The Pleistocene Boundary and the Beginning of the Quaternary

EDITED BY J. A. VAN COUVERING

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Contents

Part I Definition of the base of the Quaternary

Part II Characterization of the Pleistocene boundary-stratotype

Part III The paleontological context of the Pleistocene boundary

Part IV The Pleistocene boundary in regional sequences

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Preface: the new Pleistocene

JOHN A. VAN COUVERING

Nailing down the golden spike

The goal of the International Geological Correlation Program, Project 41 (IGCP-41), was to propose a formal definition for the base of the Pleistocene Series – by international agreement the lower (and only) boundary of the Quaternary subsystem – in accordance with the resolution passed by the XVIII International Geological Congress (IGC) in London in 1948 (King and Oakley, 1950), as well as to characterize and globally correlate this boundary. After years of study and discussion, as described by K. V. Nikiforova and M. N. Alekseev in the first chapter of this volume, the members of IGCP-41 agreed on a boundary defined by a physical reference point or "golden spike" within the outer-shelf marine section exposed at Vrica, near Crotone, Calabria. As this book demonstrates, the members of IGCP-41 were able to recognize this level in marine and nonmarine sections around the world. The Vrica boundary-stratotype for the Pleistocene boundary was ratified and adopted by the international authorities and has gained approval as a global boundary-stratotype section and point (GSSP) (Cowie and Bassett, 1989).

The participants in IGCP-41 may take pride in the establishment, at long last, of an internationally accepted and unambiguous definition for the base of the Pleistocene. With regard to this momentous advance, this book has three aims: (1) to document the Vrica boundary-stratotype; (2) to show how the boundary-stratotype can be characterized in terms of the paleontological and paleoenvironmental record, radiometric dating and magnetostratigraphy; and (3) to describe the worldwide stratigraphic context of the boundary.

Editorial history and notes on style. The great majority of manuscripts for this book were in hand by the end of 1984, although a few solicited chapters and revisions were submitted in 1987. Progress in preparing the book was regrettably slow, but all manuscripts were circulated for updating in 1992, and most chapters (notably Chapter 2, on Vrica itself, and Chapter 10, on the rapidly changing context of human evolution) also incorporate even more recent information. In particular, age determinations derived from the paleomagnetic time scale have been corrected in the text to the current orbitally tuned calibration of the geomagnetic polarity time scale (GPTS) (Table 1), although some figures could not be revised.

In editing this work, I have held to the general principle that authors should be allowed to speak with their own voices in regard to matters of taste such as "planktic" versus "planktonic," "ostracod" versus "ostracode," and "-logic" and "-graphic" (as in geologic and stratigraphic) versus "-logical" and "-graphical." As for the problems introduced by my attempts to standardize the terms and usages in this volume, I am grateful for the patience of my colleagues.

Modern stratigraphic philosophy is based on the principle that *base defines boundary,* which is to say that the base or beginning of each unit also defines the top of the preceding unit (Salvador, 1994). In view of this, the terms "Pliocene–Pleistocene boundary" or "Neogene/Quaternary boundary" actually mean the same as "Pleistocene boundary" or "Quaternary boundary," and both usages are found in this work as space or taste dictates. With regard to punctuation, many authors have used the virgule, as in "Pliocene/Pleistocene," rather than the dash, as in "Pliocene–Pleistocene," to express the boundary between two units such as these. The older term is written first, as is stratigraphically logical, but unfortunately the line of the virgule symbolizes a boundary, and its slant might be taken to suggest that the older unit rests on the younger. For this reason, I have preferred the dash where there was a choice. The term "Plio–Pleistocene" is restricted herein to mean a body of strata or time that includes both Pliocene and Pleistocene parts.

Is there a "Neogene–Quaternary" boundary?

The International Union of Geological Sciences (IUGS) currently recognizes the units "Paleogene," "Neogene," and "Quaternary" as the three periods making up the Cenozoic era (Cowie and Bassett, 1989). The expression "Neogene–Quaternary," or its abbreviation "N/Q," which appears throughout this work to indicate the chronostratigraphic boundary defined by the base of the Pleistocene epoch, is therefore correct according to international guidelines. The philosophy and history behind the terminology suggest, however, that this arrangement is vulnerable and may be changed in the future.

It has been pointed out by Berggren and Van Couvering (1974, 1979) that the original mid-nineteenth-century meaning of "Neogene," which Moritz Hörnes (1853) coined for the marine fossils of the Vienna Basin, does not agree with the sense of the term as used by M. Gignoux (1913) in the first decade of this century. I am greatly indebted to Prof. F. Steininger (personal communication, 1991) for pointing out that Hörnes based his term on the "meiocän" and "pleiocän" local faunal units outlined by Bronn in 1838. Admittedly Bronn (1838) was extending to the Vienna Basin the terms "Miocene" and "Pliocene," which had just been introduced by Charles Lyell (1833), but Hörnes, writing in 1853, made it clear that it was Bronn's units, rather than Lyell's (contra Harland et al., 1990), that he considered to be basic to the Neogene. In so doing, Hörnes ignored Lyell's post-1838 revisions regarding the limits of his epochs, not to mention Beyrich's studies, which were soon to lead to the formal proposal of the Oligocene in 1855. Thus, the original Neogene, like the *original* Miocene and Pliocene on which Bronn based his units, encompassed the entire post-Eocene fossil record. Hörnes specifically included collections from the glacial "Löss-" and "Diluvial-Bildungen" in Austria. Furthermore, in correlating the Vienna Basin Neogene assemblages with Mediterranean faunas, Hörnes (1853, p. 809) mentioned collections from Rhodes, Crete, and Sicily that would now be dated to the Pleistocene.

The term "Neogene" was thereafter much neglected until Gignoux (1913) reintroduced it. In its revised form, it appears to have been nothing more than a casual literary convenience (Gignoux, 1955, p. 467), with no definition beyond the bald statement that it consisted of the Miocene and Pliocene epochs together. Despite the superficial resemblance of this diagnosis to the original, Gignoux's meaning for "Neogene" was actually quite different, because Hörnes did not recognize the Pleistocene as separate from the Pliocene, whereas Gignoux, writing 60 years later, took this separation for granted. Furthermore, although Hörnes, like Lyell, defined his unit in terms of the fossil record, unlike Lyell he seems to have ignored its application to geologic time, as is suggested by his choice of "-gen" rather than "-cen" as an ending. Gignoux made no mention of Hörnes's faunal definition (indeed, he made no mention of Hörnes at all) in treating the Neogene as a compound of epochs, or chronostratigraphic units. In that rather careless way the Pleistocene (and Oligocene, for that matter) was excluded from the Neogene without discussion, and it was in this "modern" sense that the term was used by IGCP-41 and eventually by the IUGS. Taking all this into consideration, Berggren and Van Couvering (1974, 1979) contended that the Neogene, in its original sense, was a biochronological unit irrelevant to the time–rock unit of the Quaternary, and chronologically overlapping it.

No post-Neogene unit is specified in the definition of Neogene given by Gignoux, let alone by Hörnes. On the other hand, the most appropriate term for the unit preceding the Quaternary is, of course, "Tertiary" – and if the one is valid it is difficult to see why the other is not. Indeed, the 1948 commission (King and Oakley, 1950) stated that it was the Tertiary–Quaternary boundary which would be identified with the Pliocene–Pleistocene boundary in its recommendation.

Gignoux's revised concept has been accepted by those who advocate that "Tertiary" should be dispensed with, and that "Neogene" and "Paleogene" should be used in its place. Harland et al. (1990) proposed to effect this change by elevating both "Tertiary" and "Quaternary" to the functionless rank of sub-era, where both antique concepts could be expected to fade away. To follow the Neogene, they introduced "Pleistogene" in place of "Quaternary" as the final system/period of the Cenozoic. On the other hand, as noted earlier, the approach of the IUGS (Cowie and Bassett, 1989) is to leave "Quaternary" as is, sequential to the Neogene, although it makes a somewhat inappropriate name for a third (alas, not fourth!) division of the Cenozoic.

A third proposal for the name of a unit to follow the Neogene (*sensu* Gignoux), with more historical and aesthetic merit than either of the preceding two, is the term "Anthropogene," as described in K. V. Nikiforova's Foreword to this volume. Though it originated as a conceptual definition of a geological interval, it is no different from Paleogene and Neogene in that regard, and in the end, as a chronostratigraphic unit, it would be defined by the base of the Pleistocene, just as the equivalent Pleistogene or Quaternary would.

New geochronology at the Pleistocene boundary

Upper limit of the Olduvai subchron. Pasini and Colalongo (Chapter 2, this volume) detail a minor but important change in the position of the Pleistocene boundary-stratotype at Vrica with regard to the conventional paleomagnetic model. The age of the top of the Olduvai normal-polarity subchron has been difficult to fix, because of a mixed-polarity "flutter" in the transition (Hilgen, 1991). Analysis of closely spaced samples by Zijderveld et al. (1991) in the Vrica section showed that a thin normal-polarity zone is present above marker *e*, which in turn is slightly above the reversal which was identified as the "top of the Olduvai" by Tauxe et al. (1983) and Nakagawa et al. (Chapter 3, this volume; Pasini and Colalongo, Chapter 2, this volume). This brief normal-polarity interval is not to be confused with the Cobb Mountain event (Turrin et al., 1994) or with other short normal-polarity excursions in the upper Matuyama chronozone which have been noted by various workers (Azzaroli et al., Chapter 11, this volume), including the N3 unit above marker *s* at Vrica (Tauxe et al., 1983).

With this new analysis, a slight modification is required for paleomagnetic characterization of the Pleistocene boundary-stratotype at Vrica. Although evidence for a complex structure in the upper Olduvai has been noted in a few instances (Zijderveld et al., 1991, p. 711), most published paleomagnetic profiles that show the Olduvai subchronozone have been in more condensed (or less closely sampled) sections, in which the short reversed section that contains marker *e* at Vrica was not resolved. Thus, as Zijderveld et al. (1991) reasoned, the thin normal-polarity zone above the main zone has always been included in the Olduvai wherever this upper zone has been sampled, and to consider it now as a separate subchron is not useful or practical. In view of this, the Vrica boundary should now be described as being "just below" rather than "just above" the top of the Olduvai. The time difference is not great (Table 1), and all of the associated biochronological markers cited in this work remain valid (Pasini and Colalongo, Chapter 2, this volume).

Table 1. *Comparison of the paleomagnetic time scale in use 10 years ago and the current, orbitally tuned time scale.*

Magnetostratigraphic boundary	Age, Ma (1985)	Age, Ma (1995)
Matuyama-Brunhes	0.73	0.780
Jaramillo top	0.90	0.990
Jaramillo base	0.97	1.070
*Cobb Mt. (Gilsá)	–	1.186
Olduvai top	1.65	1.770
Vrica zone β top	–	1.785
→ *P/P boundary*	*1.65*	*1.796*
Vrica zone β base	(1.65)	1.815
Olduvai base	1.82	1.950
Réunion top	1.99	2.140
Réunion base	2.02	2.150
Gauss-Matuyama	2.48	2.581
Kaena top	2.92	3.040
Kaena base	2.99	3.110
Mammoth top	3.08	3.220
Mammoth base	3.18	3.330

Note: The earlier time scale (Berggren, Kent, and Van Couvering, 1985) gives values that average 6% older than the orbitally tuned time scale (Cande and Kent, 1995) at each point. The Cobb Mountain subchron is dated according to Turrin et al. (1994). The age limits for the Vrica interval β were calculated from the upper and lower age limits of the Olduvai subchron, assuming a constant sedimentation rate in the Vrica sequence (Pasini and Colalongo, Chapter 2, this volume). In 1985 (column 1) the accepted age of 1.65 Ma for the top of the Olduvai subchron was erroneously applied to the top of the normal zone below interval β at Vrica.

Astronomically tuned time scale. Mathematical models of long-term variations in global insolation values, as predicted on the basis of harmonics in the obliquity of the earth's axis and the ellipticity and precession of its orbit, have been refined with the aid of modern computers (Berger and Loutre, 1988). This has brought fresh success in calibrating the proxy records of astronomically forced climatic periodicities to the magnetostratigraphy and biostratigraphy in marine deposits (Hilgen and Langereis, 1989; Shackleton, Berger, and Peltier, 1990; Shackleton et al., 1995a). The application of the new astronomically calibrated time scale to the Vrica section is discussed by Pasini and Colalongo (Chapter 2, this volume), but has not been consistently applied in the descriptions of other Plio–Pleistocene sequences in this volume. Table 1 is a reference to the "old" versus "new" values, as an aid to the reader.

Standard global Plio–Pleistocene chronostratigraphic units

The adoption of a stratigraphically defined lower boundary of the Pleistocene opens the possibility for an improved worldwide standard chronostratigraphy for the Plio–Pleistocene interval. Global standard stages have recently been established or proposed, based on the same marine sequences in southern Italy that comprise the type area of the Pleistocene. Figure 1 summarizes the consensus from a recent meeting of Italian stratigraphers (Van Couvering, 1995).

Calabrian versus Selinuntian. For some time now, the lowermost stage of the Pleistocene has been the subject of controversy. The 1948 resolution (King and Oakley, 1950) invoked a basic principle of chronostratigraphy in advocating that "the Pliocene–Pleistocene boundary should be based on changes in marine faunas," and to that end recommended that "the lower Pleistocene should include as its basal member in the type area the Calabrian formation (marine)" and that "according to evidence given this usage would place the boundary at the horizon of the first indication of climatic deterioration in the Italian Neogene succession" (King and Oakley, 1950, p. 214).

Gignoux did not specify a type section for the Calabrian. According to correspondence seen by G. B. Vai (personal communication, 1994), he purposely declined to do so on the grounds that it was impossible to consider any single exposure to be an adequate example of the complete "cycle." In order to convert the Calabrian to a standard chronostratigraphic unit, Selli (1971) selected one of Gignoux's described sections, Santa Maria di Catanzaro, as the stage stratotype, but this only proved Gignoux's point. Ruggieri and Sprovieri (1977), as discussed by Rio, Raffi, and Backman (Chapter 5, this volume) and Azzaroli et al. (Chapter 11, this volume), demonstrated that the Santa Maria di Catanzaro section represents only the uppermost part of Gignoux's (and the 1948 subcommission's) concept of the Calabrian and is stratigraphically far above levels where it might be expected that the evidence for the end-Pliocene climate change would be found. Furthermore, micropaleontological

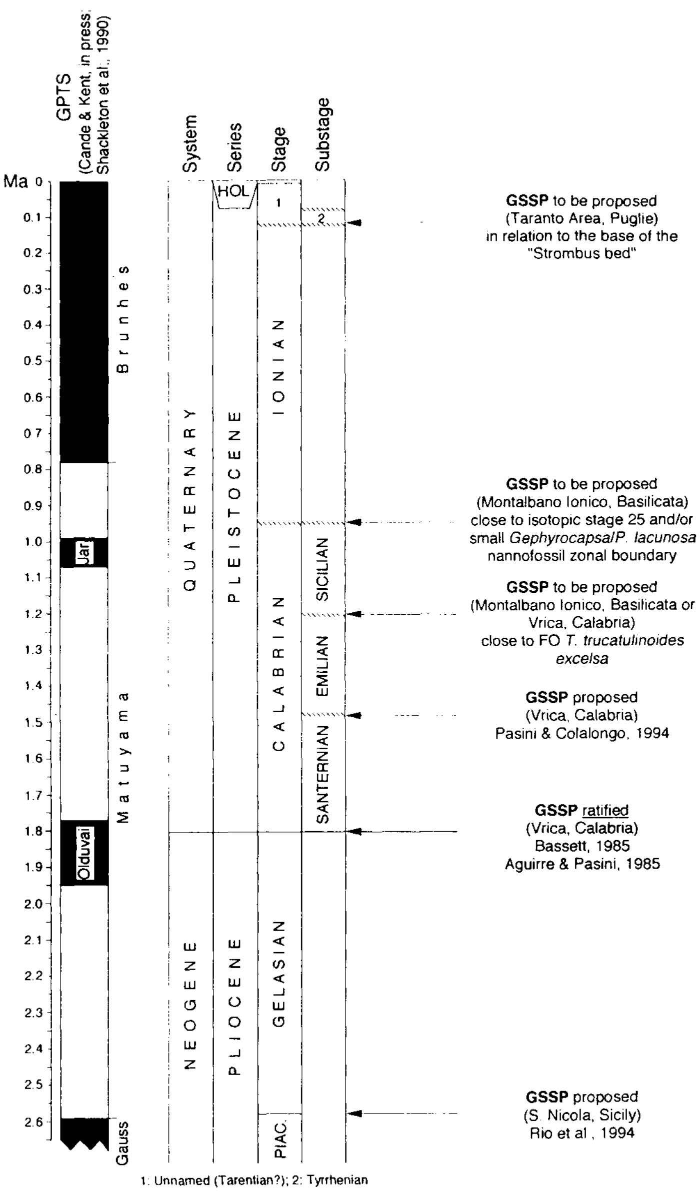

Figure 1. Chronostratigraphic units in the Plio–Pleistocene sequence of the Gulf of Taranto, southern Italy. Locations for GSSPs at the bases of the newly proposed Gelasian and Ionian stages and those of the revalidated Calabrian and its substages are shown according to the consensus at the 1994 Bari workshop (Van Couvering, 1995).

correlations have indicated that this section is in fact equivalent to strata in Sicily that had been regarded as typical of the *piano siciliano* of Doederlein since the nineteenth century (Ruggieri and Sprovieri, 1976; Rio et al., Chapter 5, this volume). Because Gignoux (1913) considered that the Sicilian "cycle" followed that of the Calabrian, the validity of the Santa Maria di Catanzaro stratotype came into question, and thus, by definition, the validity of the term "Calabrian" itself (Ruggieri and Sprovieri, 1977, 1979; Ruggieri, Rio, and Sprovieri, 1984).

The approach of IGCP-41 was to defer the problem presented by the stratotype of the Calabrian Stage and to follow the spirit, rather than the letter, of the 1948 IGC resolution in proposing to locate the Pliocene–Pleistocene boundary at Vrica. The strata in the Vrica section are in the same sub-Apennine formation that is partially exposed at Santa Maria di Catanzaro and Le Castella (Haq, Berggren, and Van Couvering, 1977) to the west, and although Vrica was not mentioned by Gignoux, it clearly demonstrates the concept of a change from warm- to cold-climate deposition at the end of the Pliocene, which was the basis of his Calabrian Stage.

In consideration of their view that the Calabrian was an "invalid junior synonym," Ruggieri and Sprovieri (1979) proposed the creation of two new pre-Sicilian stages/ages for the lowest Pleistocene, the Santernian and Emilian, based on sections in the Santerno River basin near Bologna. Those stages, together with the Sicilian, were considered to be components of a new Lower Pleistocene superstage, the Selinuntian, in place of the Calabrian *sensu largo*. Later, Ruggieri et al. (1984) revised the status of the Selinuntian, defining it as a single, composite stage/age made up of the Santernian, Emilian, and Sicilian at substage/chronozone rank. This position was also taken by Rio et al. (Chapter 5, this volume) and Azzaroli et al. (Chapter 11, this volume). In effect, the Selinuntian defined in this way occupies the same interval, and carries the same meaning, as the historical concept of Calabrian.

The question is not yet fully resolved, but many Italian stratigraphers involved in the debate now favor a return to the historical usage (Van Couvering, 1995) in which the status of the Calabrian is restored as the lowest stage of the Pleistocene in Italy, by recognizing that its lower boundary-stratotype is established at Vrica. One primary reason is the fact that, according to international guidelines (Salvador, 1994), this recognition may be merely a formality. Two points must be kept in mind: that chronostratigraphic units are fundamentally hierarchical, and that a boundary-stratotype may be defined separately from a body-stratotype. Few stages have such explicit hierarchical status as the Calabrian, which was specified as the basal unit of the Pleistocene in the same international resolution that justified the proposals made by IGCP-41. We must therefore agree, according to the concept of hierarchy outlined in the international guidelines, that any definition of the base of the Pleistocene Series cannot exist apart from the base of the Calabrian Stage. The basal limit of the Calabrian Stage would thus have been removed axiomatically from the base of the body-stratotype at Santa Maria di Catanzaro to a boundary-stratotype at the level of the claystone overlying marker *e* at Vrica, by the act of proposing this level as the definition for the base of the Pleistocene Series. It should be noted that the problem of synchronicity or "priority" between the body-stratotype of the Calabrian at Santa Maria di Catanzaro and the typical Sicilian at Ficarazzi disappears when the Sicilian is considered as a substage, as proposed by Ruggieri et al. (1984).

New stages and boundary proposals. Recent advances in micropaleontology, magnetochronology, and stable-isotope cyclostratigraphy in the upper Cenzoic marine strata in southern Italy support the recognition of new Pliocene and Pleistocene global chronostratigraphic units in the same context as the Pleistocene boundary-stratotype. Rio, Sprovieri, and Thunell (1991) have conclusively demonstrated that in the upper Pliocene, the stratotype of the Piacenzian Stage is erosionally truncated near the Gauss–Matuyama boundary, some 0.8 m.y. prior to the base of the Pleistocene as presently recognized. Under the international guidelines (Salvador, 1994), chronostratigraphic "gaps" such as that are eliminated by considering the top of a unit to be defined by the base of the next succeeding unit, so that the Piacenzian Stage could be considered to continue up to the base of the lowest stage in the Pleistocene. (An alternative proposal, to move the base of the Pleistocene *downward* to fill the gap, is not in accord with the guidelines, as discussed later). Rio, Sprovieri, and Di Stefano (1994) pointed out that the paleobiological and paleoclimatic characteristics of the *gap* in question, as seen in the well-studied section at Monte San Nicola in Sicily, distinguish it from the typical Piancenzian. Those authors proposed a new stage, the Gelasian, to embody the Upper Pliocene and to more clearly document the sharp change from warm-climate to cold-climate conditions at the level of the Vrica boundary.

The Ionian Stage has been proposed for the Middle Pleistocene by a group led by Neri Ciaranfi at Bari University (Van Couvering, 1995), to be based on the upper part of a thick sequence of highly fossiliferous Lower and Middle Pleistocene marine clayey silts at Montalbano Ionico in southern Basilicata (Ciaranfi et al., 1994). Preliminary studies of the sequence have shown continuous sedimentation from the middle Matuyama to the middle Brunhes, in the context of calcareous nannoplankton zones of "large" *Gephyrocapsa,* "small" *Gephyrocapsa,* and *Pseudoemiliana lacunosa* (Rio, Raffi, and Villa, 1990). In order to avoid the problems raised by the fact that cold-climate lowstands are erosional in shallow-marine and continental deposits, the boundary-stratotype of the Ionian has been proposed at the level of the last warm, transgressive event prior to the Menapian "glacial Pleistocene," at the 0.9-Ma paleoclimatic step, and the end of the Villafranchian (Azzaroli et al., Chapter 11, this volume). This is isotope stage 25, just above the Jaramillo and correlative to the base of the *P. lacunosa* zone (Castradori, 1993), which is well documented at Montalbano Ionico.

Boundary-stratotypes for the Santernian, Emilian, and Sicilian, which have been designated as Lower Pleistocene substages (Ruggieri et al., 1984), have been proposed in southern Italy (Figure 1). The base of the Santernian substage is identified, by hierarchic necessity, with that of the Calabrian Stage and that of the Pleistocene Series at the Vrica GSSP. The boundary-stratotype of the Emilian has been defined by Pasini and Colalongo (1994) at a point 71 m stratigraphically above the Pleistocene boundary in the Vrica section, at the level of *Hyalinea baltica* FAD and the base of the "large" *Gephyrocapsa* zone. A boundary-stratotype for the Sicilian has not been designated, but Ruggieri et al. (1984) suggested the first appearance level of *Globorotalia truncatulinoides excelsa* FAD at Ficarazzi (Sicily), which is near the base of the "small" *Gephyrocapsa* zone.

Lowering the Quaternary

Years of effort have culminated in the establishment of an internationally accepted Pleistocene boundary-stratotype coincident with a major climatic downturn. This level, near the top of the Olduvai subchron, with an age of about 1.8 Ma, is (by definition) the Quaternary boundary as well. It is now well known that the shift from the equable, stable climates of the early Pliocene to the intensely seasonal and highly cyclic climates of the late Pleistocene and Recent proceeded episodically, with progressive step-like increments, from the middle Miocene climatic maximum to modern "fully glacial" conditions by about 0.4 Ma (Thunell and Williams, 1983). Major steps toward the present glacial-climate condition have been confirmed in high-quality deep-sea records at 3.2 Ma and 2.5 Ma (Shackleton, Hall, and Pate, 1995b), prior to the 1.8-Ma downturn, and other major steps at 0.9 Ma and 0.4 Ma (Shackleton et al., 1990; Hilgen, 1991). A detailed review is beyond the scope of this brief commentary, but it is safe to say that whereas a given step or two may be strikingly distinct in one or more of the proxy records that have been developed – stable isotopes, vertebrate fossils, land microflora and macroflora, marine planktic and benthic microfossils, marine and continental epiglacial sediments – no one climatic downturn stands out from all the rest on a consensus basis. Because the step at 1.8 Ma is not universally dominant, the effects of the two most closely bracketing steps have also been advocated as criteria for the Pliocene–Pleistocene boundary.

