and the surrounding rocks were correspondingly oxidized. This would explain the widespread association of these ores with unconformities, their high grade and lenticular nature, common hematitic wall rock alteration and minor element associations. Thorium is deficient in these ores because, unlike uranium, it was not readily oxidized to a soluble ion, hence was not mobilized and transported. Minor metals such as nickel, silver and base metals occur in the ores because they too, were soluble in circulating groundwaters and precipitated by reduction. This tectonic environment has its plate-tectonic analogue in the continental sandstone-type uranium deposits of late Palaeozoic—Tertiary age which share many of these characteristics but in which buried terrestrial organic matter appears to have been the major reducing precipitant. The granitic-pegmatitic type uranium deposits of mid-Proterozoic age are probably the result of orogenic reworking of the other types which generated uranium-thorium-rich granitic-pegmatitic rocks.

Finally, the major rift separations of mid-Proterozoic time continued and, by late Precambrian-early Palaeozoic time, evolved into the various mechanisms and configurations of modern plate tectonics. These are characterized by important lateral translations and interactions of oceanic and crustal plates. As outlined above, some plate-tectonic configurations are repetitions of earlier tectonic situations, and here old deposit types and settings recur, although on reduced scales of both space and time. Other configurations represent new tectonic environments or variants, and here new deposit types occur. The space and time distribution of Phanerozoic mineral deposit types also indicates that there have been evolutionary changes in subductive, consuming plate margins through Phanerozoic time, both during the development of individual margins, and perhaps also generally for many margins (Hutchinson, 1980).

SUMMARY AND CONCLUSIONS

Evolutionary metallogenic trends show that some Precambrian mineral deposit types and their settings disappear, whereas others periodically reappear in younger rocks. In addition, new deposit types or variants also appear in younger rocks, and some of these rapidly become widespread and important. Still other deposit types, important in Phanerozoic rocks, are absent or poorly represented in older ones. These extinctions, continuations, proliferations and diversifications are reminiscent of those in faunal evolution.

These evolutionary changes in mineral deposits and settings result from changing tectonic processes and environments through time. They therefore provide useful guides to palaeo-tectonism. Evolutionary changes in tectonism are consistent with evolutionary changes in other terrestrial processes such as organic and atmospheric—hydrospheric development, but static tectonism is not. It is therefore unlikely that Phanerozoic plate-tectonic processes were operative during Precambrian time. The plate-tectonic environments of Phanerozoic deposit types, however, are well understood and, together with comparisons between Phanerozoic and Precambrian types, permit interpretive reconstruction of the sequential Precambrian tectonic environments that led to Phanerozoic plate tectonics.

The tectonic environment of early Archaean greenstone belts was similar to that of Phanerozoic accreting plate margins, but late Archaean greenstone belt tectonism was like the early stages of subduction along consuming plate margins. In Archaean time, however, subduction of rocks formed in the early stage was by vertical subsidence, hence rocks of the late stage were generated from, but were superposed directly above the former. Nevertheless, petrogenetic and ore genetic processes were similar to those of accreting, early-stage consuming plate margins. Worldwide granitic plutonism at the end of Archaean time uplifted the greenstone belts and formed a thick, stable sialic crust. This tectonism is similar to that of batholithic intrusion following the eugeosynclinal stages of orogenic belt development along consuming plate margins.

Proterozoic tectonism was dominated by rifting of continental crust, and widespread mafic magmatism accompanied the rifting. Early Proterozoic rift displacement was mainly vertical, forming major asymmetrical yoked basins and controlling the distribution of sedimentary lithofacies and related new mineral deposit types. Oxygenation of the atmosphere—hydrosphere during early Proterozoic time was important in formation of other new deposit types. Mid-Proterozoic rift displacement, however, was mainly lateral, perhaps due to global expansion, and may have formed the deep ocean basins separating the major rifted continental blocks. Lateral separation along the rifts continued and, by late Proterozoic time, evolved into the familiar Phanerozoic plate tectonics involving major lateral interaction of oceanic and continental crustal plates.

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