For example, Nikiforova describes, in the Foreword to this volume, how researchers in eastern Europe and Russia have long held (with some justification) that the establishment of lowland glaciers on the European continent at about 0.9 Ma (i.e., isotope stage 24) (Hilgen, 1991) marked the proper beginning of the Pleistocene, in accord with the original intentions of nineteenth-century stratigraphers. Itihara et al. (Chapter 24, this volume) show that Japanese stratigraphers have also dated the base of the Pleistocene to that time, coincident with the extinction of typical Pliocene macroflora.

On the other hand, the glacial maximum at about 2.5 Ma (i.e., isotope stage 100) (Shackleton et al., 1995b) appears to have been the first to have noticeably impacted the continental and shallow-marine environments, and it was also the earliest cold-climate phase in which the limits of ice-rafted debris (IRD) expanded beyond the Arctic Ocean and into the northern Atlantic and northern Pacific. Workers in China, as noted by Zhang (Chapter 26, this volume), were accustomed to using the earliest loess, together with evidence for the development of the Villafranchian mammal fauna, as indicators that the Pleistocene epoch began at levels now dated to 2.5 Ma. Likewise, evidence has been cited for distinct changes in continental mammal faunas at that time in western Europe (Azzaroli et al., Chapter 11, this volume), North America (Lindsay, Chapter 30, this volume), and Africa (Cooke, Chapter 27, this volume). Although the next major cold-climate period, at about 1.8 Ma, was more intense, its effects were less dramatic, in the context of a biosphere that had already been "winterized" to a significant extent.

Historically, as Berggren and Van Couvering (1979) described, agreement on the Pliocene–Pleistocene boundary was a vexatious problem for many years, because researchers in different fields, like the blind men examining the elephant, were unable to reconcile their separate understandings. The physical reference point, or "golden spike," adopted by the 1948 London IGC was a logical and practical solution, with the goal of shifting the argument from the debating room to the outcrop. Nevertheless, the ideology of the ice ages continues to exert a fascination. The primary example is opposition to the Vrica definition (e.g., Zagwijn, 1992; Sibrava, 1992), on the ground that the cold-climate maximum in the 2.5-Ma step, being of greater extent and severity than those known earlier in Norway and Iceland, was the true "first glacial." It is worth noting, in this regard, that the extent of ice-rafting and the latitudinal distribution of indicator biota suggest that at its coldest, the 2.5-Ma "glacial" was warmer than the present-day interglacial. Furthermore, between 2.5 and 1.8 Ma, the Tiglian phase (approximately equivalent to the marine Gelasian Stage) could be characterized as "the last preglacial," because it was characterized by global climates that were significantly warmer and more equable than those in any Pleistocene cycle that followed. The conundrum faced by the climatically correct is thus to decide whether it offends more that the Pliocene should include an interval of relatively cold climate if the Pliocene–Pleistocene boundary is at 1.8 Ma, or that the Pleistocene should include a clearly pre-glacial interval if the boundary is at 2.5 Ma.

That such considerations should be brought forward as reason to question the 1948 resolution and to reopen the boundary wars is, however, quite another matter. There is no necessary relationship between chronostratigraphy and past climate (Salvador, 1994), even though the identity of the Pleistocene epoch has been gravely debated as if it were a paleoclimatic unit ever since Edward Forbes (1846) noticed that it was also a time of "ice ages." The truth is that whereas it is pleasing, and even logical, to make an epoch boundary coincident with a notable geohistorical event such as a change in global climate, it cannot be made a requirement in chronostratigraphy. After three decades of conscientious and painstaking study, an international agreement has been reached to the effect that the intent of the 1948 resolution has been fulfilled with the selection of a physical boundary-stratotype at Vrica. It is our hope that this volume will demonstrate that in addition to being legitimate, the Vrica boundary-stratotype is in fact as practical and appropriate as any other.

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Foreword

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Quaternary deposits are everywhere. In mountainous areas they range from thin soils and scree through ephemeral accumulations, such as talus fans and stream alluvium, to sometimes quite long-lived deposits like peat, valley fills, lake beds, and glacial moraines. The acidic ground water and soils of mountains, however, leave few fossils, apart from pollen and macrofloral impressions. In lowland river valleys and coastal margins, continental Quaternary deposits are thicker, more widespread, and more fossiliferous. Even so, they are lithologically and stratigraphically discontinuous on the larger scale. Only in the marine environment, particularly on the floors of the deep ocean basins, are essentially uninterrupted and highly fossiliferous sequences of Quaternary strata found over wide regions. Because the boundary between land and sea has changed very little, relatively, during the geologically brief period of the Quaternary, the early, classic studies of this crucially important interval in the earth's history were restricted to the relatively discontinuous continental deposits. The number of studies of marine Quaternary deposits has increased greatly during the past few decades, however, and they have provided us with a new and much better documented basis from which to establish the position of the lower boundary of the Quaternary.

As far back as 1760, the Italian geologist Arduino singled out Quaternary strata as a distinct group of geological deposits characterized by lack of induration. The concept of a Quaternary time, as the most recent part of geological history, was suggested in 1825 by Desnoyers, and in 1839 Charles Lyell introduced a new term, "Pleistocene," for the epoch of youngest marine faunas. By the middle of the nineteenth century it was evident that much of the Quaternary was characterized by ancient glacial deposits; at the same time, the distinct paleofaunal change from Pliocene to Pleistocene was understood to be caused by the increased influence of glacial environments. Thus, "Pleistocene" and "Quaternary" came to be considered as coeval terms for the glacially dominated period that ended the Cenozoic.

In 1846, Forbes recommended that use of "Pleistocene" be restricted to exclude the most recent post-glacial strata, which he denoted by the term "Holocene." Eventually the term "Pleistocene" came to be used by most stratigraphers in that way, as a name for the deposits of the "Ice Age," with an upper limit equivalent to the ending of glacial conditions in temperate lowlands, presently dated to about 12,000 years ago.

The lower limit of the Quaternary is still subject to dispute, and some researchers do not even acknowledge its existence, because of the opinion that it is based on the same antiquated preconceptions as the Tertiary, not to speak of the Primary and Secondary. These workers consider the Cenozoic era to be directly divided into constituent epochs, such as the Pleistocene, or that the Pleistocene epoch is included in an extended Neogene sub-era. A second group (including some of the same researchers) would also eliminate the Holocene as a separate, independent epoch. The work of IGCP-41, on the other hand, is based on the prevailing opinion that the Quaternary should be considered an independent period of the Cenozoic and that its lower limit, the boundary between the Neogene and the Quaternary (the N/Q boundary), is equivalent to the boundary between the Pliocene and the Pleistocene.

In Russian, the terms "Post-Tertiary" and "Post-Pliocene," as synonyms for the Quaternary, were widely used until the 1920s. In 1919, A. P. Pavlov proposed to replace the name "Quaternary" with "Anthropogene," according to the view that the main event in the organic history of the final system/period of the Cenozoic was the appearance of humans. In actual fact, as E. Aguirre points out in this volume (Chapter 10), the evolution of *Homo erectus* in Africa at about 1.8 Ma is almost coincident with the age of the N/Q boundary at Vrica, as recommended by IGCP-41. The name "Anthropogene" accorded well with the preceding "Paleogene" and "Neogene," which had already replaced "Tertiary" in the Soviet schema, and it is now commonly used throughout central and eastern Eurasia. In 1963 the names "Quaternary" and "Anthropogene" were officially approved as equivalents by the Interdepartmental Stratigraphic Committee of the USSR, in the MSC (modern stratigraphic code) (Zhamoida et al., 1977). The term "Anthropogene" has an obvious priority over the recently proposed "Pleistogene" of Harland and others (Van Couvering, Preface, this volume).

Also, in consideration of priority, one should remember that "Pleistocene" was long applied in the Soviet Union according to a strict interpretation of Forbes's definition, which restricts the term to the time of fully glacial conditions. It is now known that in mainland North America, continental ice sheets appeared during the climatic deterioration at the beginning of the presently defined Pleistocene, and even before that in Greenland and Iceland, whereas in subpolar Eurasia (to which Forbes was referring) the first fully glacial conditions are identified with the Menapian (classic Mindel) paleoclimatic unit, now dated as being of early Middle Pleistocene age. The pre-Menapian deposits corresponding to the Lower Pleistocene of western Europe are therefore equivalent to the Eopleistocene ("dawn of Pleistocene") in the former USSR.

The definition of boundaries is a major issue in the debates on Quaternary geology. The first step toward fixing the place of the boundary between the Pliocene and the Pleistocene was the organization of a special temporary commission at the XVIII International Geological Congress in London in 1948. The commission, affirming that the base of the Quaternary was equivalent to the base of the Pleistocene, outlined three basic criteria for the placement of the boundary: (1) the boundary should be based on a faunal change in a section of marine deposits, (2) the boundary should be located in the classic territory of Quaternary marine deposits of Italy, the area in which these principles can be best applied, and (3) the boundary should be placed at a horizon demonstrating the first indication of a deterioration of climate in the Italian deposits, as evidenced by the first appearance of (unspecified) "northern guests" in the Mediterranean Sea. Based on those criteria, the commission recommended that the lower boundary of the Pleistocene should be drawn at the base of the Calabrian Stage in Italy. That recommendation was approved and adopted in the closing session of the XVIII International Geological Congress.

The commission was aware of the fact that in proposing the Calabrian Stage, Gignoux (1913) had referred to a number of sections throughout Italy that evidenced a change from warm, Pliocene conditions in the Mediterranean to a significantly colder climate. That environmental change was documented by the first appearance in Italian sections of animals that clearly originated in the North Atlantic bioprovince. The first species identified as a "northern guest," and still the most popular example, is the mollusk *Cyprina islandica*, now generally referred to the genus *Arctica*, which today lives in shallow subarctic waters above the 60-m isobath. Its value for close correlation is limited, and stratigraphers therefore put forward a boreal benthonic foraminifer also found in the Calabrian, *Hyalinea baltica*, as a deep-water indicator for the base of the Pleistocene (and also Quaternary).

Evidence of glacial climates can now be documented by characteristic changes in many different assemblages of microorganisms and in the composition of fossil-spore spectra. Originally it was assumed that the earth had been free from lowland ice sheets until it suddenly entered a worldwide ice age in the last part of the Cenozoic. With the aid of modern studies of microfossil groups, however, we can recognize diachronous and recurring appearances of cold-climate conditions in the Mediterranean Sea. Application of this type of data to the history of the late Cenozoic confirms that the deterioration in climate throughout the Neogene was in fact a succession of gradually intensifying global cycles, with repeated coolings having taken place in the Mediterranean long before the first appearance of the truly boreal elements that are designated as indicative of the beginning of the Quaternary. In addition, Italian geologists have recently shown that the appearances of the two most often cited "cold guests," *Arctica islandica* and *Hyalinea baltica*, were not synchronous and that *H. baltica* is certain to have appeared later than *A. islandica*.

Among the typical Calabrian sections, Gignoux (1913) included Santa Maria di Catanzaro (Calabria), and that section was later designated by Selli (1971) as the neostratotype. As far back as 1961, Ruggieri proposed a new stratigraphic zonation of the Italian Pliocene and early Pleistocene, with two zones: a Lower Pliocene zone with two subzones (A and B), assigned to the early Pliocene, and an upper zone also divided into two subzones (C, with *A. islandica*, and D, with *A. islandica* plus *Hyalinea baltica*). Ruggieri considered subzone D to be the base of the Pleistocene because it coincided with the first appearance of *H. baltica*, with subzone C representing the late Pliocene. Later on, Ruggieri and Sprovieri (1976) recognized that subzone C, with its famous "cold guest," was actually equivalent to the lowermost Pleistocene, corresponding with the accepted general concept of the Calabrian Stage. Subzone D was coeval with the beginning of the Emilian Stage, and the overlying Sicilian Stage was characterized by the two species named earlier plus *Globorotalia truncatulinoides excelsa*.

As is now well known, the same authors (Ruggieri and Sprovieri, 1977) later felt it necessary to abandon the name "Calabrian" for the lowermost Pleistocene stage when it was discovered that the Calabrian of Santa Maria di Catanzaro appeared to be synchronous with the typical Sicilian, and the Calabrian Stage defined in that section by Selli (1971) was, in their view, therefore an invalid "synonym" of the type Sicilian proposed by Doederlein in 1872 in a section at Ficarazzi (Palermo). In place of the Calabrian, Ruggieri and Sprovieri (1977) proposed a new stage, the Santernian, defined in the Santerno Valley tributary to the Po in northern Italy.

At the IGCP/Plio-41 symposium in Italy in 1975, R. Selli and G. Pasini proposed the section at Vrica, 4 km south of Crotone (Calabria), as a candidate to serve as the boundary-stratotype of the N/Q or Pliocene–Pleistocene boundary. Study of this section by the working group of IGCP Project 41 and INQUA Subcommission 1-d, "Pliocene–Pleistocene Boundary" (Selli et al., 1977), showed that the section met all the requirements of the modern stratigraphic codes for such a boundary (Hedberg, 1976; Zhamoida et al., 1977). Further investigations have only strengthened the acceptance of the Vrica section as a location for the Pliocene–Pleistocene boundary-stratotype, or neostratotype according to the Soviet rules of stratigraphic nomenclature (Zhamoida et al., 1977). Detailed descriptions of the section are given in publications of Aguirre and Pasini (1985) and Nikiforova (1985), as well as in various chapters in this volume.

The three stages Santernian, Emilian, and Sicilian were united by Ruggieri and Sprovieri (1979) into one superstage, Selinuntian, although the short durations of those units prompted those authors subsequently to reconsider them as substages or chronozones of a Selinuntian Stage (Ruggieri, Rio, and Sprovieri, 1984). Most recently there has been general agreement (Preface, this volume) to restore the Calabrian as the lowest stage in the Pleistocene, according to the historical argument that the boundary-stratotype for this stage was automatically established at the same lithologic point as the definition of the Pliocene–Pleistocene boundary-stratotype.

Thus, the section at Vrica proposed by Selli and Pasini in 1975, and thoroughly studied afterward, meets all the international requirements for adequate definition of the boundary-stratotype. If we understand the restored Calabrian to embody the original Calabrian concept of Gignoux, then the Vrica proposal also meets every condition set by the 1948 London IGC for definition of the Pliocene–Pleistocene boundary.

It should be noted in conclusion that the most urgent task now facing the INQUA Commission on Stratigraphy is standardization of Quaternary stratigraphical and geochronological units and their terminology, as concerns both the name of the system itself and its subdivisions. As the stratigraphic range of the Quaternary does not exceed one normal biozone, that of *Globorotalia truncatulinoides*, recognition of series, stages, and chronozones within the Quaternary may not be justified. In the Stratigraphic Code of the USSR (Zhamoida et al., 1977), subdivisions even finer than zones are described, some of which (division, link) have been included in the standard stratigraphic scale and may be appropriate as Quaternary subunits. These problems are not proper subjects for this volume, of course, but are considerations that arise now that the work on IGCP Project 41 is completed.

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Part I

Definition of the base of the Quaternary

1 International Geological Correlation Program, Project 41: "Neogene/Quaternary Boundary"

KSENIA V. NIKIFOROVA and MIKHAIL N. ALEKSEEV

Formation of the project

Research on Project 41, "Neogene/Quaternary Boundary," was initiated in 1974 under the auspices of the International Geological Correlation Program at the second session of the IGCP board held in Vienna. At the same session, a separate proposal by the Geological Survey of India to define this boundary in the Indo-Pakistani subcontinent was included in IGCP Project 41 (Sastry, Chapter 23, this volume).

The first session of the working group for Project 41 was held in Barcelona in September 1974. In attendance were E. Aguirre (Spain), A. Azzaroli (Italy), M. Alekseev (USSR), W. A. Berggren (USA), H. B. S. Cooke (Canada), L. K. Gabunia (USSR), C. Ghenea (Romania), J. Michaux (France), K. V. Nikiforova (USSR), A. Rónai (Hungary), M. V. A. Sastry (India), R. Selli (Italy), K. O. Lange (UNESCO), and R. Gonzales (Spain). At the conclusion of that meeting, it was agreed that the membership of the Project 41 working group would be as follows:

K. V. Nikiforova (chairman), Geological Institute, USSR [now Russian] Academy of Sciences, Moscow, Russia
E. Aguirre (treasurer), Ciudad Universitaria, Madrid, Spain
M. Alekseev (secretary), Geological Institute, USSR [now Russian] Academy of Sciences, Moscow, Russia
A. Azzaroli, Museum of Geology and Paleontology, Florence, Italy
L. Benda, Niedersachsisches Landesamt für Bodenforschung, Hannover, [West] Germany
W. A. Berggren, Woods Hole Oceanographic Institution, Woods Hole, MA, USA
H. B. S. Cooke, Dalhousie University, Halifax, N.S., Canada
L. K. Gabunya, Institute of Paleobiology, Academy of Sciences of the Georgian SSR [now Georgian Academy of Sciences], Tbilisi, Georgia
E. D. Gill, University of New South Wales, Canterbury, Australia
C. Ghenea, Geological Institute, Bucharest, Romania
H. M. S. Hartono, Geological Research and Development Centre, Bandung, Indonesia
R. W. Hey, Cambridge University, Cambridge, UK
J. C. Ingle, Stanford University, Stanford, CA, USA
M. Itihara, Osaka City University, Osaka, Japan
H.-D. Kahlke, Institute of Quaternary Paleontology, Weimar, [East] Germany
J. Michaux, University of Montpellier, Montpellier, France
H. Nakagawa, Tohoku University, Sendai, Japan
R. Pascual, National University, La Plata, Argentina
K. Prasad, Geological Survey of India, Bangalore, India
A. Psilovikos, Aristotle University of Thessaloniki, Greece
A. A. Rónai, Hungarian Geological Institute, Budapest, Hungary
M. V. A. Sastry, Geological Survey of India, Calcutta, India
R. Selli, Institute of Geology, Bologna, Italy
N. J. Shackleton, Cambridge University, Cambridge, UK
P. Vella, Victoria University, Wellington, New Zealand

The members of the working group agreed that the main task of Project 41 would be detailed studies of the stratigraphy of Pliocene and Lower Quaternary deposits developed on continents, islands, and ocean floors, aimed at definition and worldwide recognition of the Neogene–Quaternary boundary. Establishment of that boundary would be of great importance for the universal stratigraphic scale and for strengthening the standardized basis for mineral prospecting, geological mapping, map legends, neotectonics, geomorphology, and so forth.

In accordance with the adopted program, the group agreed that various methods should be used for solving this problem: biostratigraphic, climatostratigraphic, and physical methods such as paleomagnetic, paleotemperature, and radiometric age determinations. Studies of marine and continental faunas and floras and of climatic changes at various intervals of the Pliocene and

Quaternary would provide the research data for paleogeographic reconstructions of certain time sections. Of particular importance would be the use of physical methods that could provide more precise dating, applicable to the problems of synchroneity and metasynchroneity of geological events on a regional or global scale.

Work on Project 41 would elucidate the geological history and structure of continental shelves that hold oil and gas prospects. Continental and near-shore deposits of Quaternary age include placers of precious metals and rare ores. Superficial deposits of tropical continents include large-scale oxide ore bodies and concentrations of heavy minerals.

The project was conceived as an interdisciplinary study. Specialists from various fields – geology, geomorphology, paleoclimatology, paleontology, micropaleontology, archeology, anthropology, chemistry, and physics – took part. Close cooperation with international and national organizations such as IUGS (International Union of Geological Sciences) and INQUA (International Quaternary Association) was deemed essential. In this respect the project conforms fully to the aims and goals of UNESCO (United Nations Educational, Scientific, and Cultural Organization).

Background

The position of the N/Q (Neogene–Quaternary) boundary had already been discussed at several meetings of the IUGS and INQUA, as well as in international symposia and colloquia convened especially to study this problem. The problem of defining the boundary was first debated at the XVIII International Geological Congress in London in 1948, which recommended that the deposits of the Calabrian Stage of Italy (in which the northern immigrants *Arctica islandica* and *Hyalinea baltica* appear for the first time) be defined as the basal member of the Lower Pleistocene in the Quaternary system. The Villafranchian was considered to be the terrestrial equivalent of the Calabrian. Later it was found that the term "Villafranchian" applied to a much wider time interval and that its lower two-thirds belongs to the Pliocene, if the Calabrian represents the oldest part of the Pleistocene.

At the international colloquium organized by the INQUA Subcommission 1-d, "Pliocene/Pleistocene Boundary," held in the USSR in 1972, the following recommendation was adopted: "the lowermost level in the section of Le Castella in Calabria with remains of *Hyalinea baltica* (Schrotter) [is] selected as the initial definition of the base of the Pleistocene in a marine section of the Mediterranean basin." The colloquium also noted the need for more exact correlation between marine and continental deposits and the establishment of stratigraphic analogues of the Calabrian in other territories, including sections of sea-floor sediments. The recommendation of the 1972 colloquium in the USSR served as the basis for the subsequent recommendations of the XXIV International Geological Congress in Montreal in 1972.

It must be noted here that in subsequent studies by Italian stratigraphers it has been shown that *H. baltica* unquestionably appears in the southern Italian sequences much later than *Arctica islandica*, at a level that has been identified with the base of the Emilian substage. The first appearance of *A. islandica* in Italian strata is still considered to be an accurate indication of the earliest part of the Pleistocene.

The next international colloquium on the Pliocene–Pleistocene boundary organized by INQUA Subcommission 1-d was held in Christchurch, New Zealand, in December 1973 during the IX INQUA Congress. That congress provided an important opportunity for firsthand examination of the Upper Pliocene and Lower Pleistocene deposits, as well as the position established for the N/Q boundary in New Zealand and Australia. Some of the papers read at that symposium were published in Bulletin 13 of the Royal Society of New Zealand, *Quaternary Studies*, in 1975.

Organization and objectives of Project 41

Plan of work

The basic goals of the international working group on IGCP Project 41, "Neogene/Quaternary Boundary," were as follows:

1. analysis of the available data and solution of approachable problems through investigations carried out by the national working groups, or in cooperation with scientists from other countries involved in these activities
2. field studies regarding new objectives and critical sequences proposed by national working groups
3. field investigations, consultations, and exchange of knowledge on the geology of Pliocene and Lower Pleistocene deposits in the developing countries in order to stimulate scientific work
4. interdisciplinary studies in regions of difficult access, with priority given to studies of abyssal oceanic sediments, in consultation with the INQUA Subcommission on Stratigraphy of Deep-Sea Sediments

The field-team concept

The field-team approach was adopted as the best method to study important local stratigraphy and correlation of Pliocene and Lower Quaternary deposits. The following are some of the particular sections studied by the national working groups:

Italy: stratotypical Calabrian and Villafranchian sections; review of the present state of investigations of the problem "N/Q Boundary" in the type area.

Japan: key sections of the Pliocene and Lower Quaternary deposits of the Akita, Kinki, and Kanto regions.

USSR: nearly complete sections of Pliocene and Lower Pleistocene deposits in the European part of the former USSR and in middle Asia.

United States: key sections of the Pliocene and Lower

Pleistocene marine and continental deposits of California and the interior areas.

India: field conference, study of Pliocene and Lower Pleistocene deposits of the Siwaliks.

Correlation

The program of stratigraphic studies was planned to correlate in time and space the Pliocene and Lower Quaternary deposits in various paleogeographical provinces of different continents and islands, and on the ocean floor, as follows:

Deposits in continents and islands. Elaboration of stratigraphy and correlation of deposits located in different paleogeographic provinces, as well as correlations between marine and continental deposits, would pose a problem. The stratigraphic sections in Italy, England, and The Netherlands had been the best studied in the regions of western, central, and northern Europe, although the sections in western and eastern Germany would also be of importance.

In eastern Europe, successions in the European part of the USSR, southern Moldavia, Ukraine, Azerbaijan, and Georgia had been well studied. The sections in Romania, Yugoslavia, and Hungary would be very important. The main problems in Europe appeared to involve correlation of marine and continental deposits, using a combination of methods such as radiometric and paleomagnetic analyses, and coordination with the sections of subaqueous and subaerial series.

In Asia, determination of the N/Q boundary would involve studies in the sections of Turkmenia, middle Asia, Turkey, Transuralian Russia, western and eastern Siberia, the northeastern USSR, the Far East, the Indian subcontinent, Burma, Indonesia, China, and the Japanese islands. The principal problems in those regions would also lie in the correlation of marine and continental deposits, as well as volcanogenic formations of boreal and tropical zones, and in the definitions of the nature of sedimentation, erosional phases, and sea-level fluctuations.

Africa would require study of the sections of continental fossil-bearing deposits of Pliocene and Lower Quaternary age in the African rift zone. The main task here would be to detail the chronostratigraphic correlations of the Rift Valley sections with the climatically influenced sections of the southern Mediterranean coast (Algeria, Tunisia, Morocco) and those of the northern coast of the Mediterranean.

In the case of North America, the most important objectives would be comparisons of the marine sediments of California and the Gulf of Mexico with the continental sediments. It would be necessary to correlate the older glacial deposits and to work out the detailed stratigraphic schemes based on the use of a number of methods.

In South America, objectives would include thorough studies of the stratigraphy of Pliocene and Lower Pleistocene deposits of the Pampas, radiometric ages for volcanogenic deposits, elaboration of the stratigraphy of Plio–Pleistocene marine sediments, and, above all, correlation of climatic fluctuations between the southern and northern tropical zones on the South American continent.

Australia and New Zealand would offer an opportunity for thorough study of the Pliocene and Lower Quaternary sections of marine, continental, and volcanogenic deposits, thus providing the data to work out a climatostratigraphic scheme for the deposits of those ages. A more precise definition of the positions of datum planes with regard to global chronostratigraphy would be the final objective.

Deposits of ocean basins. The objectives of studies in the ocean basins would be analyses of the precise stratification of abyssal deposits and establishment of the boundary of the *Globorotalia tosaensis* and *Globorotalia truncatulinoides* zones, as well as the stratigraphic positions of other oceanic fossils in the Pacific, Indian, and Atlantic oceans. A further major objective would be the zonation and correlation of sections of Upper Pliocene and Lower Pleistocene deposits at high, middle, and low latitudes. Establishment of stratigraphic analogues of the *G. tosaensis* and *G. truncatulinoides* zones, as well as nannoplankton and radiolarian zones typical of middle latitudes, would be required for deposits laid down in high latitudes. Contributing to that effort would be studies of the paleomagnetism, radiometric age, and paleotemperatures of such deposits.

History of the working group activities

During 1975, national and regional working groups were organized and began to implement research activities in the German Democratic Republic, the Federal Republic of Germany, India, Spain, Japan, Italy, the USA, and the USSR, and international groups began work on the magnetostratigraphy and stratigraphy of deep-sea Pliocene and Lower Quaternary deposits. They were joined by working groups in Australia and New Zealand in 1976.

The Italian national working group carried out extensive studies on the establishment of a stratotypical section in marine deposits in southern Italy, as well as research on Pliocene and Lower Quaternary continental deposits.

The symposium of the working group in Italy, 1975

The second symposium was held in Bologna, Italy, in October 1975, and the presented papers were published in the *Giornale di Geologia*, vol. 41 (1977). The meeting was attended by 47 representatives from 12 countries (United Kingdom, France, Federal Republic of Germany, The Netherlands, Hungary, Canada, Spain, Italy, Japan, Yugoslavia, the USA, and the USSR). Twenty-five contributions from various countries were delivered at the Bologna session, most of which were preliminary to the chapters in this volume.

The group spent six days making excursions to the sections of Pliocene and Quaternary deposits in central Italy (basins of the Santerno, Arno, and Tiber rivers) and in Calabria. The participants agreed that there were some difficulties in the published interpretation of the Santa Maria di Catanzaro sections, as well as in the Le Castella outcrop, with regard to locating a boundary-stratotype. It was noted that the Vrica outcrops of the Crotone area, not far north of Le Castella in Calabria, form a section about 300 m thick in which there are no apparent interruptions in sedimentation. It was already known that sedimentation at Vrica had taken place under conditions of rather deep water, as compared with those farther to the south, and that changes of both foraminiferal and ostracod complexes could be observed in the section (Pasini et al., 1975). The participants decided to study the Vrica section in detail and to submit the results at the session of the working group in 1977 at the X INQUA Congress in Birmingham. The working group members also suggested that by the end of 1977 the Vrica section would have been sufficiently well studied to enable them to recommend it as a potential stratotype of the N/Q boundary. Paleontological and radiometric studies of the Vrica section were to be carried out by the Italian national working group, and the paleomagnetic determinations by Japanese scientists.

To that end, the conference adopted a resolution to the effect that the working group of IGCP-41, "Neogene/Quaternary Boundary," had drawn the attention of the appropriate Italian authorities to the great importance of the geological sections in Calabria for the establishment of a time scale for use all over the world. Thick deposits at Vrica, near Crotone, were currently being investigated and promised to yield results of major importance to this purpose.

More detailed evaluations of the small and large mammalian markers, to establish a reliable subdivision of the continental late Pliocene and early Pleistocene, would be necessary. It was recommended that an international meeting of mammalogists be held in the western Mediterranean region to achieve a common understanding of the detailed mammalian biostratigraphic subdivisions that could be relevant to the aim of Neogene/Quaternary correlation.

The IGCP-41 working group symposium in Japan, 1976

The third IGCP-41 symposium took place as a joint meeting of the IGCP-41 working group and the IUGS working group on the Pliocene–Pleistocene boundary in May 1976, during the First International Congress on Pacific Neogene Stratigraphy, in Tokyo.

The participants at the symposium examined the key sections of the Pliocene and Lower Quaternary on the Boso Peninsula and in the Kinki and Kakegawa regions. The Pliocene–Quaternary sequences of Japan are key sections for eastern Asia and the North Pacific region and can be correlated with the stratotypical section of the same age in Italy.

In a number of districts on Honshu Island, three groups of deposits have been defined: the Osaka, Kobiwako, and Tokai groups. In the Osaka group, a little below ash bed Ma-0, an uncomformity indicating a transgression was discovered, with a radiometric age of 1.6–1.7 Ma. This transgression could perhaps correspond to the Calabrian transgression of Italy and, consequently, to events near the N/Q boundary.

The fresh-water Kobiwako group is subdivided into two parts. Early regression in the basin is estimated to have occurred between 1.7 and 2.0 Ma, followed by the formation of Biwa Lake. Thus, the origin of this lake may be coincidental with the Neogene–Quaternary boundary as well. The Tokai group is also subdivided into two parts. The earlier part is close in age to the early Kobiwako group, but the later part is much younger than Biwa Lake.

Technical reports at that session were devoted mainly to the micropaleontology of the N/Q boundary level in deep-sea records.

Round-table meeting in Spain, 1976

In order to help correlate the N/Q boundary in continental environments by means of land-mammal biostratigraphy, the divisions and nomenclature for a mammal biostratigraphic scale were adopted in a round-table meeting in Madrid, 1976, by representatives of the various IUGS and IGCP projects for the Neogene. Those standards were published in *Trabajos Sobre Neogene-Cuaternario, No. 7*, Madrid, Instituto Lucas Mallada CSIC.

Birmingham, 1977

A joint meeting of the INQUA subcommission on the Pliocene–Pleistocene boundary and the IGCP working group on Project 41 was held in Birmingham, England, during the X INQUA Congress. Papers presented at this meeting were published with the proceedings of the congress.

The participants of the meeting heard accounts of the ongoing studies of the N/Q boundary conducted during the intercongress period 1974–1977, prepared by the chairmen of the subcommission and the working group, as well as reports on the studies that were carried out in various countries. The participants were informed that a considerable amount of progress had been made during this period, including collection of abundant material on the geology, biostratigraphy, climatostratigraphy, and, to a lesser extent, the radiometric ages of Pliocene and Lower Quaternary deposits of the various continents, islands, and oceans.

General agreement among the investigators was obtained on the following matters:

1. The Neogene–Quaternary boundary should be drawn in accordance with general stratigraphic principles (i.e., based on changes in the open-ocean marine faunas).
2. Italy should be the stratotypical area for determination of the N/Q boundary.
3. Detailed correlations between marine and continental deposits will be necessary in order to establish the correlation of the N/Q boundary. Data on paleomagnetic stratigraphy and the radiometric ages of deposits must also be taken into consideration. The boundary

defined in marine sections in Italy must be taken as a basis, and then correlated with the continental sequences. These analyses can provide reliable grounds for global correlation.

4. The participants at the meeting emphasized the necessity for further micropaleontological investigations aimed at finding more reliable criteria for definition of the Neogene–Quaternary boundary.

In addition, the participants at the meeting concentrated on the considerable advances in our knowledge in the fields of biostratigraphy and magnetostratigraphy and agreed that the use of radiometric dating would be one of the most important methods for wide-range correlation, although such data were not yet available for many parts of the world. The participants expressed a desire for this method to be used more extensively in the work of Project 41. The participants at the meeting supported the suggestion to take the Vrica section as the N/Q boundary-stratotype and recommended that a detailed, complex study of this section be proposed.

The meeting approved the membership of an editorial board for the final report, to consist of E. Aguirre, M. Alekseev, W. A. Berggren, F. P. Bonadonna, H. B. S. Cooke, R. W. Hey, H. Nakagawa, K. V. Nikiforova, G. Pasini, and R. Selli. At that time, the board recommended further study of the following stratigraphic and chronological levels:

1. The boundary between the Gilbert and Gauss paleomagnetic epochs (3.3–3.5 Ma) that tentatively corresponds to the base of the Astian and Piacenzian of the Italian sections and the Akchagylian in the USSR
2. The beginning of the Matuyama paleomagnetic epoch (approximately 2.5 Ma), corresponding to the boundary of the Lower and Middle Villafranchian and the base of deposits containing the fauna of the Khaprovian complex of mammals in the USSR
3. The beginning of the Olduvai event of normal magnetization, which, at that time, was thought to be dated to approximately 1.79 Ma and to be related to a shift to faunas with arctic elements (mollusks, ostracodes, and foraminifers), as well as to the base of the Upper Villafranchian in Italy, the Apsheronian in the USSR, and deposits containing the mammalian fauna of the Odessa complex in the USSR
4. The boundary between the Brunhes and Matuyama paleomagnetic epochs (0.7–0.8 Ma), which is tentatively correlated with the base of the Cromerian in The Netherlands and, in the USSR, with the base of the Bakunian deposits, as well as with the base of the series of continental deposits enclosing the mammalian fauna of the Tiraspolian complex

IGCP review, 1977

During the meeting of the governing board of the IGCP in Paris in March 1977 it was decided that it was time for an appraisal of the program as a whole. The board requested all project leaders to submit reports to the Secretariat describing the progress and achievements of their projects. Statements for 1973–1977 were presented and published in a special issue of *Geological Correlation* in 1978.

At that time, four different concepts regarding the placement of the lower boundary of the Quaternary were noted:

1. At the location of the cooling event below the Piacenzian in Italian sections, which was related to the *Globorotalia miocenica* zone of Bolli and Premoli Silva, as well as to the Upper Ruscinian and Lower Villafranchian mammalian faunas, and which is close to the boundary between the Gilbert and Gauss paleomagnetic epochs.
2. At the location of the significant climate changes that marked the base of the Middle Villafranchian, coinciding approximately with the boundary between the Gauss and Matuyama paleomagnetic epochs.
3. At the base of the Calabrian Stage, which contains arctic elements (beds with *Arctica islandica*). (At that time, the horizon was believed to be correlated to the base of the *Globorotalia truncatulinoides* zone and the beginning of the Olduvai paleomagnetic event, as well as to Upper Villafranchian faunas of Italy.)
4. At the base of the Cromerian in The Netherlands (base of the "glacial Pleistocene" in Europe), which is close to the boundary between the Matuyama and Brunhes paleomagnetic epochs.

It was pointed out that only two of those suggested boundaries (1 and 3) could be seriously regarded as definitions of the Neogene–Quaternary boundary, because only they were in accord with the general principle of placing such a boundary at a widely recognized horizon in marine sequences, in this case involving planktonic organisms that could allow widespread correlation. The other two suggested boundaries were in nonmarine or local sequences defined by a climatic change.

It was firmly established by the 1948 London IGC resolution that the stratotype sections for the Neogene–Quaternary boundary should be in Italy and should be situated within Mediterranean marine deposits of the Calabrian in relation to the appearance of cold-water Atlantic immigrants. The IGCP-41 working group noted that the Vrica section met all of the necessary requirements to be chosen as a stratotype; it is characterized by continuously deposited strata laid down under bathyal conditions and is rich in fossil organisms, including planktonic groups.

By the time of the Paris IGCP board meeting in 1977, research had been expanded considerably within the framework of the project. For example, additional studies in Java had been initiated with the help of Japanese working group members. Also, the Indian working group had organized studies on the stratigraphy of the Pliocene and Lower Quaternary deposits in the Nicobar and Andaman islands and had extended its interests to remote northern and northeastern regions of the country.

Symposium in Tadjikistan (USSR), 1977

Findings in southern Tadjikistan in 1977 were the subjects of an IGCP-41 symposium. The group studied the following reference sections: Akjar, Karamaidan, Karatau, Lakhuti, Khonako, and Kuruksai. The papers presented during the meeting were devoted to the stratigraphic subdivision of the Upper Pliocene–Quaternary deposits in Tadjikistan, their paleomagnetic characteristics, their fauna and flora, a comprehensive analysis of the loess-soil formations, and studies of the newest tectonic movements and the structural characteristics of the Upper Cenozoic molasse. The problems of the geology of Paleolithic deposits and the ecology of ancient humans in Tadjikistan were examined. Additionally, papers were presented on the stratigraphy of Pliocene and Quaternary deposits, the flora and fauna of Kirgizia, Uzbekistan, Turkmenia, and Kazakhstan, and the problems of stratigraphy of the Pamir glacial deposits and their correlation with glacial sequences in middle Asia. The presented papers appeared in the *Proceedings of the second symposium on the Neogene/Quaternary boundary* (Nikiforova and Dodonov, 1980).

In earlier years, the stratigraphic subdivisions of the Upper Pliocene and Quaternary deposits had been based mainly on geological-geomorphological criteria, together with paleontological and archeological data. Practically no analyses on stratigraphy of overlying deposits and their correlation with alluvial complexes had been carried out. More recently, considerable success has been achieved in studies of the Upper Cenozoic deposits in Tadjikistan by applying interdisciplinary techniques to micropaleontological (i.e., palynological), paleopedological, paleomagnetic, and thermoluminescent methods. Thus, the Tadjik depression has become an important reference region for solving various problems of Quaternary stratigraphy in Central Asia.

Participants at the meeting agreed that interdisciplinary studies of the Upper Cenozoic deposits of Tadjikistan should be continued and widened to establish the climatostratigraphic essence of the previously mentioned alluvial, alluvial-proluvial, and glacial deposits.

Activities in 1978

During 1978, a working group in Greece joined Project 41 under the coordination of A. Psilovikos. A group in Bangladesh also joined the project, with a new branch of the Geological Survey of Bangladesh created to study Quaternary geology. Additional studies in Java were continued by Indonesian and Japanese members of the working group. The work of Project 41 will be important in developing countries of Asia and the Pacific region, complementing IGCP Project 32, "Stratigraphic Correlations between Sedimentary Basins of the ESCAP Region." In 1977, Dr. H. F. Doutch, an expert from the ESCAP Secretariat, participated in the Tadjikistan symposium.

In November 1978, the 15th session of the United Nations Committee for Coordination of Joint Prospecting for Mineral Resources in Offshore Areas of ESCAP met in Bangkok and considered the possibility of cooperation with the stratigraphic research of IGCP Project 41. Considerable attention was given to the studies produced by Project 41 on stratigraphic correlation between the sedimentary basins of the ESCAP region. The developing countries of the ESCAP region were invited to take part in implementation of Project 41 research, and there were discussions with the professional staff of the Geological Survey of Indonesia about possible specific research projects. Various regional meetings were held to discuss the N/Q boundary problem in the USSR (in Bashkiria, Transcaucasia, the lower Volga River basin, etc.), in addition to field conferences, drilling, and field work in the USSR, Hungary, India, Japan, Indonesia, Germany, Spain, Greece, and the USA. New groups of organisms were used for definitions of stratigraphic boundaries in this interval. The number of countries participating in the work increased, and the developing countries began to participate more actively in the work of Project 41. Also in 1978, the INQUA subcommission on European stratigraphy met to discuss the problem of the N/Q boundary. The almost complete absence of correlations between the stratigraphy of continental deposits and the oceanic zonation was widely discussed. That may have influenced the majority of the western European scientists to conclude that the N/Q boundary should be positioned at the base of the Middle Villafranchian (about 2.5 Ma), because that level was marked by an intensive cooling of the climate. However, that cooling is not strongly reflected in the deep-sea sediments, and significant changes in the assemblages of marine organisms have not been recorded at that level.

Activities in 1979

A field conference held in the northwestern part of India in 1979 was devoted to the problems of stratigraphy, paleontology, tectonics, and volcanism of the Upper Pliocene–Lower Pleistocene, as well as to the magnetostratigraphy and geochronology of the Pliocene–Quaternary transitional interval. The presentations were given at sessions on (1) stratigraphy, (2) paleontology and paleogeography, (3) tectonics, volcanism, magnetostratigraphy, and geochronology, and (4) paleoanthropology and archeology. All reports were later published by the Geological Survey of India (Sastry et al., 1981). The scientific sessions were followed by excursions to sections of the Siwalik series in the Himachal Pradesh and Jammu provinces and the Karewa series in Kashmir.

The editorial board met in Madrid in February 1979 to discuss the preparation of the final report on Project 41: clarification of the program, its mode of presentation, and the date of its completion. It was decided to include J. A. Van Couvering (USA) and M. V. A. Sastry (India) on the editorial board. Dr. Van Couvering was named to be editor of the final report.

At the International Pacific Science Congress in Khabarovsk, USSR, in 1979, the secretary of the central working group, M. N. Alekseev, reported on the activities of Project 41. Reports devoted to the N/Q boundary problem on the eastern coast of Asia and in Siberia were presented and discussed at a section on Upper Cenozoic stratigraphy. An excursion to the Kolyma lowland was organized to study the stratigraphy of the Quaternary. Also in 1979, M. N. Alekseev convened a meeting at Sangiran, Java, to review cooperative work between the Geological Survey of Indonesia and Japanese scientists on a detailed study and geological survey of the critical area in central Java, emphasizing correlations between continental and marine Plio–Pleistocene deposits. That led to a more precise understanding of the positions of stratigraphic boundaries in the portion of the section intermediate between the Pliocene and the Quaternary.

Further physical research into the problem of the N/Q boundary was carried out by the national working groups of Hungary, India, Indonesia, the German Democratic Republic, Spain, Italy, Greece, the USA, the USSR, and some other countries. The Project 41 leaders maintained close contact with the investigators of Project 32, "Stratigraphic Correlation between Sedimentary Basins of the ESCAP Region," Project 25, "Stratigraphic Correlation of the Tethys–Paratethys Neogene," and Project 114, "Biostratigraphic Datum-Planes of the Pacific Neogene." Considerable micropaleontological research was conducted by the national working group of Italy to document the Vrica section (Italy) as a key stratotype section for the N/Q boundary. In addition, magnetostratigraphic studies by H. Nakagawa revealed a normal-polarity interval in the predominantly reverse-polarized section that potentially can be considered to represent the Olduvai episode of the Matuyama geomagnetic epoch near the N/Q boundary. Samples were collected for further determination of the radiometric ages of the ash interlayers. A meeting of the Spanish national working group was held to discuss new research on selecting the local parastratotype section. The coastal area from the Portuguese frontier around to the Ebro delta was investigated, and studies of new sites with mammalian fauna and detailed geomorphological study in La Mancha indicated that the Campo de Calatrava volcanic formations in western La Mancha could be used for radiometric age determinations to help date the Spanish N/Q sequence. It was recommended that magnetostratigraphic investigations be carried out in the southern Meseta and in the Baza basin and in the territories of the Cadiz coast, Murcia-Alicante, and the Ebro River.

During 1979, Project 41 engaged 19 regional working groups to conduct broad investigations in the territory of the USSR. The reports of those regional groups were discussed at a meeting of the Soviet working group in March 1979. The Hungarian working group drilled research boreholes in the Great Hungarian Plain, with the goal of interdisciplinary geological analysis. They encountered Quaternary and Upper Pliocene sequences that in some places were continuous and complete, without stratigraphic lapses. The borehole samples were investigated from many points of view, including magnetostratigraphy, and the preliminary results were published.

Activities in 1980

A meeting of the Project 41 working group was held during the 26th session of the International Geological Congress in Paris, jointly with INQUA Subcommission 1-d on the Pliocene–Pleistocene boundary. Twenty-five delegates from various countries attended the meetings. The recent activities of the working group were presented by the leader of the project, K. V. Nikiforova. That was followed by a report on the previous two years of work by the INQUA subcommission and the IUGS working group on the Pliocene–Pleistocene boundary, and a report on the results of a mail consultation conducted by its chairman, E. Aguirre. The current state of research on the sections at Le Castella, Santa Maria di Catanzaro, and Vrica was discussed by G. C. Pasini, and an introduction to the report on Plio–Pleistocene datum levels in the deep sea was given by J. A. Van Couvering. After discussions on the reports, the meeting unanimously adopted the following principles:

1. The lower boundary of the Quaternary would have to be established in accordance with the general principles of stratigraphy (i.e., the decision must conform to the guidelines recommended by the International Commission on Stratigraphic Classification) (Hedberg, 1976).
2. The recommendation of the IGC (London, 1948) should be slightly modified to state that the boundary must be designated as a stratigraphic plane (boundary-stratotype) in a continuous sequence of open-marine deposits.
3. The 1948 recommendation is understood to mean that the base of the Quaternary (viz., the Pliocene–Pleistocene boundary) should be defined by the base of the Calabrian Stage in southern Italy. It was therefore recommended that the Calabrian Stage should be redefined (taking into account modern research, which indicated that the Catanzaro section is not satisfactory to express Gignoux's concept of the Calabrian Stage) to make its base unambiguous.
4. Multiple criteria should be used in selecting a stratigraphic plane for the base of the Calabrian and thus the N/Q boundary-stratotype; that is, all the available evidence that could help wide-range correlations should be taken into account. The positions corresponding to the N/Q (Pliocene–Pleistocene) boundary in other areas would have to be determined by working out local stratigraphic scales and correlating them to the stratotype section. For compilation of the local scales, a synthesis of biostratigraphic, climatologic, magnetostratigraphic, and radiometric techniques should be used. Special attention would have to be paid to the current difficulties of identifying the boundary, by means of correlation, in different latitudes and in continental sequences.

As a first criterion, it was proposed that the N/Q boundary could be placed in the Vrica section at the FAD (first-appearance datum) of the "cold guest" ostracode *Cytheropteron testudo*, whatever its paleoclimatic significance might be. At the time of that meeting it was thought that the first *C. testudo* could be found in the Vrica section 10 m above sapropel *e*, but in later studies (Pasini and Colalongo, Chapter 2, this volume) it was established that this FAD actually occurs somewhat earlier.

An alternative possibility for selecting a level in the Vrica section as the N/Q boundary-stratotype would be to place it within the stratigraphic interval between the level of sapropel *e* and the level of volcanic ash *m*, as close as possible to a paleomagnetic reversal. It was recommended that the final decision on the exact placement of the N/Q boundary-stratotype be deferred until the paleomagnetic record of the Vrica section could be further investigated. F. P. Bonadonna agreed to coordinate that work, which was to include a new detailed sampling and study of the section, with the cooperation of scientists from several institutions and countries.

Studies within the framework of the national working groups were continued throughout 1980. A field conference devoted to the problems of biostratigraphy and magnetostratigraphy was held in Spain, with scientists from Spain, Italy, USA, USSR, and France taking part. The most suitable paleontologically characterized Plio–Pleistocene sections were selected, with the aim of paleomagnetic investigations. In 1980 the working group in Japan summarized its results from thorough studies of sections of the Boso Peninsula (Honshu). The first appearance of *Globorotalia truncatulinoides* was established in the middle part of the Kiwada Formation. Japanese and Indonesian scientists continued studies of Pliocene–Quaternary sections in the central part of Java.

Scientists in the People's Republic of China managed to establish that the loessic series of Malan and Lishi belong to the Brunhes zone of positive paleomagnetic polarity. On the basis of paleomagnetic measurements, the Wucheng loess was attributed to the Matuyama reversed-polarity zone, with the upper part of these loessic series falling in the Upper Matuyama.

The USA working group concentrated on studies of sections in Arizona and California, in preparation for a field conference at Tucson, Arizona, in March–April 1981.

In the USSR, important studies were carried out in preparation for the scientific excursions of the XI INQUA Congress in 1982, primarily in middle Asia, Moldavia, Transbaikalia, and Yakutia. In the eastern provinces of the USSR, local biostratigraphic subdivisions and the alluvial series on the high terraces of the Lena River were correlated to the general magnetostratigraphic scale, and many studies dealt with climatostratigraphic interpretations in both the European and the Asian parts of the USSR.

Activities in 1981

The meeting in Tucson was followed by a field conference on the key sections of Pliocene and Pleistocene deposits in Arizona and California, coordinated by the USA working group. Members from China, Hungary, India, Italy, Spain, the USA, and the USSR took part. The meeting was held jointly with IGCP Project 128, "Late Cenozoic Magnetostratigraphy," and the INQUA subcommission on the Pliocene–Pleistocene boundary.

At the Tucson meeting, the findings from ongoing studies were discussed, and further work on the Vrica section was proposed, to be carried out during 1981 by scientists from Italy and the USA. Following the Tucson meeting, magnetostratigraphic investigations, further collection of radiometric samples, and detailed micropaleontological studies in the Vrica section were carried out. Paleontological studies of the continental deposits of northern Italy were included as stratigraphic analogues of the Pliocene–Lower Pleistocene marine beds.

During 1981, the Spanish working group continued biostratigraphic study of some key sections of Neogene–Quaternary age to develop correlations to the stratotypical section of Italy. Work also continued in various regions of the European and Asian parts of the USSR to compile a series of key sections with clearly distinguishable chronologic, biostratigraphic, and magnetostratigraphic evidence from the Olduvai event to the Gilbert–Gauss boundary. A significant scientific event in this respect was the All-Union Meeting on Quaternary Research at Ufa, Kuibyshev, in August 1981, with part of the program devoted to the problem of the N/Q boundary. The geological excursions included a special visit to Pliocene and Lower Pleistocene deposits in Bashkiria and Kuibyshev Zavolzhie. In addition, the fifth meeting of the Soviet working group, held in April 1981, recommended additional research, including drilling in the deltas of the Dniestr, Don, Dnieper, Lena, and other rivers.

The national working group of China studied the stratigraphic subdivision of the Pliocene–Quaternary deposits in the regions of Pingliang, Xifeng, and Wuchi. In the course of studying the Plio–Pleistocene deposits of the Beijing plain, a biological boundary corresponding to the magnetostratigraphic boundary between the Matuyama and Gauss paleomagnetic zones was established. According to the Chinese working group, the N/Q boundary in China should be drawn at this level, now dated to 2.5 Ma. Future studies were to be carried out in the western part of the North China Plain.

Adoption of a proposal

A meeting of the IGCP-41 working group was held in Moscow during the XI INQUA Congress in 1983, jointly with the INQUA subcommission on the Pliocene–Pleistocene boundary. The purpose of the meeting was to consider the progress made toward implementation of Project 41, the results of studies on the N/Q boundary-stratotype, the problems of correlation of the N/Q stratotype with key N/Q sequences in various parts of the world, and preparation of the final report. At that meeting it was proposed, discussed, and adopted that the Vrica section, located in Calabria, be designated as the Pliocene–Pleistocene boundary-stratotype section.

The Vrica section, located in Calabria, southern Italy, had been extensively described by Selli et al. (1977) and Colalongo, Pasini, and Sartoni (1981), among other reports (Pasini and Colalongo, Chapter 2, this volume). That Plio–Pleistocene section satisfied all the internationally accepted guidelines for an adequate boundary-stratotype: good vertical development (more than 300 m), complete exposure, stratigraphical continuity, abundance and variety of well-preserved fossils, facies (bathyal marine sediments) favorable for recognizing time-significant biohorizons in long-distance correlation, no structural complication or metamorphism, suitability for magnetostratigraphic investigations, and accessibility. The Vrica section offered the possibility of selecting a Pliocene–Pleistocene boundary-stratotype that was consistent with the original concepts of Pliocene and Pleistocene, as described by Lyell and as elaborated by generations of later workers. Moreover, the magnetostratigraphy of the section had already been determined, and some radiometric ages of several included ash levels had been obtained, as reported by Nakagawa et al. (1980), Nakagawa (1981), and Tauxe et al. (1983), as reviewed by Nakagawa et al. (Chapter 3, this volume).

At the 1983 meeting, the members of IGCP-41 and INQUA Subcommission 1-a (i.e., 1-d) initially proposed to the INQUA Commission on Quaternary Stratigraphy that the Pliocene–Pleistocene boundary-stratotype be defined in the Vrica section. The base of the bed of silty-marly claystone conformably overlying the sapropelic bed *e* at the Vrica section, where this level is exposed in profile B, as described by Colalongo et al. (1982), was selected as the boundary marker point or "golden spike," for many reasons. One of those was the fact that several paleontologic markers (first and last occurrences of microfossils) straddling the proposed boundary-stratotype occur in Italian shallow-water sections near the first appearance of *Arctica islandica*, which is one of the main criteria for locating the base of the Pleistocene. The biostratigraphy at the selected boundary level has been reviewed in detail (Pasini and Colalongo, Chapter 2, this volume). It was noted that these paleontologic markers are found in the same order and in the same position with respect to the Olduvai subchron in the Vrica section and in oceanic deep-sea cores. Furthermore, the Mediterranean first appearance of *A. islandica*, historically the index for the beginning of the Calabrian (and, in present terms, the Santernian as well) (Preface, this volume), clearly postdates the upper levels of the Piacenzian stratotype, as defined by Colalongo, Elmi, and Sartoni (1974). According to modern stratigraphic guides, this means that the base of the lowest Pleistocene stage, if defined in a level that accurately reflects the appearance of *A. islandica* in Italian sequences, will also define the top of the Piacenzian, according to the principle of "base defines boundary," at a level that essentially accords with the previously accepted upper limit of the Piacenzian. This condition appears to be met in the Vrica definition, although *A. islandica* itself is not autochthonous in this deep-sea section. Therefore, the recommended Pliocene–Pleistocene boundary-stratotype will coincide with the top of the Piacenzian, the youngest stage of the Pliocene, as well as with the established marine biostratigraphic concept for the beginning of the Pleistocene in Italy.

With the presentation of that final report, the work of the IGCP Project 41 was brought to a successful close.

Subsequent activities

Acceptance of the Vrica section as a stratotype for establishing the N/Q boundary was provisionally approved by members of the working groups of IGCP Project 41 and the INQUA Subcommission 1-d on the Pliocene–Pleistocene boundary. The final decision was submitted to the IUGS Commission on Stratigraphy and then to the XXVII International Geological Congress in 1984. In 1985, the complete Vrica proposal was published (Aguirre and Pasini, 1985) together with the announcement (Bassett, 1985) that the terms of the proposal had been formally ratified by the ICS (International Commission on Stratigraphy) as of May 31, 1985. The Vrica boundary-stratotype was subsequently adopted by the IUGS Executive as a global stratotype section and point (GSSP) (Cowie et al., 1986) and was included as such in the IUGS 1989 Global Stratigraphic Chart (Cowie and Bassett, 1989).

The subject of continental analogues to the earliest marine Pleistocene beds was not addressed in the proposal to IUGS. At present, it can be considered that the basal marine Pleistocene (e.g., the Santernian of the Po Valley) (Ruggieri, Rio, and Sprovieri, 1984) relates only to the upper Villafranchian. The top of the Olduvai paleomagnetic zone coincides very closely with the appearance of the Late Villafranchian mammalian fauna in the continental deposits of the Mediterranean, as exemplified in the fauna of the Olivola and Tasso intervals (Azzaroli et al., 1988; Azzaroli et al., Chapter 11, this volume). The Olivola fauna is presently attributed to the beginning of the Eburonian Stage in The Netherlands; in turn, the beginning of climatic cooling is recorded in faunal and floral markers seen in the Eburonian, as well as in the Baventian in East Anglia and in the lower Apsheronian in the former USSR.

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Part II

Characterization of the Pleistocene boundary-stratotype

2 The Pliocene–Pleistocene boundary-stratotype at Vrica, Italy

GIANCARLO PASINI and MARIA LUISA COLALONGO

Introduction

The Plio–Pleistocene sequence of deep-water marine beds at Vrica, Calabria, Italy, was proposed by INQUA Subcommission 1-d, "Pliocene/Pleistocene Boundary," and by the working group of IGCP Project 41, "Neogene/Quaternary Boundary," as the location for a physical boundary-stratotype for the base of the Pleistocene, according to modern chronostratigraphic guidelines. The recommended level has been adopted formally by the IUGS (Nikiforova and Alekseev, Chapter 1, this volume). That section is herein characterized in terms of sedimentology, paleoecology, biostratigraphy, biochronology, and magnetostratigraphy, based on a decade of studies by stratigraphers in different countries. According to criteria established by prior recommendations, the physical location of the Pleistocene boundary-stratotype is identified with the base of the claystone conformably overlying marker bed *e* of the Vrica section. This level is very close to the Olduvai normal-polarity subzone and is approximately coeval with the beginning of a cold-climate phase marked by the first appearance of the "northern guest" *Arctica islandica*, a mollusk confined to boreal waters during interglacial periods.

Background

At the XVIII International Geological Congress in London in 1948, the council unanimously accepted the recommendation of the temporary commission to advise on the question of the definition of the Pliocene–Pleistocene boundary, which made the following essential points: "The Commission considers that the Pliocene–Pleistocene boundary should be based on changes in marine faunas. . . . The classic area of marine sedimentation in Italy is regarded as the area where this principle can be implemented best" (King and Oakley, 1950).

Following that recommendation, during the joint meeting in India in 1979 of the working group for IGCP Project 41, "Neogene/Quaternary Boundary," and the INQUA Subcommission 1-a (since changed to 1-d), "Pliocene/Pleistocene Boundary," it was further resolved that "the territory of Southern Italy (the region of Calabria) is selected as type area for establishing the Pliocene–Pleistocene boundary, so consequently the type section should be chosen in the marine sequences . . . of Southern Italy."

Accordingly, among the sections located in Calabria, three were considered as candidates for the location of the Pliocene–Pleistocene boundary: (1) Santa Maria di Catanzaro, (2) Le Castella, and (3) Vrica. Studies by Colalongo, Pasini, and Sartoni (1981) indicated that neither the Santa Maria di Catanzaro section nor the Le Castella section was suitable as a standard section for defining the Pleistocene boundary-stratotype, because of evidence that each contains a hiatus corresponding to the uppermost Pliocene and the lowermost Pleistocene. Thus, the only one of those sections suitable for defining the Pleistocene boundary-stratotype was at Vrica, where a continuous, highly fossiliferous bathyal sequence is exposed.

The Vrica section is located about 4 km south of the town of Crotone (Figure 2.1). Like the areas of Le Castella and Santa Maria di Catanzaro, which are 19 km to the southwest and 55 km to the west-southwest, respectively, the Vrica area is an emergent portion of the Crotone sedimentary basin. According to Barone et al. (1982), the Crotone basin is a graben-like structure containing postorogenic sediments deposited since Tortonian times on top of the nappes making up the emerged Calabrian Arc. After the early Pleistocene, a large portion of the Crotone basin emerged north of the Catanzaro isthmus, exposing Miocene sediments and a thick Plio–Pleistocene sequence.

The Plio–Pleistocene sediments of the Marchesato peninsula (Figure 2.1), including the Vrica area, are very gently folded, with axes oriented approximately north–south; the folds are cut by normal faults (Selli, 1977). The Vrica section is located in a regular monocline, dipping 5–15° westward (Figure 2.2).

Physical stratigraphy of the Stuni–Vrica sequence and the Vrica section

In the extensive badlands area located between Crotone and C. Colonne (Figures 2.1 and 2.4) a well-exposed Plio–Pleistocene sequence crops out. The lowermost part of this sequence is formed by gray claystones of the "Semaforo Formation" of Selli (1977) (Figures 2.2 and 2.3), which is middle Pliocene in age. At

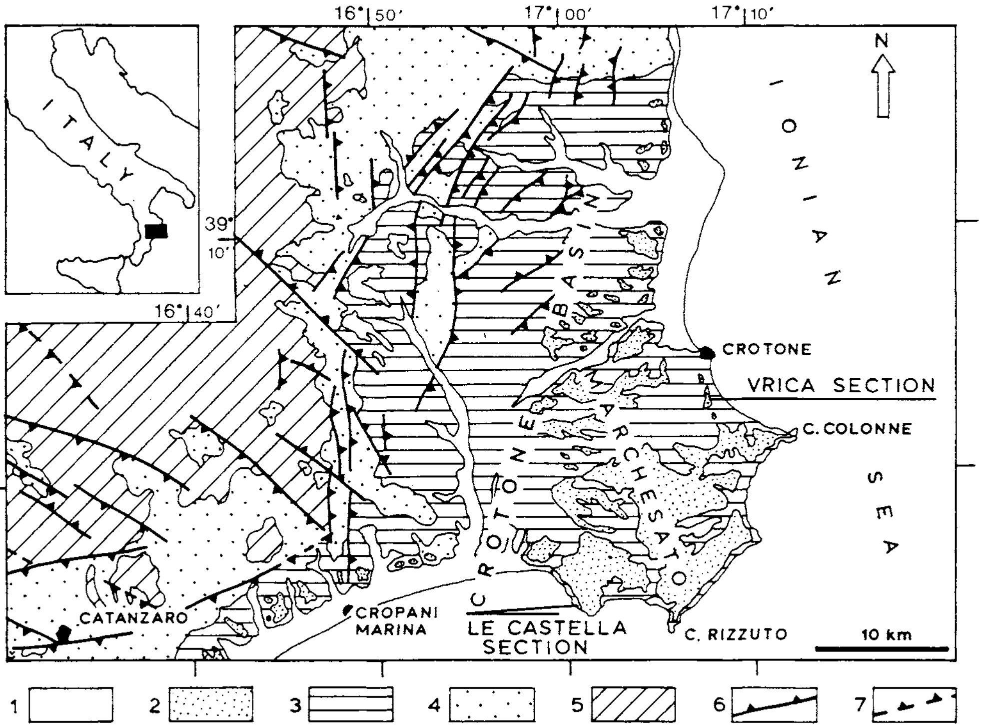

Figure 2.1. Geologic map of the Marchesato Peninsula, an emerged portion of the Crotone sedimentary basin. The submerged part of the basin is located southeast of this peninsula: 1, continental sediments (Pleistocene and Holocene); 2, sediments of the Upper Pleistocene marine terraces; 3, Middle-to-Upper Pliocene and Lower Pleistocene sediments; 4, Tortonian–Lower Pliocene sediments; 5, rocks of the substrate; 6, main faults; 7, possible faults. (From Tortorici, 1982, courtesy of Società di Geologica Italiana.)

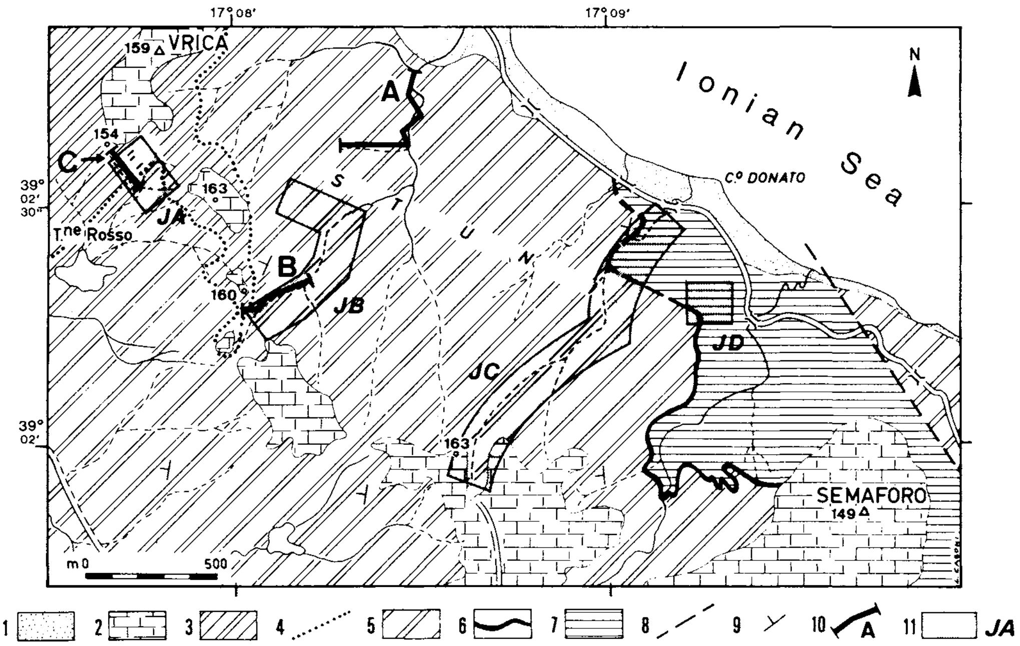

Figure 2.2. Geology of the Vrica area: 1, beach sand and dunes; 2, calcarenites and sands of the "Milazzian" marine terrace; 3, gray silty-marly claystones, with intercalated sapropelic layers (Lower Pleistocene); 4, first appearance of *Hyalinea baltica*, 35 m above the *m* volcanic ash layer (maximum age 1.99 Ma); 5, gray silty-marly claystones, with intercalated sapropelic layers (Lower Pleistocene, Upper and Middle Pliocene); 6, lower volcanic ash layer (*l.a.*), dated to 2.22 Ma; 7, gray claystones (Middle Pliocene); 8, faults; 9, strike and dip; 10, component-sections measured and sampled by Pasini and co-workers (Figure 2.3, second column); 11, component-sections measured and sampled by Nakagawa and co-workers (Figure 2.3, third column). (Adapted from Selli et al., 1977.)

the top of these claystones is the middle Pliocene volcanic ash horizon *l.a.* (=lower ash), with a thickness of about 20 cm. The overlying portion of the sequence, about 380 m thick, is formed by gray silty-marly claystones with several shale layers and a few, rare sandy horizons, belonging to the Upper Pliocene–Lower Pleistocene "Tripolacea Formation" and "Papanice Formation" of Selli (1977). Approximately 100 m below the top of the sequence there is a second volcanic ash horizon, layer *m*, varying in thickness between 2 and 7 cm. The lower and middle parts of this sequence are well exposed in gullies in the Stuni region, and the upper part crops out near the site of Vrica, a ruined farmhouse used as a geodetic point (Figure 2.2). Thus the Stuni–Vrica sequence, as the full section is called, is a composite of correlated component-sections from different areas (e.g., Fig-

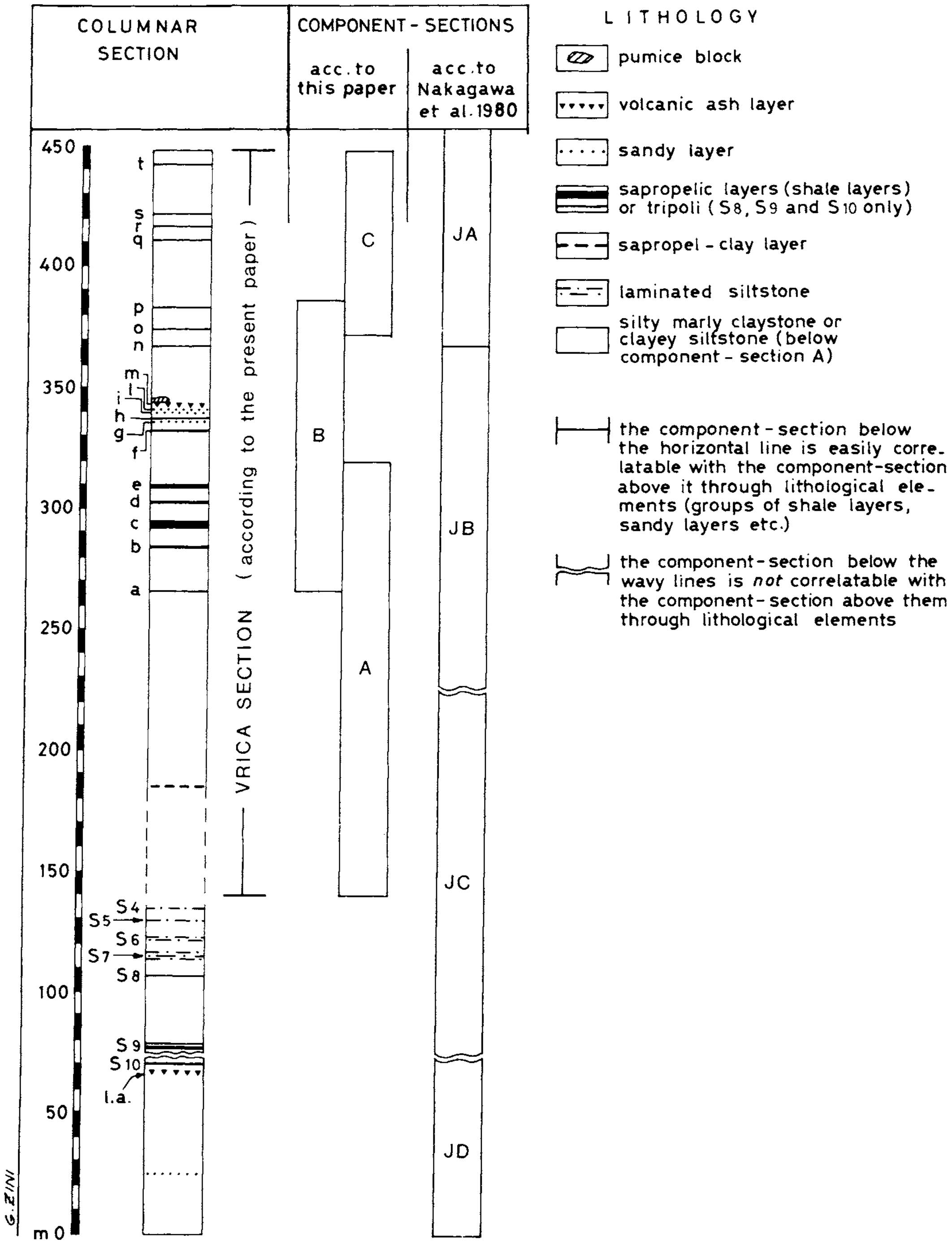

Figure 2.3. The Stuni-Vrica sequence and the Vrica section. The component-sections selected by Pasini et al. (1975), and described in this chapter are compared with those selected by Nakagawa et al. (1980), marked originally as A, B, C, and D, but shown here as JA, JB, JC, and JD, for clarity. The columnar section from zero to 140 m is taken from Nakagawa et al. (1980), and that from 140 m up to the top is from Selli et al. (1977).

ures 2.5–2.7). The Stuni–Vrica sequence has been measured and sampled independently by Italian and Japanese geological teams (Pasini et al., 1975; Selli et al., 1977; Nakagawa, 1981; Nakagawa et al., 1980; Nakagawa et al., Chapter 3, this volume).

The Italian geological team has used the name "Vrica Section" in a restricted sense to mean the middle and upper parts of the Stuni–Vrica sequence (Figure 2.3), specifically a stratigraphic interval 306 m thick measured from 125 m below the marker bed *a* to the exposed top. The level of the Pleistocene boundary-stratotype is approximately in the middle of this measured sequence, about 140 m below the top.

The Vrica stratigraphic section, from which the Italian team has collected 275 samples, is represented in three partly overlapping component-sections identified as A, B, and C (Figures 2.2–2.7). The correlations between the component-sections are based on conspicuous and clearly identifiable groupings of sapropelic shale layers. The beds of the lowermost 45 m of component-section A (from the sapropel–clay layer

Figure 2.4. Aerial photograph of the Vrica outcrop area. For the locations of the component-sections A, B, and C, compare this photograph with Figures 2.19 and 2.20. (I.G.M.I. photo no. 227/10145, 1955, reproduced with permission of the S.M.A. in publication no. 231, of April 7, 1970.)

down to the base: see Figure 2.3) are not well exposed, and the section is synthesized by correlation between scattered outcrops of claystones.

The Japanese geologists initially reconstructed a Plio–Pleistocene sequence in the Vrica area starting from a level 67 m below the *l.a.* (lower ash) horizon up to the top, for a total thickness of about 450 m, in four component-sections D, C, B, and A in ascending stratigraphic order (Nakagawa, 1977, 1981; Nakagawa et al., 1980) (Figure 2.8). In Figures 2.2 and 2.3, and throughout this chapter, the Japanese component-sections are given here as JD, JC, JB, and JA, to distinguish them from the Italian A, B, C, and D component-sections. The component-sections JB and JA were measured in the same gullies where the Italian team measured the component-sections B and C (Selli et al., 1977) and therefore relate to the same boundary-stratotype section and point.

In recent years, the Vrica section has been extended downward and upward. At Monte Singa, Calabria, about 100 km southwest of Vrica, a thick sequence of marine marls and clays with sapropelic layers spans the middle and upper parts of the Pliocene and the lower part of the Pleistocene. This sequence includes the Pliocene–Pleistocene boundary level according to the criteria presented here, and its upper part has been accurately correlated to the Vrica section by Zijderveld et al. (1991). The same authors also correlated the uppermost part of the Vrica section to the Crotone section (located in the environs of this town), where about 40 m of strata overlying marker bed *t* are exposed.

Sedimentology and paleoecology

The Vrica section, 306 m thick, is a well-exposed, continuous sequence of silty-marly claystones sparsely intercalated with laminated sapropelic shales and fine sands, with one volcanic ash layer (Figure 2.3). As noted, 275 samples were collected for biostratigraphic and sedimentologic studies.

Silty-marly claystones

The Vrica section is dominated by silty-marly claystones, gray or blue-gray in color, obscurely bedded, and containing 60.0–78.9% clay, 20.5–38.3% silt, and 0.7–9.3% fine to very fine sand. Its carbonate content ranges from 14.5% to 25.1%. These sediments were formed mainly from clay minerals and the remains of calcareous nannoplankton and foraminifera (with planktic species more abundant than benthic). Variable, though subordinate, amounts of volcanic glass and detrital minerals (feldspars, micas, glauconite) are present. Rarely, the claystones show surfaces and beds with red staining, probably due to limonite. Other claystone beds show burrows filled with indurated clay (often reddish), and some are distinguished by abundant brachiopods, echinoid fragments, or pteropods. Well-preserved mollusks, mostly small gastropods, are rare; a few solitary corals, otoliths, and fish teeth have also been collected.

In addition to calcareous nannofossils and foraminifera, the well-preserved microfauna includes abundant ostracodes. The presence within the claystones and throughout the section of psychrospheric ostracodes such as *Agrenocythere pliocenica, Bathycythere vanstraateni, Zabythocypris antemacella,* and *Bythoceratina scaberrima mediterranea* (Colalongo and Pasini, 1980a), deep-water benthic foraminifera including *Articulina tubulosa, Bolivina albatrossi, Cassidulina carinata, Discospirina italica, Eggerella bradyi, Gyroidina soldanii, Hoeglundina elegans, Planulina wuellerstorfi, Karreriella bradyi, Pleurostomella alternans, Pullenia bulloides*, and *Rhabdammina linearis* (D'Onofrio, 1981), and the deep-water mollusks *Propeamussium duodecimlamellatum* and *Malletia excisa* (Selli et al., 1977) indicates that the deposition of the clays took place in a bathyal environment, in water depths of approximately 500–800 m. This paleodepth range is in agreement with the interpretation of the fish fauna, as discussed later.

Shale layers (sapropelites)

Fourteen conspicuous layers of thinly laminated shales are interbedded with the massive claystones of the Vrica section.

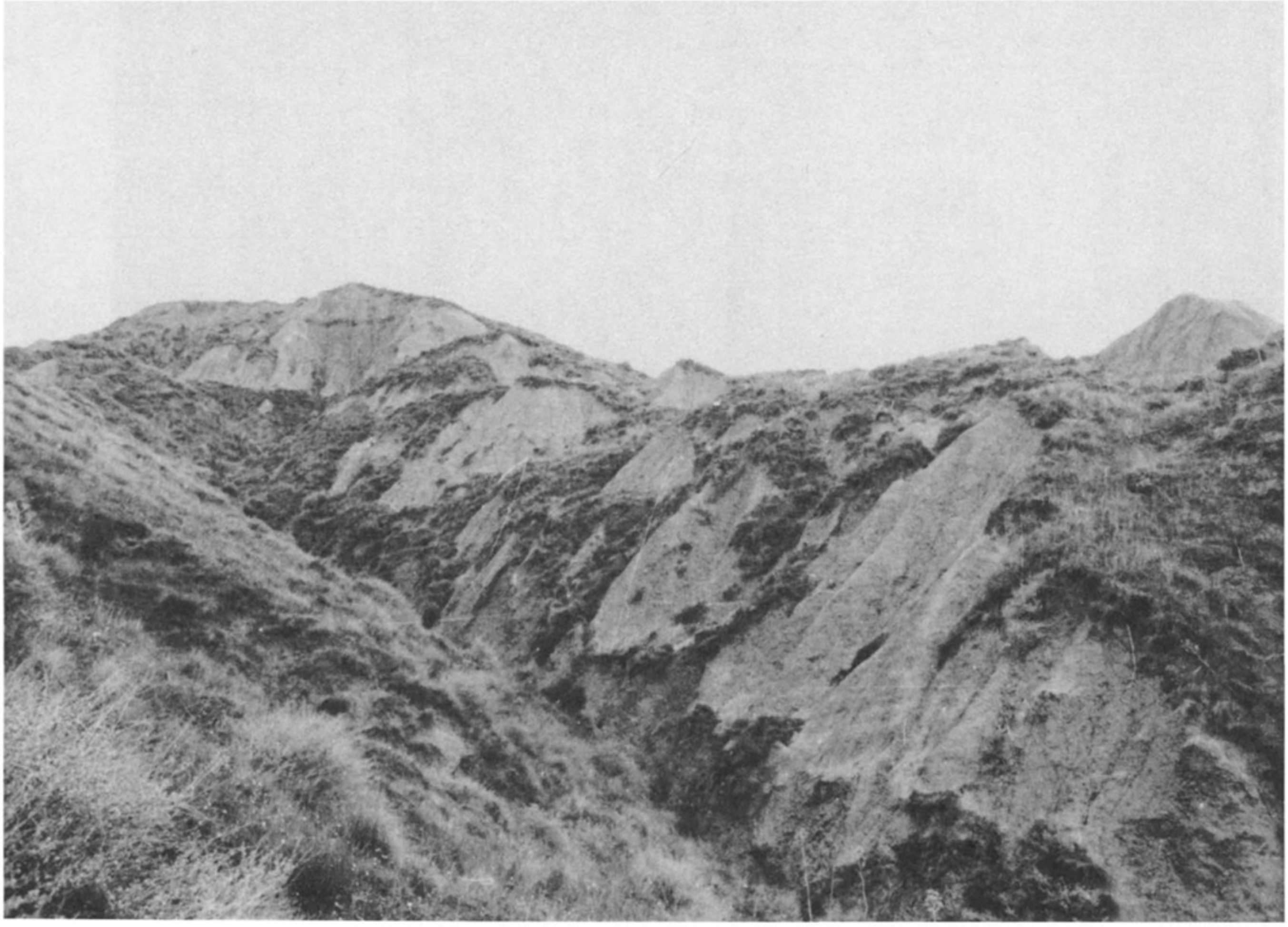

Figure 2.5. Portion of component-section A of the Vrica section, located above the sapropel-clay layer (see Figure 2.3).

These shale layers are useful marker beds; their thicknesses, measured in the line of section (Figure 2.2), are approximately as follows (from top down):

shale layer *t*	75 cm
shale layer *s*	65 cm
shale layer *r*	35 cm
shale layer *q*	80 cm
shale layer *p*	55 cm
shale layer *o*	55 cm
shale layer *n*	100 cm
shale layer *h*	90 cm
shale layer *f*	115 cm
shale layer *e*	190 cm
shale layer *d*	155 cm
shale layer *c*	340 cm
shale layer *b*	115 cm
shale layer *a*	70 cm

The shale units are gray-pink in color and show undisturbed primary laminae parallel to the bedding that measure from a few millimeters to less than 1 mm in thickness. The composition of the shales is 56.7–78.3% clay, 20.6–36.8% silt, and 0.9–11.3% fine sand, with a carbonate content ranging from 11.3% to 19.2%.

In our opinion (Pasini and Colalongo, 1982) the laminites of the Vrica area should be interpreted as sapropelic layers and possibly as true sapropels (Kidd, Cita, and Ryan, 1978) from which some organic carbon has been removed by surface weathering. Like typical sapropels (Sigl et al., 1978), the Vrica organic laminites appear to have been formed during short periods of oxygen deficiency in the bottom water, most probably during intervals of poor ventilation related to overwhelming surface productivity, judging from the following observations. First, because primary laminae deposited on well-oxygenated bottoms are quickly destroyed by mud-eating organisms, such laminae can be preserved only where such organisms are rare or absent because of anoxic conditions at the sediment interface. In the Vrica laminites, a very few clay-filled burrows, all starting from the overlying claystones, indicate that mud-eating organisms that lived *after* the deposition of the sapropelite shales

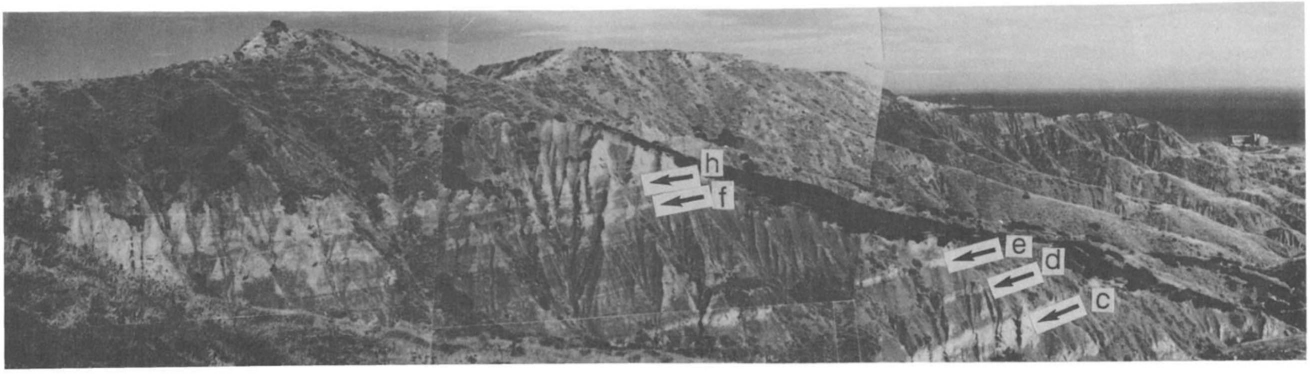

Figure 2.6. Component-section B of the Vrica section (in the foreground). Marker beds *c*, *d*, *e*, *f*, and *h* are indicated. On the right side, the uppermost part of component-section A and the Costa Tiziana Hotel are visible.

Figure 2.7. Component-section C of the Vrica section, exposed along the ravine to the left.

sometimes deepened their burrows down into the shale layers. Second, the planar and undisturbed lamination also suggests the absence of bottom currents. Finally, the virtual absence within the shale units of the benthic epifauna that is common in the enclosing mudstones is another indication of a severe oxygen deficit in bottom waters. According to D'Onofrio (1981), the presence of dwarfed, rare bolivinids and brizalinids is due to "the presence of oxygen-depleted waters" during deposition of the shale layers.

Correlative evidence comes from the presence of fish remains in all of the sapropelic layers examined (9 out of the 15 shale layers, *b* to *p*) (Landini and Menesini, 1978a). The abundance and good preservation of the fish fossils indicate the absence of scavengers and aerobic bacteria (Brongersma-Sanders, 1957). In addition, the presence of minerals that are formed through alteration of iron sulfides, such as gypsum, goethite, limonite, and (rarely) jarosite, suggests that the Vrica laminites contained primary sulfides, strongly indicating euxinic conditions. It may be that some of these layers are coeval (considering the geographic location of the area) with the Pliocene and Quaternary sapropels of the eastern Mediterranean (Stanley, 1978; Kidd et al., 1978), but Raffi and Thunell (Chapter 4, this volume) were not able to identify any of the Vrica sapropelic layers with Plio–Pleistocene sapropels in cores from DSDP (Deep Sea Drilling Project) site 125.

The fish species identified by Landini and Menesini (1978a) from the sapropelic layers today live exclusively or preferentially at depths exceeding 500 m, which supports the evidence from the benthic microfauna that the Vrica sediments were deposited in a bathyal environment.

The sapropelic laminites of the Vrica section have been further studied by Howell, Rio, and Thunell (1990), Hilgen (1991), and Lourens et al. (1994). In the two latter papers the sapropelites were calibrated to astronomically forced climate cycles.

Sandy layers

About 100 m below the top of the Vrica section, three closely spaced sandy layers (*g*, *i*, and *l*) are interbedded with the claystones. These layers are pale gray in color, weathering to reddish. The thicknesses of the sandy layers in the line of section are as follows:

sandy layer *l*	≈3–6 cm
sandy layer *i*	≈3–6 cm
sandy layer *g*	≈6–10 cm

These layers, which are very useful (kilometers wide) marker beds, contain 1.0–3.5% fine sand (3 to 2 ϕ units), 37.1–60.4% very fine sand (4 to 3 ϕ units), 15.2–21.1% silt, and 22.0–40.8% clay. The carbonate content ranges from 12.3% to 14.3%.

The sandy layers are composed of calcareous nannoplankton, foraminifera, volcanic glass (more or less altered), clay minerals, and relatively coarse-grained terrigenous debris, including quartz (mostly metamorphic), altered plagioclases, micas, and sparse lithic fragments. The terrigenous debris, mostly of metamorphic origin, probably came from the Paleozoic crystal-

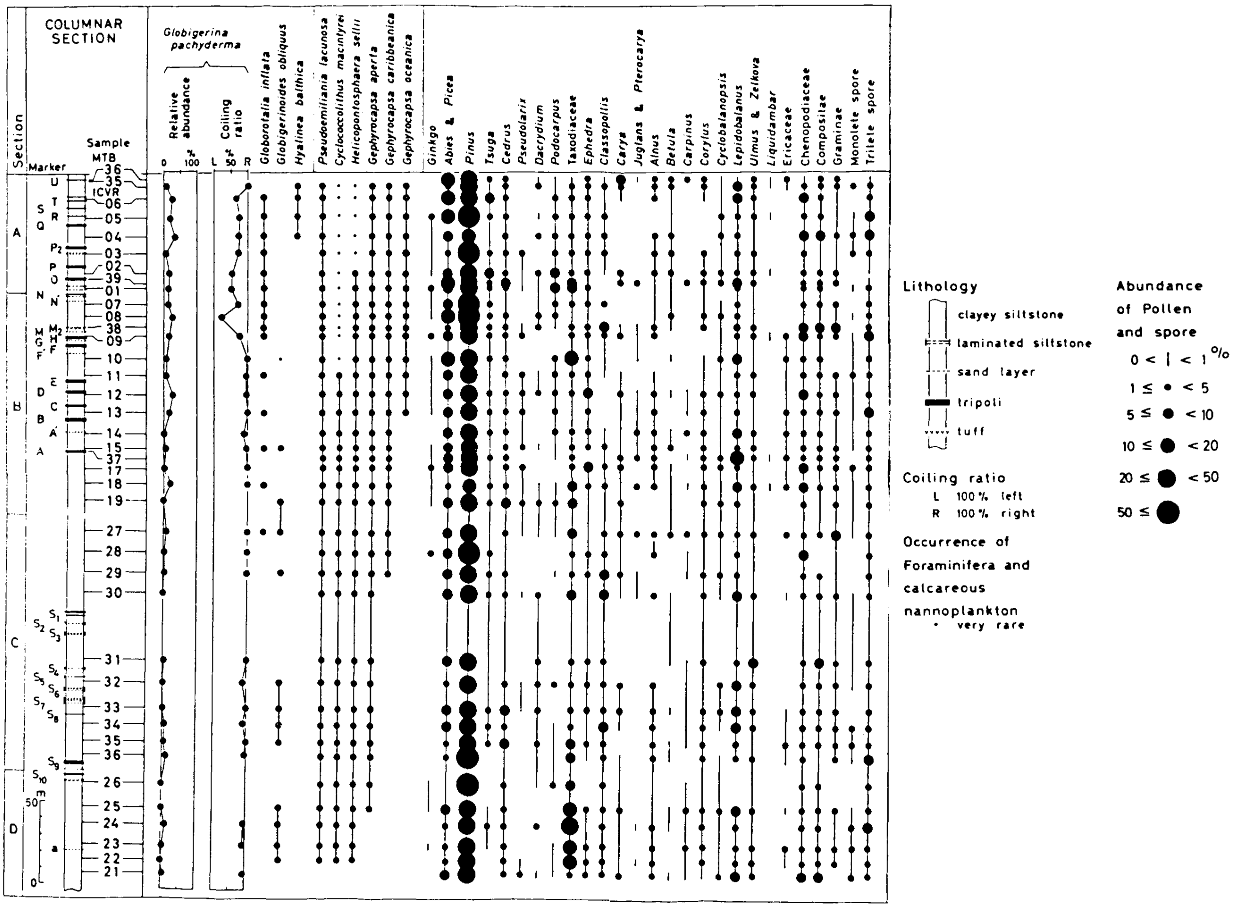

Figure 2.8. Stratigraphic distributions of foraminifera, calcareous nannofossils, pollen, and spores in the Stuni-Vrica sequence, according to Nakagawa and co-workers. (Adapted from Nakagawa, 1981, courtesy of the Geological Survey of India.)

line rocks of the Sila massif, about 40 km west-northwest of the Vrica area.

Volcanic ash layer m *and pumice block*

Layer *m*, a 2–7-cm-thick silt-textured layer of pale gray color (reddish where weathered), crops out on the line of component-section B about 6 m above the top of the sapropelic shale layer *h*, and 2.5 m above the sandy layer *l*. Macroscopically it is very similar to the sandy layers *g*, *i*, and *l* and is also a useful marker. It has the following granulometry: 3.9% fine sand, 10.9% very fine sand, 51.4% silt, and 33.8% clay; the carbonate content is relatively low, at 3.3%.

Mineralogically, layer *m* is made up of about 80% very fresh, colorless, often fibrous or vesicular volcanic glass, with small amounts of fresh, euhedral to subhedral andesinic plagioclases, well-preserved biotite, altered micas, amorphous ferric aggregates, microfossils, and carbonate fragments. The glass is chemically rhyo-dacitic, with a refraction index of 1.512 ±0.001 (Savelli and Mezzetti, 1977). Plagioclase and biotite crystals are included within the glass shards in some samples. The abundance of glass and the freshness of the volcanic materials, including the glass itself, indicate a primary ashfall that was essentially synchronous with eruption.

A block of pale-gray pumice, about 30 cm in diameter and perfectly preserved, was delivered to one of the authors (G.P.) by a quarryman from Crotone who had extracted it from a clay quarry some 3 km north of component-section C and about 1 km south of the center of Crotone, near the coastal road. In that quarry the volcanic ash layer *m*, the three sandy layers *g*, *i*, and *l*, and the sapropelic shale layers *f* to *p* crop out. The quarryman was able to show the exact spot where the pumice block was found, about 1 m above the volcanic ash layer *m* (Selli et al., 1977) (Figure 2.3).

The pumice block consists of fresh, colorless glass (refraction index = 1.500 ±0.001) and includes phenocrysts of hornblende and plagioclase (Savelli and Mezzetti, 1977; Obradovich et al., 1982). The mineralogy and chemistry of the pumice are unlike those of the underlying volcanic ash layer *m*, which contains biotite and plagioclase. Instrumental neutron activation analysis (INAA) of the glass fraction indicates that the pumice is in fact very similar to the thick volcanic ash layer in the lower part of the Stuni–Vrica sequence (layer *l.a.* in Figure 2.3) (Obradovich et al., 1982).

Biostratigraphy and biochronology

The Vrica section is very rich in fossils. Groups that have been studied include calcareous nannoplankton, planktic and benthic foraminifera, ostracoda, mollusks, fish, and pollen, as well as diatoms (Palmer, Chapter 6, this volume). Besides these groups, silicoflagellates, radiolarians, octocorals, brachiopods, pteropods, and echinoderms are present.

In this section we consider the biostratigraphy and the biochronology of the studied groups, in particular the fossils whose first appearances (FA) or last occurrences (LO) in the Mediterranean and also in the extra-Mediterranean regions are considered by most authors as key events in late Pliocene or early Pleistocene time. These biotic events can now be related to the boundary-stratotype (*sensu* Hedberg, 1976) of the Pleistocene as adopted in the Vrica section.

Calcareous nannoplankton

The calcareous nannoplankton of the Vrica section have been studied by Nakagawa (1981), Nakagawa et al. (1980), Cati and Borsetti (1981), Raffi and Rio (in Pasini and Colalongo, 1982), Backman, Shackleton, and Tauxe (1983), Lourens et al. (1994), and Rio, Raffi, and Backman (Chapter 5, this volume) (Figures 2.8–2.11).

Rio et al. (Chapter 5, this volume) emphasize that because of reworked specimens, the extinction levels of species in the section can be accurately identified only through careful semiquantitative or quantitative analysis (Raffi and Rio, cited in Pasini and Colalongo, 1982; Backman et al., 1983). The latter authors have also standardized the informal morphometric taxonomy of the *Gephyrocapsa* group as used by Rio (1982), in place of the historically varied taxonomic concepts that have made the studies in previous publications difficult to compare.

Among the major nannofossil events identified by Rio et al. (Chapter 5, this volume) in the Vrica section, the most important, from both the stratigraphic and biochronologic points of view, are (from younger to older) as follows: the first appearance of *Gephyrocapsa* spp. with coccoliths larger than 5.5 μm (i.e., "large" *Gephyrocapsa*); the extinction of *Calcidiscus macintyrei*; the first appearance of *Gephyrocapsa oceanica s.l.*, *sensu* Rio (1982) (i.e., "medium-sized" *Gephyrocapsa*, between 4.0 and 5.5 μm; *teste* D. Rio); the synchronous extinctions of *Discoaster brouweri* and *D. triradiatus*. The relationships of these events to one another and to the magnetostratigraphy are the same in the Vrica section as in cores from the Mediterranean, the Atlantic, the Pacific, and the Caribbean.

Lourens et al. (1994) calibrated the Lower Pleistocene sapropels of the Vrica–Crotone composite section (as discussed earlier) to the astronomical record in order to obtain an accurate, high-resolution chronology for the Pleistocene part of this section. Those authors analyzed oxygen isotopes from microfossil samples and used their new age model to construct a º^{18}O curve. They concluded that "identification of oxygen isotope stages in the Vrica/Crotone composite and their correlation to obliquity is consistent with the astronomical calibration of these stages proposed by Shackleton et al. (1990)" (Lourens et al., 1994). In addition, those authors carried out a biostratigraphic study of the calcareous nannofossils and selected planktic foraminifera in the Vrica–Crotone composite section to correlate calcareous plankton events to the oxygen-isotope stages. The stratigraphic positions of the nannofossil events indicated by Lourens et al. (1994) are very similar to those shown herein (Figure 2.9).

Raffi et al. (1993) studied the Plio–Pleistocene nannofossil biostratigraphy of DSDP site 607 (midlatitude North Atlantic) and Ocean Drilling Program (ODP) site 677 (eastern equatorial Pacific) and pointed out that "the successful astronomical calibration of oxygen isotope stratigraphies from Deep Sea Drilling Project site 607 and Ocean Drilling Program site 677 in the Matuyama Chron [as discussed by Shackleton et al., 1990] permits calibration of the biostratigraphic events to these uniquely resolved isotope chronologies." In ODP Leg 138 sites (eastern equatorial Pacific), Shackleton et al. (1995a) dated numerous first and last occurrences of Neogene microfossils belonging to different groups (nannofossils, radiolaria, planktonic foraminifera, diatoms) based on the astronomically calibrated time scale of Shackleton et al. (1995b).

The results obtained by Lourens et al. (1994), Raffi et al. (1993), and Shackleton et al. (1995a), as they pertain to the calcareous nannofossil biostratigraphy in the Vrica section, will be discussed in order of increasing age. Ages shown in italics are those calculated for the Vrica section by Lourens et al. (1994).

Event 4: 1.56 Ma(?). First appearance (FA) of "large" *Gephyrocapsa*. This event (Figures 2.9, 2.11) occurs at isotope stage boundary 53/52 in the Vrica section, dated to 1.56 Ma, and practically in the same position in the Singa section (Lourens et al., 1994). In DSDP site 607, the FA of "large" *Gephyrocapsa* occurs in the top part of stage 49, dated to 1.479 Ma; in ODP site 677 this event is recorded in the top part of stage 48, dated to 1.457 Ma (Raffi et al., 1993). In ODP Leg 138 sites the estimate for the time of this event is 1.44 Ma (Shackleton et al., 1995a). According to Berger et al. (1994), the FA of "large" *Gephyrocapsa* occurs in the western equatorial Pacific at 1.515 ±0.025 Ma. If we accept the age estimate of Lourens et al. (1994), this event would have occurred in the Mediterranean about 80 k.y. earlier than in the midlatitude North Atlantic. According to F. J. Hilgen (personal communication), one option would be to accept that the "large" *Gephyrocapsa* FA was in fact earlier in the Mediterranean than in the open oceans; a second option would be a hiatus in both the Vrica and Singa sections between the LO of *C. macintyrei* and the FA of "large" *Gephyrocapsa*. Considering that the *C. macintyrei* LO has almost the same age in the Vrica section as in the Singa section (Lourens et al., 1994), it would be required that such a hiatus would have to represent precisely the same time interval in both sections, which are about 100 km apart and had different sedimentation rates. In our opinion, the presence of such a hiatus is very unlikely, and the noted event probably was diachronous between the Mediterranean and the open oceans.

Event 3: 1.67 Ma. Extinction of *Calcidiscus macintyrei*. This event is recorded in the Vrica section at stage boundary 59/58. In DSDP site 607 it occurs at stage boundary 58/57, dated to 1.640 Ma; in ODP site 677 it is recorded in the top part of stage 55, dated to 1.597 Ma (Raffi et al., 1993). The estimate for the extinction of *C. macintyrei* in ODP Leg 138 sites is 1.58 Ma (Shackleton et al., 1995a). In ODP site 806 this event is dated to 1.627 ±0.025 Ma (Berger et al., 1994).

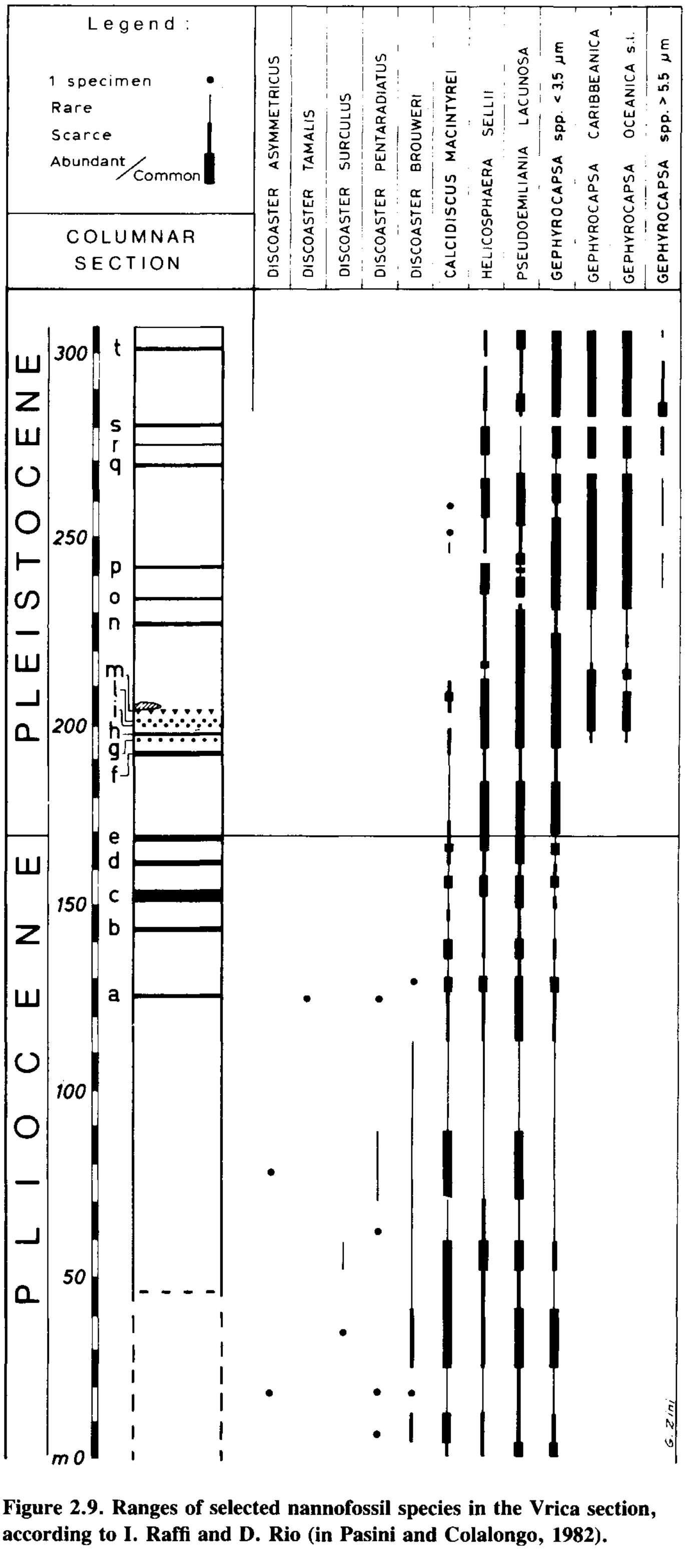

Figure 2.9. Ranges of selected nannofossil species in the Vrica section, according to I. Raffi and D. Rio (in Pasini and Colalongo, 1982).

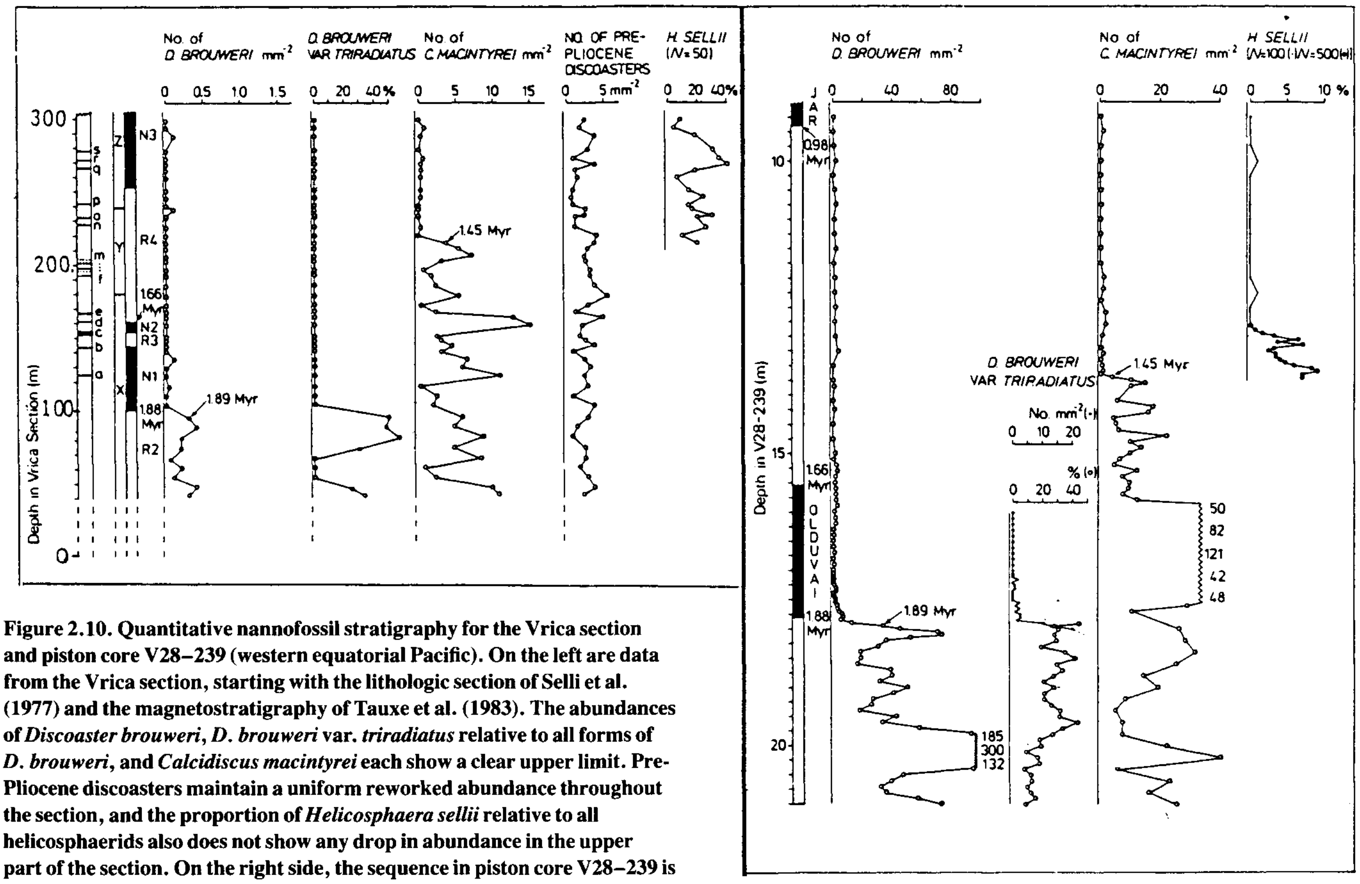

Figure 2.10. Quantitative nannofossil stratigraphy for the Vrica section and piston core V28–239 (western equatorial Pacific). On the left are data from the Vrica section, starting with the lithologic section of Selli et al. (1977) and the magnetostratigraphy of Tauxe et al. (1983). The abundances of *Discoaster brouweri*, *D. brouweri* var. *triradiatus* relative to all forms of *D. brouweri*, and *Calcidiscus macintyrei* each show a clear upper limit. Pre-Pliocene discoasters maintain a uniform reworked abundance throughout the section, and the proportion of *Helicosphaera sellii* relative to all helicosphaerids also does not show any drop in abundance in the upper part of the section. On the right side, the sequence in piston core V28–239 is from Backman and Shackleton (1983), with magnetostratigraphy from Shackleton and Opdyke (1976). (From Backman et al., 1983, with permission of the editors of *Nature*.)

Event 2: 1.71 Ma. First appearance of *Gephyrocapsa oceanica s.l.*, *sensu* Rio (1982) (i.e., medium-sized *Gephyrocapsa* between 4.0 and 5.5 μm). This event in the Vrica section is documented immediately below marker bed *g*, just above the Olduvai subzone (Figures 2.9 and 2.18), and occurs within stage 60 both here and in DSDP site 607, where it is dated to 1.700 Ma (Raffi et al., 1993). In ODP Leg 138 sites the estimate for the time of this event is 1.69 Ma (Shackleton et al., 1995a), and in ODP site 806 it is 1.664 ±0.025 Ma (Berger et al., 1994).

Event 1: 1.95 Ma. Extinction of *Discoaster brouweri*. This event is recorded in stage 72 in the Vrica section (Lourens et al., 1994) (Figure 2.4), practically in coincidence with the lower boundary of the Olduvai normal-polarity subzone, as defined by Zijderveld et al. (1991). In DSDP site 607 it occurs at stage boundary 72/71, dated to 1.950 Ma, simultaneously with the extinction of *Discoaster brouweri* and *D. triradiatus*. Raffi et al. (1993, pp. 399–400) affirm in this regard that "the results from core V28–239 [western equatorial Pacific] and those from DSDP site 606 [subtropical North Atlantic] (Backman and Pestiaux, 1987) . . . seem compatible with those from site 607, which indicate that both *D. brouweri* and *D. triradiatus* have their last occurrence virtually at the Olduvai . . . [lower] boundary, although this disappearance event probably occurred a few thousand years after the reversal, within the basal Olduvai. Both the disappearance event and the reversal boundary are assigned age estimates, however, of 1.95 Ma [according to] Shackleton et al. (1990)." In ODP site 677 the extinction levels for *D. brouweri* and *D. triradiatus* are not clear, but the estimate for the extinction of *D. brouweri* in ODP Leg 138 sites is 1.96 Ma (Shackleton et al., 1995a).

According to Nakagawa (1981), Raffi and Rio (in Pasini and Colalongo, 1982), Backman et al. (1983), and Rio et al. (Chapter 5, this volume), *Helicosphaera sellii* is present throughout the Vrica section (Figures 2.8–2.11), and therefore the top of this section is older than the regional extinction of this species. According to Lourens et al. (1994), the *H. sellii* LO occurs in the Crotone section about 35 m above marker bed *t*, at the isotope stage 38/37 boundary.

Planktic foraminifera

The planktic foraminifera assemblages of the Vrica section are very diverse and abundant, in genera as well as species, and are generally very well preserved. Figure 2.12 shows the stratigraphic distributions of selected planktic foraminifera in the Vrica section, based on studies by Colalongo and Sartoni (in Selli et al., 1977), Colalongo et al. (1980, 1981, 1982), Pasini and

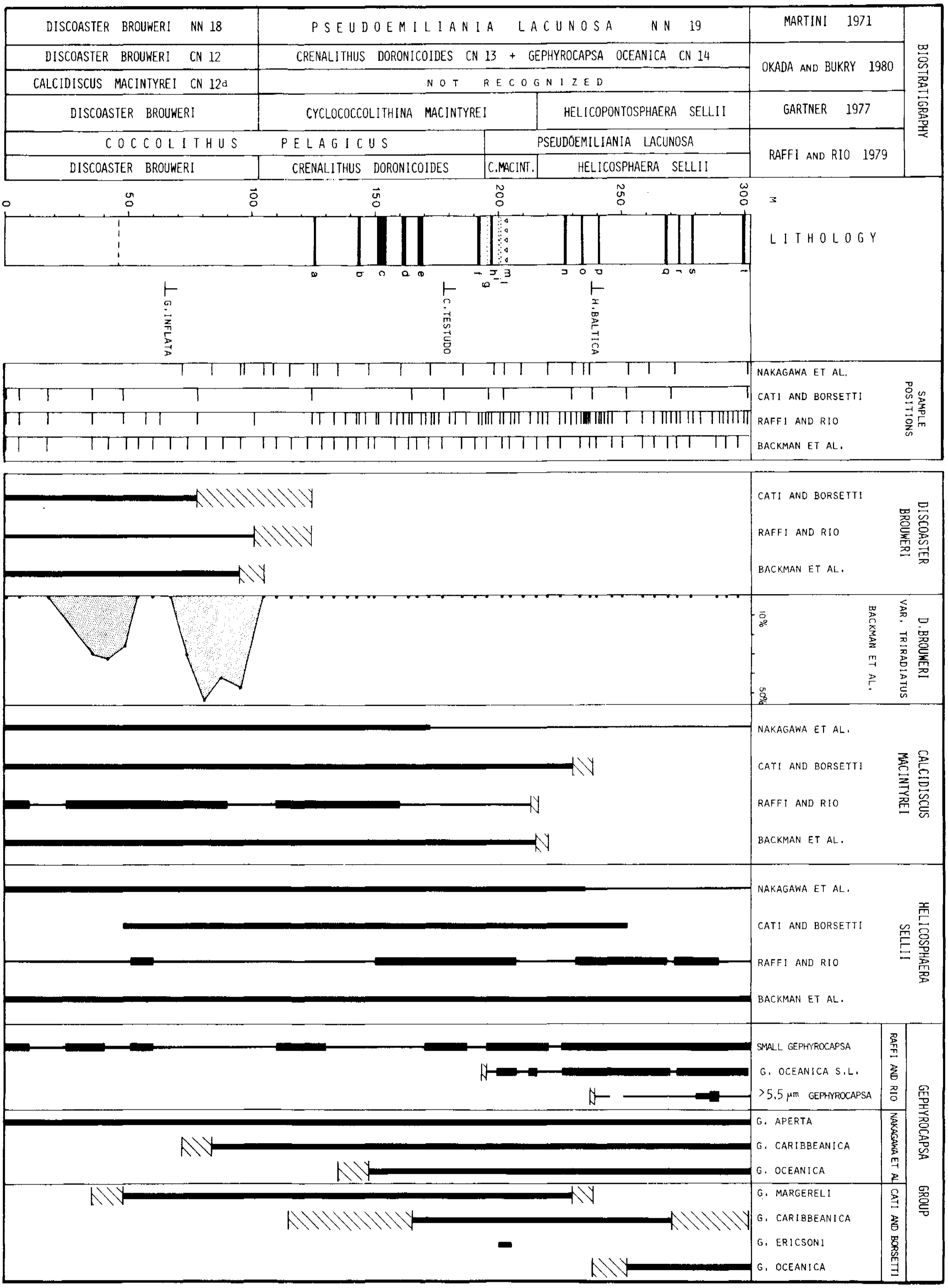

Figure 2.11. Summary of nannofossil biostratigraphy in the Vrica section. Data are from Rio et al. (Chapter 5, this volume), Nakagawa et al. (1980), Cati and Borsetti (1981), I. Raffi and D. Rio (in Pasini and Colalongo, 1982), and Backman et al. (1983).

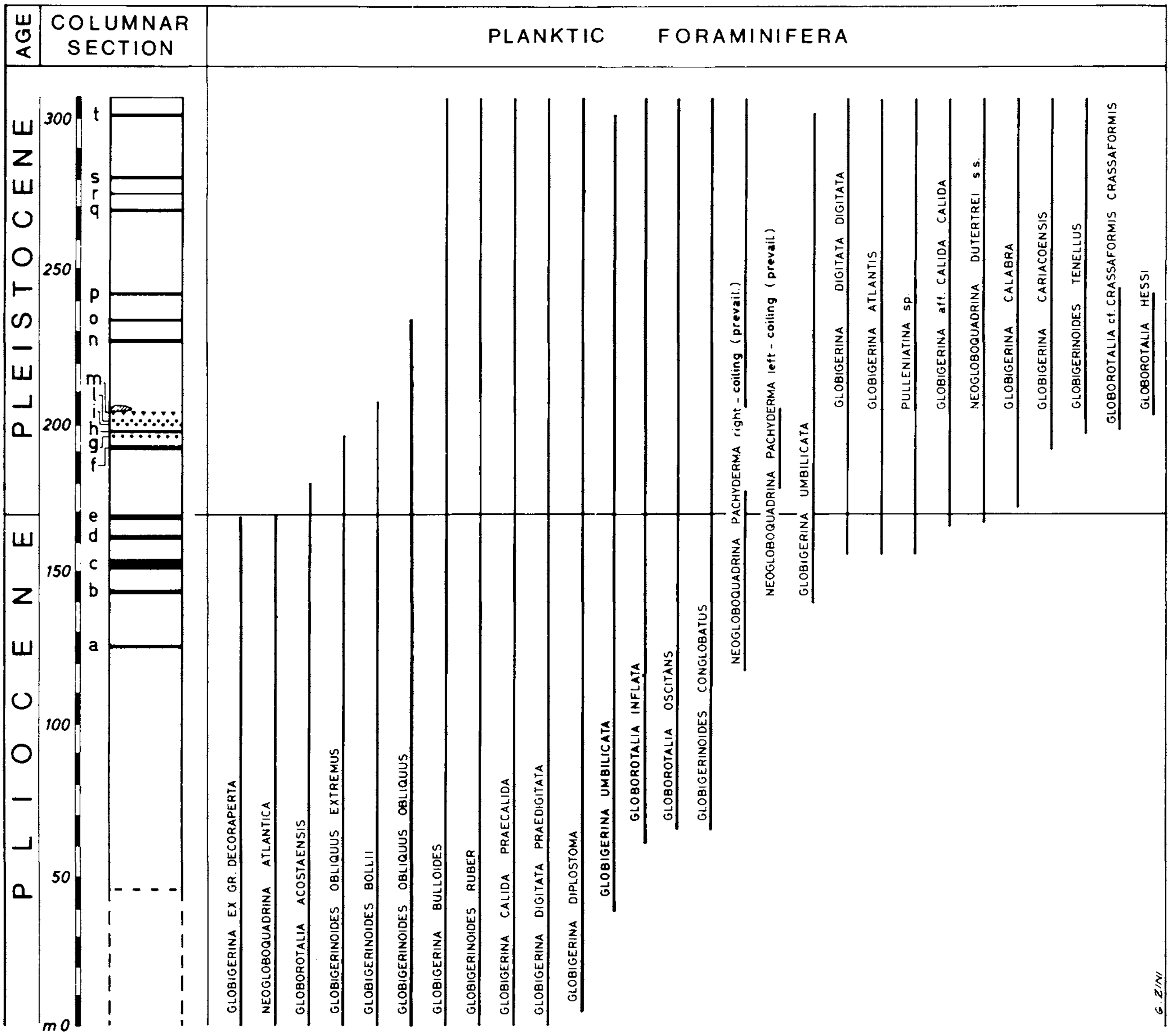

Figure 2.12. Distribution of selected planktic foraminifera in the Vrica section. Data are from M.-L. Colalongo (in Selli et al., 1977) and Colalongo et al. (1981).

Colalongo (1982), as well as on recent research carried out by one of us (M.L.C.). Planktic foraminifera at Vrica have also been studied by Nakagawa et al. (1980) and Nakagawa (1981), as indicated in Figure 2.8, as well as by Spaak (1983), Zijderveld et al. (1991), Sprovieri (1993), and Lourens et al. (1994).

According to Colalongo et al. (1984), among the planktic foraminifera events recorded in the Vrica section, the most reliable for interregional correlation and dating are the *Globorotalia inflata* FA, the beginning of dominantly sinistral coiling in *Neogloboquadrina pachyderma*, and the *Globigerina cariacoensis* FA.

Globorotalia inflata first occurs in the Vrica section between 62 and 67 m below marker bed *a*. Lourens et al. (1992) recorded the first appearance of *G. inflata* in the upper Singa section in oxygen-isotope stage 78, which those authors date to 2.076 Ma. According to Sprovieri (1993), who found that fluctuations in planktic foraminifera abundances in Mediterranean Plio–Pleistocene sequences were in phase with astronomically forced oxygen-isotope stages, the *G. inflata* FA in all studied sequences is coincidental with oxygen-isotope stage 80. Sprovieri affirmed that the most reliable estimate for the age of fluctuation 80 is 2.13 Ma, obtained by using the time scale of Hilgen (1991). On the other hand, Raymo et al. (1989) found the *G. inflata* FA within isotope stage 78 at DSDP sites 607 and 609, and perhaps slightly earlier at DSDP site 552A; those three sites are located in the North Atlantic, at latitudes of 41°, 50°, and 56°, respectively.

The beginning of dominant sinistral coiling in *Neogloboquadrina pachyderma* occurs about 178 m above the base of the Vrica section and is practically coincident with the top of the Olduvai subzone as defined by Zijderveld et al. (1991) (Figure 2.18). This event is also almost coincidental with the top of the Olduvai in sites 650 and 651 of ODP Leg 107 (Tyrrhenian Sea) and in DSDP sites 552, 607, 609, and 610 in the North Atlantic, whereas in DSDP site 611 in the North Atlantic it slightly predates the top of the Olduvai (Raymo et al., 1989; Channel et al., 1990). According to Lourens et al. (1994), in the Vrica section the first substantial increase in sinistrally coiled neogloboquadrinids occurs in isotope stage 64 at an age of 1.80 Ma, very close to the top of the main normal-polarity zone of the Olduvai, which is dated by these authors at 1.79 Ma. In the Singa section, Lourens et al. (1992) located that event in the same position as in the Vrica section. Sprovieri (1993) noted that the first common occurrence of left-coiling specimens of *N. pachyderma* in the Vrica section and other Mediterranean sequences is coincident with the abundance fluctuation linked to oxygen-isotope stage 64. According to Raymo et al. (1989), in DSDP sites 607, 609, and 552A (North Atlantic) the first abundant occurrence of left-coiled *N. pachyderma* is also found within isotope stage 64. The beginning of dominant sinistral coiling in *N. pachyderma* has been used as a criterion for recognizing the Pliocene–Pleistocene boundary in northern Italy in the Santerno section (Colalongo, 1968), the Rio Vendina–Crostolo section (Colalongo, Cremonini, and Sartoni, 1978), and the Po plain (Dondi and Papetti, 1968), in southern Italy in the Pisticci section of Puglie (Lentini, 1971), and in other sequences.

The first appearance of *Globigerina cariacoensis* is in the Vrica section about 2 m below the midpoint of marker bed *f*. Lourens et al. (1994) dated the midpoints of marker beds *f* and *e* at 1.738 and 1.809 Ma, respectively. By linear interpolation between the ages of these two sapropels, an age of about 1.744 Ma is obtained for the *G. cariacoensis* FA in the Vrica section, which occurs in this section within isotope stage 62. The *G. cariacoensis* FA is recorded in sediments considered basal Pleistocene in many other sections, such as the Capo Rossello–Punta Piccola section and the Monte San Nicola section in southern Sicily (Rio, Sprovieri, and Raffi, 1984), the Santerno section of northern Italy, sections in the Marche region of central Italy and sections in the Calabria region of southern Italy according to our own unpublished data, in DSDP site 132 of the Tyrrhenian Sea (Colalongo et al., 1981), and in DSDP site 125 of the Ionian Sea (Rio et al., Chapter 5, this volume).

Other planktonic foraminifera events

The following are recorded in the Vrica section at the indicated approximate stratigraphic distances above the base of the section:

197 m: first appearance of *Globigerinoides tenellus*, an event recorded in sediments assigned to the basal Pleistocene in the Caraffa di Catanzaro section of Calabria (Pasini, Selli, and Colalongo, 1977a), the Capo Rossello section (Sprovieri, 1978), the Rio Vendina–Crostolo section (Colalongo et al., 1978), DSDP site 132, and sections sampled in the Romagna and Marche regions (M.L.C., unpublished data).

196 m: last occurrence of *Globigerinoides obliquus extremus*, an event recorded in sediments considered to be basal Pleistocene in age in DSDP site 132 and in the regions of Romagna, Marche, and Calabria (M.L.C., unpublished data).

169 m: last occurrence of dextrally coiled *Neogloboquadrina atlantica*, and, at 156.5 m (ca.), first occurrence of *Globigerina digitata digitata*. According to R. Sprovieri (personal communication), those two events are recorded close to the Pliocene–Pleistocene boundary, as presently understood, in the Monte San Nicola section and in the Capo Rossello–Punta Piccola section. According to our unpublished data (M.L.C.), these two events are also recorded close to the Pliocene–Pleistocene boundary in DSDP site 132 (Tyrrhenian Sea).

65 m: first appearance of *Globorotalia oscitans*, an event recorded in sediments considered "upper Pliocene" in age (*sensu* Colalongo and Sartoni, 1979, and Iaccarino and Salvatorini, 1982) in the Capo Rossello–Punta Piccola section, the Monte San Nicola section and DSDP site 125 (Sprovieri, 1993), the Rio Vendina–Crostolo section (Colalongo et al., 1978), the Marche region (D'Onofrio, 1968), and, according to our unpublished data (M.L.C.), DSDP site 132 and in the Calabria region.

40 m: first appearance of *Globorotalia umbilicata*. The appearance of this taxon is recorded in Pliocene sediments, close to the *G. inflata* FA, in the Capo Rossello–Punta Piccola section, the Monte San Nicola section, and DSDP site 125 (Sprovieri, 1993). According to our unpublished data (M.L.C.), this relationship also occurs in DSDP site 132 and Italian land sections in the Romagna, Marche, and Calabria regions.

The ages for most of the aforementioned planktic foraminifera events in the Vrica section can be obtained from the accurate astronomical calibration of sapropels in this section made by Lourens et al. (1994). It should be noted that several of the planktic foraminifera events noted here had not been widely mentioned in earlier publications on the Mediterranean successions. That possibly was because their importance for Mediterranean biostratigraphy had not been clearly recognized, even though the species are not uncommon in most samples of the appropriate ages.

Benthic foraminifera

The benthic foraminifera assemblages of the Vrica section are also abundant and diverse and are generally very well preserved (D'Onofrio, 1981). The stratigraphic distributions of selected benthic foraminifera in the Vrica section, according to D'Onofrio, are shown in Figure 2.13. Nakagawa (1981) and Nakagawa et al. (1980) reported only the range of *Hyalinea baltica* among the benthics in the Vrica section (Figure 2.8).

According to Colalongo et al. (1984), among the benthic

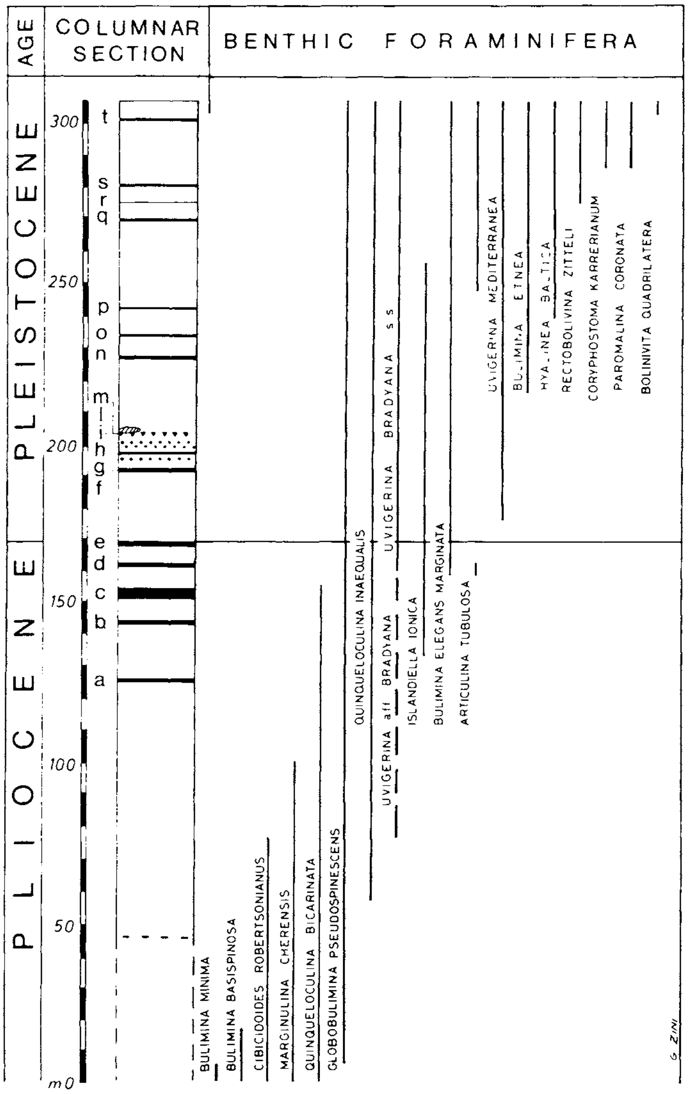

Figure 2.13. Distribution of selected benthic foraminifera in the Vrica section. (Adapted from d'Onofrio, 1981, courtesy of *Giornale di Geologia*.)

foraminifera events observed in the Vrica section, the three most important are located as follows (in meters above the base of the section, approximately located):

239 m: first appearance of *Hyalinea baltica*. In the Vrica section, this event is evident between marker beds *o* and *p*. According to Nakagawa (1981) and Verhallen (1991), the first occurrence of *H. baltica* is between marker beds *p* and *q* (Figure 2.8). This discrepancy is very probably due to the fact that whereas we collected 22 samples between marker beds *o* and *p*, the Japanese team and Verhallen collected only a few samples from this stratigraphic interval, and by accident did not find any *H. baltica* in their samples.

This first appearance of *Hyalinea baltica* is recorded in sediments considered Lower Pleistocene in age in the section at Tiepido, Emilia (Rio et al., Chapter 5, this volume), and in sections at Santerno (Colalongo, 1968), Monte Cassiano, Marche (D'Onofrio, 1968), Caraffa di Catanzaro (Pasini et al., 1977a), Capo Rossello (Sprovieri, 1978), and several other Italian sections. Furthermore, the *H. baltica* FA occurs in sediments assigned to the lower part of the Pleistocene in the Po plain of northern Italy (Dondi and Papetti, 1968) and in the Puglie region of southern Italy and in Sicily (Wezel, 1968).

The *H. baltica* FA is synchronous with the "large" *Gephyrocapsa* FA in the Vrica section (Figure 2.18) and in the Capo Rossello–Punta Piccola section (Rio et al., Chapter 5, Figure 5.5, this volume). It slightly predates this nannoplankton event in the Tiepido section and slightly postdates it in the Santerno section (Rio et al., Chapter 5, Figure 5.5, this volume). Both of these latter sections are characterized by very high sediment accumulation rates.

216 m: first appearance of *Bulimina etnea*. This event is recorded between marker beds *m* and *n*. It is also seen in sediments assigned to the basal Pleistocene at Santerno (Colalongo, 1968) and at Caraffa di Catanzaro, Calabria (Pasini et al., 1977a), and in Sicilian sections at Agrigento (Sprovieri, 1968), Monte Navone (Di Geronimo, 1969), and Capo Rossello (Sprovieri, 1978). The *B. etnea* FA occurs immediately above the *Calcidiscus macintyrei* LO in the Vrica section (Figure 2.18), but below that datum, between it and the *G. oceanica s.l.* FA, in the Capo Rossello–Punta Piccola section (R. Sprovieri, personal communication).

160 m: first appearance of *Uvigerina bradyana*. This event is recorded immediately below marker bed *d*. According to Colalongo (1968) and R. Sprovieri (personal communication), the *U. bradyana* FA approximates the Pliocene–Pleistocene boundary in the Santerno section (Romagna, northern Italy) and in the Capo Rossello–Punta Piccola section (Sicily), respectively. In the Vrica section (Figure 2.18) and in the Capo Rossello–Punta Piccola section (R. Sprovieri, unpublished data) the *U. bradyana* FA occurs between the *Discoaster brouweri* LO and the *Globigerina cariacoensis* FA.

According to Ruggieri and Sprovieri (1977), the *H. baltica* FA marks the base of the "Emilian stage," which these authors consider to be the second of three chronostratigraphic units making up the Italian Lower Pleistocene. According to Rio, Ruggieri, and Sprovieri (1982), recent advances in nannofossil biostratigraphy allow these units to be recognized, with approximate limits, in extra-Mediterranean regions. In view of the unsatisfactory exposures at this level in the Santerno River section, we have proposed (Pasini and Colalongo, 1994) to define the boundary-stratotype and GSSP for the base of the Emilian substage at a level 2 m below marker *p* in the Vrica section, where *H. baltica* is first observed.

Ostracodes

The ostracofauna of the Vrica section is very well preserved and rich in species (Colalongo and Pasini, 1980a). Figure 2.14 shows

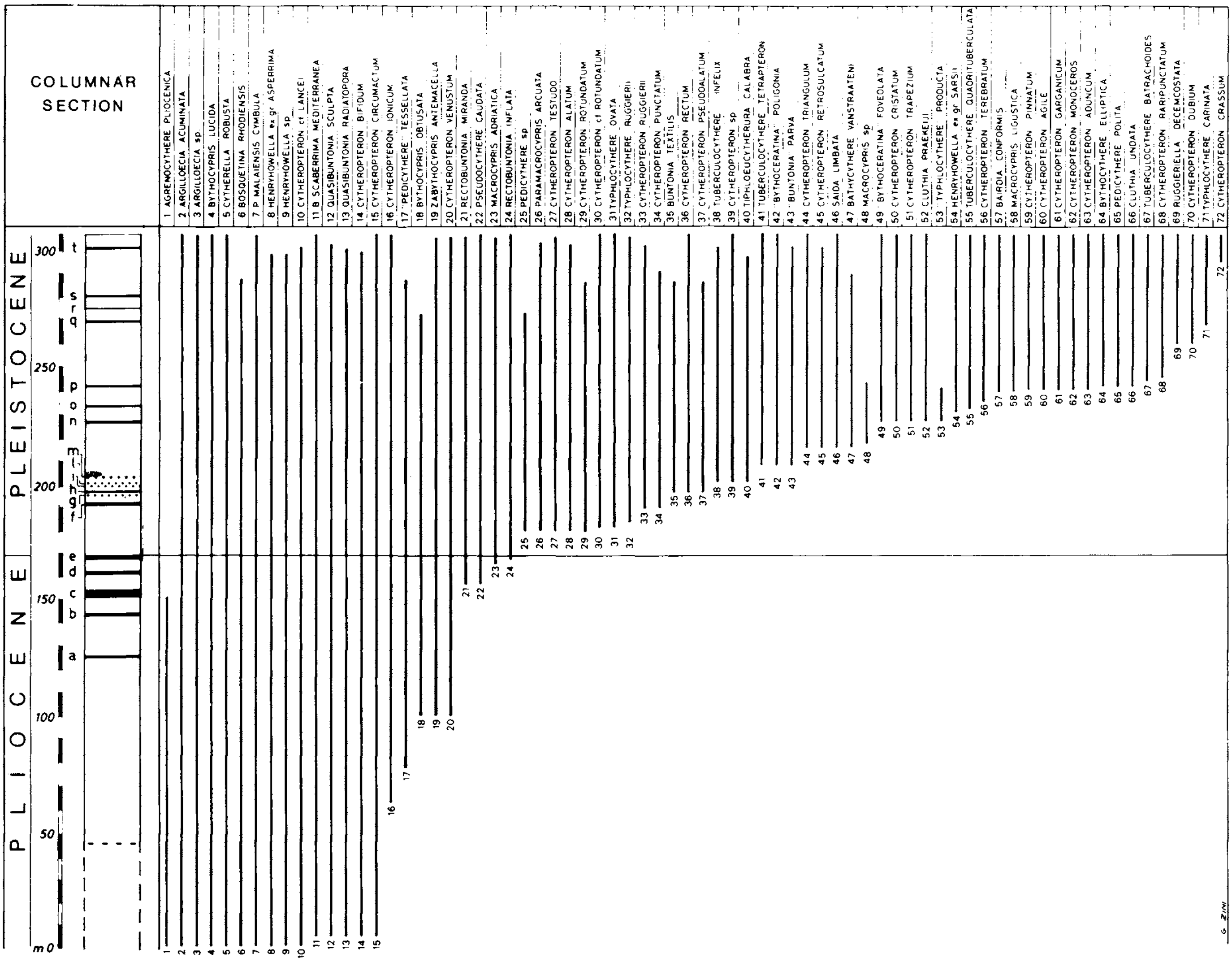

Figure 2.14. Distribution of presumed autochthonous ostracodes in the Vrica section. (Adapted from Colalongo and Pasini, 1980a, courtesy of *Giornale di Geologia*.)

the stratigraphic distributions of the ostracodes that we consider to be autochthonous, and Figure 2.15 shows the distributions of the ostracodes that we consider to be displaced from shallow-water sediments contemporaneous with the Vrica sediments (Colalongo and Pasini, 1980a). Besides the forms listed in these figures, numerous specimens of *Krithe* spp. and *Parakrithe* spp. are present throughout the section.

From Figures 2.14 and 2.15 we get the impression of a sharp change in the ostracode fauna close to marker bed *e*. Below this marker bed, the fauna consists of a comparatively small number of species that continue to the top of the section, and above this level many first appearances are recorded.

Ostracode events below marker bed e. Between the base of the Vrica section and marker bed *e*, the *Zabythocypris antemacella* FA and the *Macrocypris adriatica* FA, at about 69 m and 4 m below layer *e*, respectively, are important biostratigraphic markers. In the Italian land sections, the *Z. antemacella* FA is contemporaneous with or slightly predates the *G. inflata* FA (G.P., unpublished data); in ODP site 654 (Tyrrhenian Sea) this ostracode event is recorded a little above the *G. inflata* FA (Colalongo et al., 1990). In addition, the *M. adriatica* FA has been recognized by the present authors in Upper Pliocene sediments of the Marche region.

No ostracode extinction events are recorded in this lower interval except for that of *Agrenocythere pliocenica*, which disappears about 14 m below layer *e*. That was a local event that appears to have predated the final extinction of this species in the Mediterranean basin, because one of us (M.L.C.) has found ostracofaunas rich in *A. pliocenica* (with larval stages) in younger samples, including material from Le Castella, that contain *Globorotalia truncatulinoides excelsa* – compare Cola-

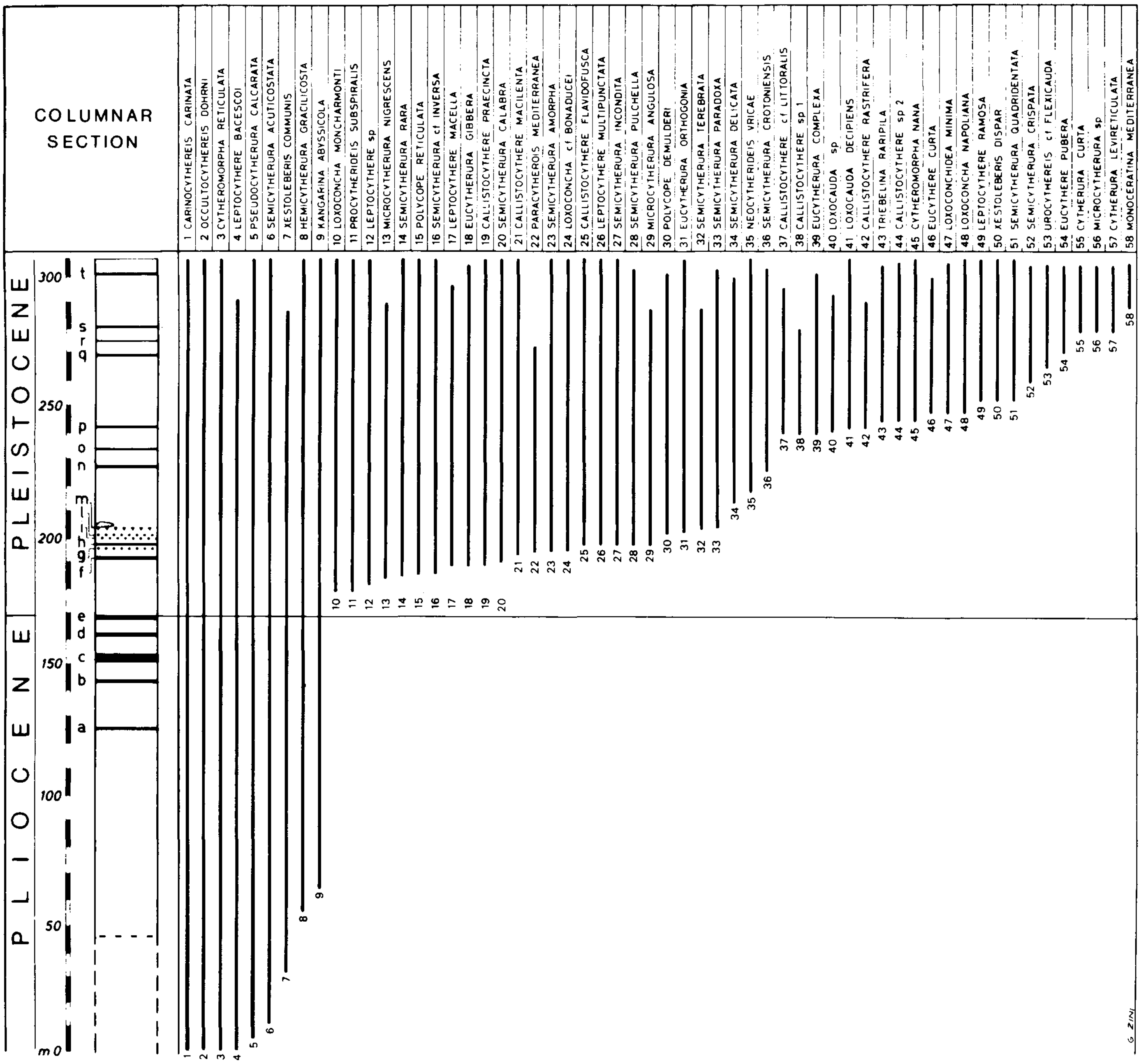

Figure 2.15. Distribution, in the Vrica section, of the ostracodes displaced from shallow-water marine sediments considered to have been contemporaneous with the sediments of the Vrica section. (Adapted from Colalongo and Pasini, 1980a, courtesy of *Giornale di Geologia*.)

longo (1965) with Pasini et al. (1977b) – and also in samples from the Marche region containing *H. baltica* (Colalongo, Nanni, and Ricci Lucchi, 1979).

Ostracode events above marker bed e. The first appearance of *Cytheropteron testudo*, considered as a "northern guest" by Ruggieri (1977a, 1980), is about 9 m above the top of marker bed *e*. Until the end of 1982 the *C. testudo* FA in Italian bathyal sediments was considered approximately coeval with the appearance of the boreal clam *Arctica islandica* in Italian shallow-marine sediments (Ruggieri, 1977b; Pelosio, Raffi, and Rio, 1980; Colalongo et al., 1981). The appearance of this famous "northern guest" in the Mediterranean has long been one of the main criteria for the beginning of the Pleistocene, and the *C. testudo* FA has therefore also been considered as a marker for the base of the Pleistocene (Colalongo and Sartoni, 1977; Pelosio et al., 1980; Colalongo and Pasini, 1980a; Colalongo et al., 1981, 1982; Pasini and Colalongo, 1982). According to more recent data, however, *C. testudo* first occurs in Sicily near the base of the Upper Pliocene (i.e., Gelasian Stage of Rio, Sprovieri, and Di Stefano, 1994).

Among the species listed in Figures 2.14 and 2.15, the following appear in the Italian region in Lower Pleistocene strata: *Cytheropteron alatum* (Colalongo et al., 1979); *C. rotundatum* and *C. punctatum* (M.L.C., unpublished data); *Buntonia textilis* (Ruggieri, 1980); *Procytherideis subspiralis* (G.P., unpublished data); *Microcytherura nigrescens* (Ruggieri, 1976); *Callistocythere praecincta* (Ciampo, 1976); *Semicytherura calabra* and *Polycope demulderi* (Ruggieri, 1980); *Callistocythere rastrifera* (Ruggieri et al., 1976); *Triebelina raripila* and *Cytheromorpha nana* (Ruggieri, 1980); *Leptocythere ramosa* (G.P., unpublished data); and *Semicytherura quadridentata* (Ruggieri, 1976). All of these species appear in the Vrica section between 9 and 82 m above marker bed *e*.

In our study of the ostracodes of the Vrica section (Colalongo and Pasini, 1980a) we erected several new species which first appear above marker bed *e*. The first occurrences of some of these species, namely *Typhlocythere ovata, T. carinata, Cytheropteron pseudoalatum, Typhloeucytherura calabra, "Bythoceratina" poligonia, Saida limbata, Cluthia praekeji, Tuberculocythere quadrituberculata, T. batrachoides, Neocytherideis vricae, Bythocythere elliptica, Pedicythere polita, Cluthia undata*, and *Ruggieriella decemcostata*, were later recognized by the present authors (unpublished data) in Lower Pleistocene sediments of different Italian regions, in some places associated with *Hyalinea baltica*. In addition, we pointed out (Colalongo and Pasini, 1980a) that *Cytheropteron garganicum, C. monoceros, Loxoconchidea minima*, and *Eucythere pubera*, previously found only in Holocene sediments, appear in the Vrica section a little above the *H. baltica* FA.

Mollusks

The stratigraphic distribution of selected mollusks in the Vrica section, according to Tampieri (in Selli et al., 1977), is shown in Figure 2.16. Among these, only *Pseudoamussium septemradiatum* seems to be a true "northern guest" (e.g., Pelosio and Raffi, 1973; Ruggieri and Sprovieri, 1977). The paleoclimatic significance of presently boreal taxa in the bathyal (psychrospheric) sequences of the Mediterranean Lower Pleistocene (such as the Vrica section) is still an open problem, however, according to S. Raffi (personal communication). In any case it must be pointed out that in the Vrica section the *P. septemradiatum* FA slightly postdates the *Gephyrocapsa oceanica s.l.* FA, which in other Italian shallow-water sections approximates the first appearance of *Arctica islandica* (Rio et al., Chapter 5, Figure 5.5, this volume).

According to the mollusk distribution scheme of Ruggieri (1962), the *Hinia turbinellus* LO and the *Turris contigua* LO seem to occur in the Italian sections close to the Pliocene–Pleistocene boundary. Ruggieri and Sprovieri (1977) note also that the extinction of *Gimnobela brevis pliorecens* occurs in Italy in the lowermost Pleistocene.

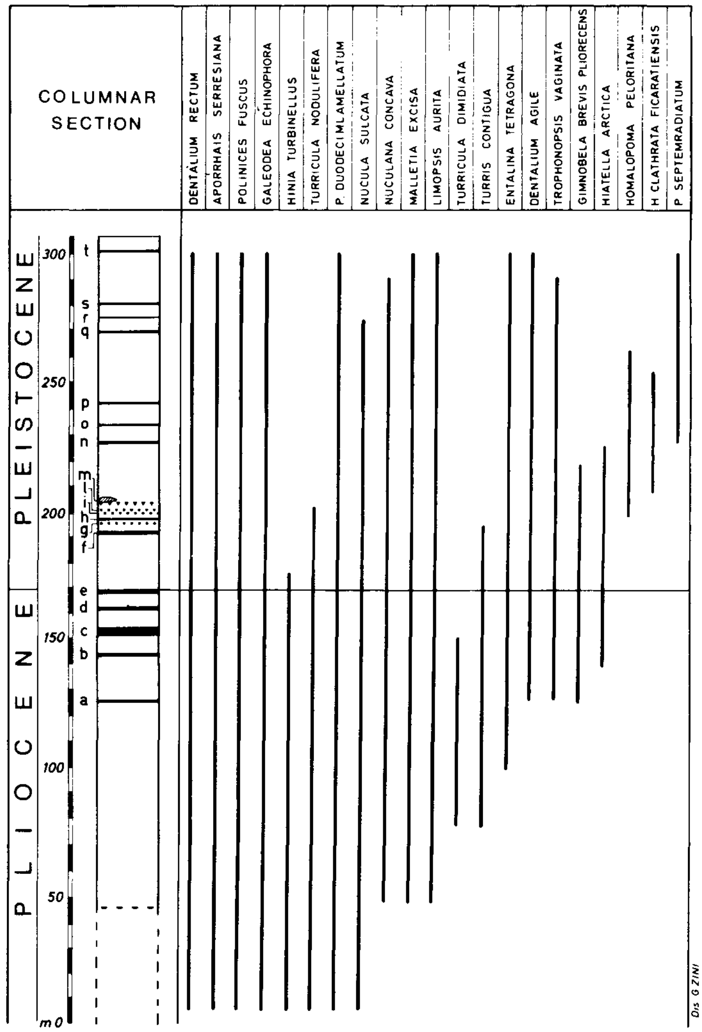

Figure 2.16. Distribution of selected mollusks in the Vrica section. (Adapted from R. Tampieri, in Selli et al., 1977, courtesy of *Giornale di Geologia*.)

Fish

The sapropel layers *b, c, d, e, f, h, n, o*, and *p* have yielded an abundant ichthyofauna of generally well preserved remains (Landini and Menesini, 1978a,b). In all, more than 700 fossil fishes have been identified to species (Table 2.1). All species survive to the present in the Mediterranean, except for *Engraulis encrasicholus macrocephalus* and *Tavania crotonensis*. Of the living forms, *Cyclothone pygmaea, Chauliodus sloanei, Hygophum hygomi, Lobianchia dofleini*, and *Lampanyctus crocodilus* live exclusively or preferentially at depths exceeding 500 m.

Pollen

According to Accorsi, Bertolani Marchetti, and Bandini Mazzanti (1978; Selli et al., 1977), the pollen of the Vrica section generally shows marked prevalence of terminocrats (Figure

Table 2.1 *Ichthyofauna of some sapropelic layers of the Vrica section*

LIST OF SPECIES	LAYERS OF ORIGIN										PRESENT GEOGR. DISTRIB.							
	b	c	d	e	f	h	n	o	p	Total sum	Med. E	Med. W	Atl. NE	Atl. NW	Atl. SE	Atl. SW	Pacif.	Ind.
Engraulis encrasicholus macrocephalus	55	8								63								
Engraulis sp.							1			1								
Cyclothone braueri	13	8	6	34	16	3	5	18	5	108	+	+	+	+				
Cyclothone pygmaea	10	11	2	52	6	3	3	15	6	108	+	+	—					
Cyclothone spp.	23	10	13	39	14	9	6	42	17	173								
Maurolicus muelleri	16	6	2	11	12	3	3	41	6	100	—	+	+	+	+	+	+	+
Vinciguerria poweriae	3									3	+	+	+	+			+	
Vinciguerria attenuata									1	1	—	+	+	+	?	?		
Ichthyococcus ovatus					1					1	+	+	+	+				
Argyropelecus hemigymnus		2			1	2	1	11		17	—	+	+	+				
Chauliodus sloanei				2						2	+	+	+	+	+		—	—
Electrona rissoi	1		1		2			4	1	9	+	+	+		+		+	—
Hygophum hygomi		1		2						3	—	+	+	+	+	+	+	+
Hygophum benoiti	1	2						1	1	5	+	+	+					
Lobianchia dofleini	5			38	10			16		69	+	+	+	+	+		—	—
Lampanyctus crocodilus			2	1	7			1		13*	—	+	+	—				
Lampanyctus pusillus	8		1	3	2					14	—	+	+	+	+	+	—	—
Ceratoscopelus maderensis	3			2	1			2	1	9	+	+	+	—				
Diaphus rafinesquei				3						3	+	+	+	?				
Diaphus holti				1						1	—	+	+					
Benthosema glaciale				1						1	+	+	+	—				
Belone belone		1					1			2	+	+	+					
Micromesistius poutassou					1					1	—	+	+	+				
Gadiculus argenteus argenteus	1	2								3	—	+	—					
Microichthys coccoi	1									1	—		?					
Stephanolepis sp.		1								1								
Tavania crotonensis	4		1							5								

Symbols: (+) species of wide geographic distribution; (−) species of restricted geographic distribution; (?) uncertain presence.

Source: Adapted from Landini and Menesini (1978a,b), courtesy of Societá di Paleontologia Italiana.

2.17). Among these, *Pinus* (as *P. diploxylon* and *P. haploxylon*) predominates, in comparison with other genera of mountain conifers (e.g., *Picea, Cedrus, Tsuga, Podocarpus*). Mediocrats are represented mainly by *Carya, Pterocarya, Corylus, Quercus, Ulmus, Zelkova, Carpinus, Tilia, Castanea, Liriodendron*, and *Liquidambar*. Taxodiaceae (*Taxodium* type) are present only in some intervals and in small percentages; they probably correspond to coastal forests. Herbaceous taxa are neither frequent nor significant. According to the aforementioned authors and to C. A. Accorsi (personal communication), the pollen diagram indicates a cool climate for the time interval represented by the sediments between the base of the section and marker bed *a*. Between marker bed *a* and marker bed *f*, mediocrat pollen species are in general much more abundant, and terminocrats scarcer, suggesting a milder climate. The sharp increase in terminocrats and the decrease in mediocrats between marker beds *f* and *h* should correspond to renewed cooling. The pollen diagram between marker bed *h* and the top of the section indicates a cool climate, with minor fluctuations.

According to Nakagawa et al. (1980; Chapter 3, this volume), the pollen show abundant conifers, especially *Pinus*, throughout the section (Figure 2.8). Pollen of broadleaf genera range from 10% to 30%. Those authors comment that the "paleoclimate is considered to be moderate to cool temperate, and a gradual cooling is suggested by the upward increase of *Abies, Picea* and *Tsuga*, and decrease of *Podocarpus*, Taxodiaceae, *Carpinus, Juglans, Pterocarya* and Ericaceae." In particular, according to those authors, it is possible to recognize unstable forest around marker bed *m* and the beginning of a slight climatic deterioration above that level.

Combourieu-Nebout, Sémah, and Djubiantono (1990) carried out detailed pollen analyses of 60 samples from the Vrica

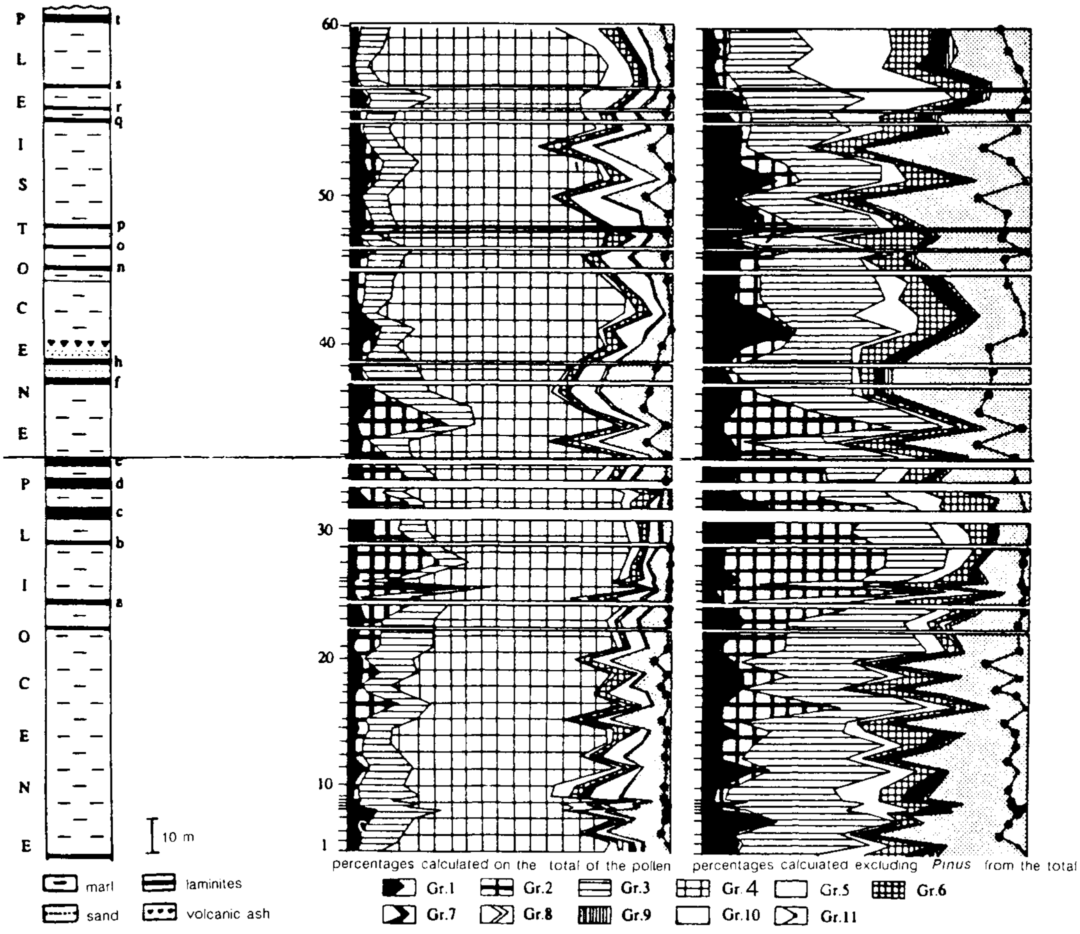

Figure 2.17. Synthesis of pollen spectra from the Vrica section, with lithology to left and pollen diagrams to right: Gr. 1, Taxodiaceae, *Engelhardtia*, Palmae, etc.; Gr. 2, *Cathaia*; Gr. 3, *Quercus*, *Carya*, *Ulmus-Zelkova*, *Carpinus*, etc; Gr. 4, *Pinus* and indeterminable Abietaceae; Gr. 5, *Tsuga*; Gr. 6, *Cedrus*; Gr. 7, *Abies* and *Picea*; Gr. 8, taxa not classified; Gr. 9, Mediterranean xerophytes (*Olea*, *Phillyrea*, *Pistacia*, etc.); Gr. 10, open-vegetation herbaceous plants; Gr. 11, *Artemisia* and *Ephedra*. (From Combourieu-Nebout et al., 1990, with permission of Academie des Sciénces de Paris.)

section, 5 of which were collected from the crucial interval between marker beds *e* and *f* (Figure 2.17). According to those authors, in the lower part of the sequence (from the bottom to sample 20, located about 15 m below marker bed *a*) there are high percentages of altitudinal elements (*Tsuga, Cedrus, Abies*, and *Picea*) and of herbaceous taxa. This association indicates a relatively cool period, which on the basis of biostratigraphic and chronostratigraphic data those authors ascribed to the "Praetiglian glacial phase" typified in northern Europe. Above sample 20, herbaceous taxa decrease, and *Cathaia* and Taxodiaceae begin to increase. From marker *a* to marker *e*, a long period dominated by forest in a subtropical to temperate–warm climate is indicated by relatively high proportions of Taxodiaceae, *Engelhardtia*, Palmae, *Cathaia, Quercus, Carya, Ulmus–Zelkova*, and *Carpinus*. This interval has been correlated by Combourieu-Nebout et al. (1990) to the "Tiglian interglacial phase." From marker bed *e* and continuing to the top of the section, the pollen spectra recorded by those authors show a steady increase in herbaceous taxa (especially *Artemisia*) and a corresponding decrease in forest taxa, with progressive enrichment of altitudinal trees. This interval, which is considered by Combourieu-Nebout et al. (1990) to have been a period of cool and xeric climate, is ascribed to the "Eburonian glacial phase." The most important point in that detailed research is that the major floristic shift that those authors identified as the transition from the "Tiglian interglacial phase" to the "Eburonian glacial phase" corresponds to the Pleistocene boundary-stratotype at the top of marker bed *e*.

K/Ar and fission-track dating

The volcanic ash layer l.a.

This layer, 20 cm thick, crops out about 65 m above the bottom of the component-section JD of the Japanese team (Nakagawa et al., Chapter 3, this volume) and at least 75 m below the base of the Vrica section as defined here (Figure 2.3). This layer is grayish in color and is characterized by the presence of hornblende, plagioclase, zircon, and glass shards (Obradovich et al., 1982).

Foraminifera assemblages examined by one of us (M.L.C.) from the siltstone immediately below and above the *l.a.* layer are typical of the *Globorotalia crassaformis* subzone (i.e., upper part of the *Globorotalia* ex gr. *G. crassaformis* zone), which corresponds to the lower portion of the Upper Pliocene (i.e., Gelasian Stage of Rio et al., 1994). In addition, the LO of dominant left-coiled *Neogloboquadrina atlantica* is found about

9 m below the *l.a.* layer. Zijderveld et al. (1991) noted the LO of *N. atlantica* (which, in our opinion, means left-coiled *N. atlantica*) in the Singa section at a level corresponding to 2.41 Ma.

Separated fractions from the *l.a.* horizon were dated by Obradovich et al. (1982) to 2.22 ±0.03 Ma (K/Ar, hornblende), 2.0 ±0.16 Ma (fission-track, zircon), and 2.2 ±0.2 Ma (fission-track, glass). Obradovich et al. (1982) also demonstrated that the age estimates previously obtained for this volcanic ash of 3.1 and 3.4 Ma, reported by Selli (1970), were incorrect.

Volcanic ash layer m

This layer (Figure 2.3) is 2–7 cm thick and crops out about 6 m above the top of sapropelic layer *h*. The *m* ash was first dated at 2.07 ±0.33 Ma by Bigazzi and Bonadonna (in Selli et al., 1977), at 2.2 ±0.2 Ma (K/Ar, hornblende) by Savelli and Mezzetti (1977) and Savelli (in Selli et al., 1977), and at 2.5 ±0.1 Ma (K/Ar, glass) by Boellstorff (1977). Obradovich et al. (1982) demonstrated that all of those age analyses were erroneous because of misidentification of samples or because of the unsuitability of the material used for dating. In re-dating the *m* ash, Obradovich et al. (1982) found two distinct fractions of older detrital biotite mixed with the primary volcanogenic biotite. The amount of detrital biotite ranged from 0.91% to 4.72% of total biotite, related inversely to the grain size, so that the coarsest fraction yielded the youngest K/Ar age: 1.99 ±0.08 Ma. Even the coarse fraction, however, was contaminated with reworked detrital biotite, and it can be concluded that the age of the *m* volcanic ash is certainly younger than 1.99 Ma. The *m* ash cannot be dated by fission-track analysis because the glass shards are unsuitable and uranium-retentive minerals such as zircon and apatite are not present. It might be possible to obtain a more precise age determination for this ash by analyzing a small sample of hand-picked large euhedral grains in a more sensitive system (J. D. Obradovich, personal communication, 1993).

Pumice block

A whole-rock K/Ar age of 2.0 ±0.1 Ma was obtained by Savelli and Mezzetti (1977; Selli et al., 1977) on the pumice block found about 1 m above volcanic ash layer *m*. It must be remembered that whole-rock K/Ar analyses of glassy materials such as pumice, especially those exposed to marine conditions, are highly unreliable (Dalrymple and Lanphere, 1969). Subsequently, however, Obradovich et al. (1982) obtained a mineral K/Ar age of 2.35 ±0.16 Ma for hornblende from the pumice block.

As noted earlier, the mineralogy and chemistry, as well as the K/Ar age, of the detached pumice block differ conspicuously from those of the autochthonous *m* ash. Obradovich et al. (1982) interpret this "as indicating that the pumice block has been reworked from an older pumice layer and thus has no relevance to the age of the *m* marker bed."

Magnetostratigraphy

In 1977, H. Nakagawa presented a preliminary version of the magnetostratigraphy of the Stuni–Vrica sequence at the joint meeting of INQUA Subcommission 1-a and the IGCP-41 working group in Birmingham, and later at the Neogene–Quaternary boundary symposium in Dushanbe (Nakagawa et al., 1980). The Japanese team subsequently reported revised polarity sequences (Nakagawa, 1981, 1982), and other measurements on the Stuni–Vrica sequence were published by Tauxe et al. (1983). Nakagawa et al. (Chapter 3, this volume) repeated their paleomagnetic sampling and analysis of the Stuni–Vrica sequence with new equipment for their final version of the magnetostratigraphy of this sequence.

The biostratigraphy of that part of the Stuni–Vrica sequence stratigraphically below the Vrica section, as defined by Selli et al. (1977) and in this chapter, has not yet been studied in sufficient detail to evaluate the physical continuity of this part of the section. For this reason, with regard to the data presented by Nakagawa and co-workers, we have taken into account only the paleomagnetic information that is inarguably within the Vrica section, in sections JA and JB (Figures 2.3 and 2.18).

Nakagawa et al. (Chapter 3, this volume) correlated the normal-polarity zone of the lower part of the Vrica section, including two short reversed-polarity intercalations separated by a short intermediate-polarity interval (Figure 2.18), with the Olduvai subchron. Their correlation is supported by the planktic microfossil events *Gephyrocapsa caribbeanica* FA, *Discoaster brouweri* LO, *Globigerinoides obliquus* LO, *Gephyrocapsa oceanica* FA, and *Calcidiscus macintyrei* LO in or near the normal-polarity zone of the Vrica section. These are events that are recorded in or near the Olduvai subchron in deep-sea sections, and to some extent in Japanese land sections of the Boso Peninsula (Itihara et al., Chapter 24, this volume).

Tauxe et al. (1983) studied almost all the Vrica section, but they omitted the lower part of component-section A because that part was synthesized from scattered outcrops, and they discarded all samples above marker bed *s* because they were judged to be unsuitable for paleomagnetic polarity measurements. According to Tauxe et al. (1983), two normal-polarity zones are recognizable in the Vrica section (Figure 2.18). The lower zone (N1–N2), interrupted by a short reversed-polarity intercalation, is located in the lower part of the section. The younger normal-polarity zone (N3) was observed immediately below marker bed *s*. Tauxe et al. (1983) correlated the N1–N2 normal-polarity interval with the Olduvai subchron because it lies between the *Discoaster brouweri* LO at its base and the *Calcidiscus macintyrei* LO above its top, as discussed earlier (Figures 2.10 and 2.18). With regard to the normal-polarity zone N3, the top of the Vrica section does not extend to levels as young as the lower boundary of the zone of "small" *Gephyrocapsa*, which is below the base of the Jaramillo subzone (Rio et al., Chapter 5, this volume) (Figures 2.2–2.4). The top of the Vrica section is therefore older than the Jaramillo, and the N3 zone, if confirmed, must represent a

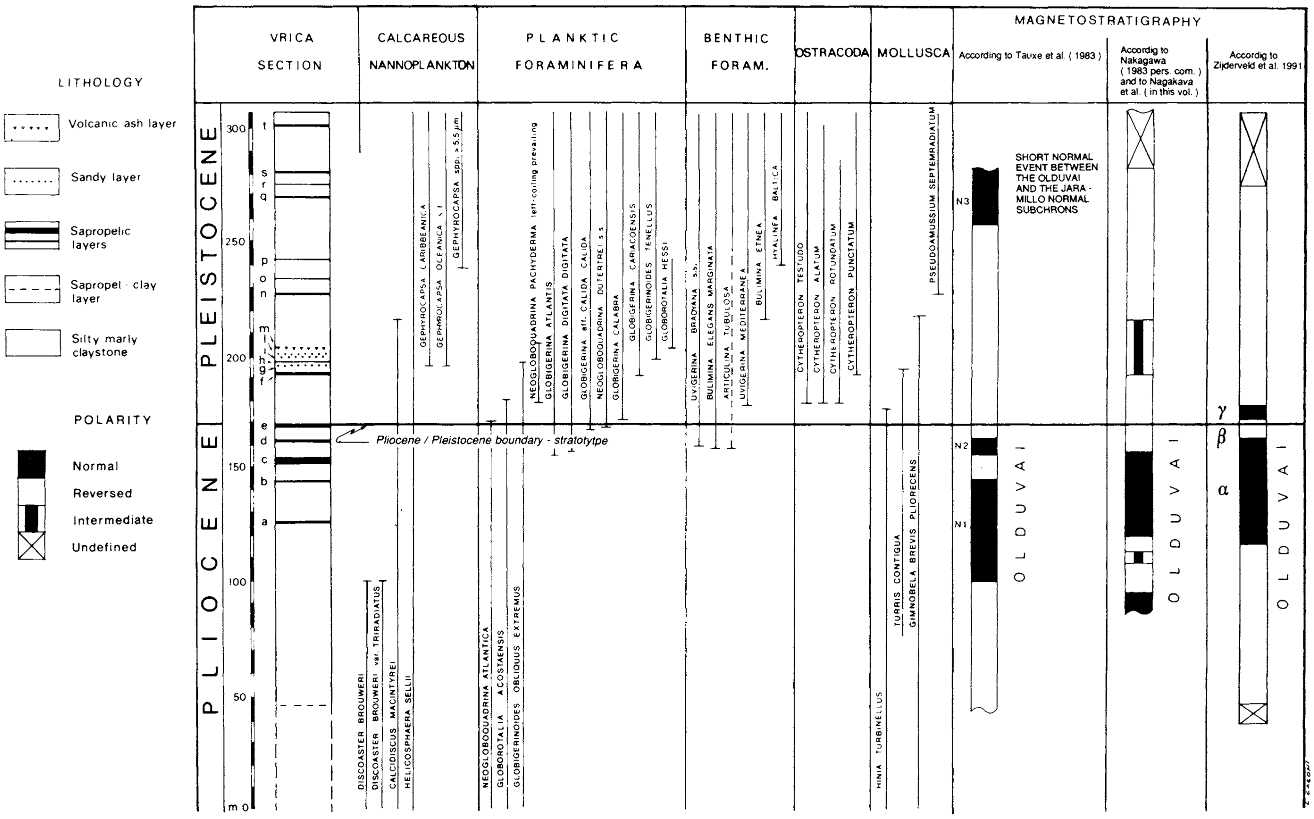

Figure 2.18. Distribution, in the Vrica section, of the most significant planktic and benthic organisms from the biostratigraphic and biochronologic points of view. To the right, magnetostratigraphy of the Vrica section according to Tauxe et al. (1983), and according to H. Nakagawa (personal communication, 1983) and Nakagawa et al. (Chapter 3, this volume). Regarding the latter paleomagnetic log, we have considered only the information from component-sections JA and JB (Figure 2.3).

normal-polarity interval between the Olduvai and the Jaramillo subchrons.

Zijderveld et al. (1991) carried out a closely spaced paleomagnetic sampling in the Vrica section, with a sampling density much higher than that of Tauxe et al. (1983), by use of a motorized corer to collect unweathered samples. The magnetostratigraphy of the Vrica section proposed by those authors is shown in Figure 2.18. According to Zijderveld et al. (1991), a 10-m-thick reversed-polarity interval, designated here as interval β, includes the Pleistocene boundary-stratotype. Interval β is located above the 45-m-thick normal-polarity interval α, and below the 5-m-thick normal-polarity interval γ, also as designated here. After reviewing the subject in detail, Zijderveld et al. (1991, p. 711) concluded that the base of the normal-polarity interval α represents the lower boundary of the Olduvai subchron, and the top of the normal-polarity interval γ above sapropel *e* "most probably represents the upper Olduvai polarity transition as generally found elsewhere." That interpretation places the short reversed-polarity interval β in the top part of the Olduvai subzone. Zijderveld et al. (1991) did not find the normal-polarity zone N3 reported by Tauxe et al. (1983) below marker *s* at the top of the Vrica section, nor in the fresh rocks of the parallel Crotone section; according to Zijderveld et al. (1991, p. 712), "given the very weathered state of the top part of the Vrica section these normal polarities are almost certainly due to secondary magnetizations related to the weathering and do not represent an extra normal subchron."

The presence of a reversed-polarity interval within the upper Olduvai had previously been recognized in the Boso Peninsula in central Japan (Nakagawa et al., 1982), in core V20–109 from the northern Pacific Ocean (Ninkovitch et al., 1966), in several deep-sea cores of the Indian Ocean (Opdyke and Glass, 1969), in loessic deposits on the Chinese Loess Plateau (Heller et al., 1991), and in other sequences.

Adoption of the Vrica section

The Vrica section was first described in the guidebook for the symposium on the Neogene/Quaternary boundary in Bologna and Crotone in October 1975 (Pasini et al., 1975). The members of the IGCP Project 41 working group, after visiting the main Plio–Pleistocene sections of Calabria, recognized that "there are some difficulties in the interpretation of the Santa Maria di Catanzaro and Le Castella sections as [Pleistocene] boundary stratotype [sections]" and that the Vrica section represented a potential Pleistocene boundary-stratotype section (Selli and Cati, 1977, pp. 29–30).

The results of further research on the Vrica section were presented in 1977 by one of us (G.P.) with C. Savelli and R. Selli (in Selli et al., 1977) and by H. Nakagawa at the joint meeting of INQUA Subcommission 1-a and the IGCP Project 41 working group during the X INQUA Congress in Birmingham. At the close of that joint meeting, a resolution was passed, emphasizing that "the participants of the Meeting support the suggestion to take the Vrica Section (Crotone in Italy) as the Neogene/Quaternary boundary stratotype [section], and find it necessary to further a detailed complex study of this section."

At the next joint meeting of the IGCP-41 working group and the INQUA Subcommission on the Pliocene–Pleistocene boundary, held in Chandigarh and Srinagar (India) in October–November 1979 (Sastry et al., 1981), it was concluded that in view of the resolution adopted by the 1948 International Geological Congress in London, a physical reference point for the Pliocene–Pleistocene boundary should be chosen in the marine sequences of southern Italy, and that "the Vrica Section in Calabria, Italy, described by Selli et al. (1977) appears more suitable [than the Santa Maria di Catanzaro and Le Castella sections for defining the Pleistocene boundary-stratotype] because it meets the general criteria enumerated above," and furthermore that "the works of Selli et al. [1977], Nakagawa et al. [1980], and Arias et al. [1980] show that the Vrica deposits contain many elements of value for wide range correlation."

Essentially the same conclusions were reaffirmed in subsequent meetings, as further work on the Vrica section continued to demonstrate its suitability for the boundary-stratotype. One of us (G.P.) presented a report describing and correlating the principal Plio–Pleistocene sections of Calabria (including the Vrica section) and DSDP site 132 (Tyrrhenian Sea) at the XXVI International Geological Congress, Paris, 1980 (Colalongo et al., 1981). At a joint meeting of INQUA Subcommission 1-a and the IGCP-41 working group during this congress, the propositions from that report were accepted, and it was resolved that "the N/Q [Neogene/Quaternary] boundary [stratotype] should be placed in the Vrica Section." Again, at the joint international field conference of IGCP Project 41 ("Neogene/Quaternary Boundary") and Project 128 ("Late Cenozoic Magnetostratigraphy") in Arizona and California, March–April 1981, a resolution was passed that "the Vrica Section in Calabria is the most suitable candidate for the P/P [Pliocene–Pleistocene] boundary stratotype," and that "paleontologic information already available from Vrica offers a sound basis for world-wide correlations in marine sections."

At the next joint meeting of the INQUA Subcommission 1-a and of the IGCP-41 working group at the XI INQUA Congress, in Moscow, August 1982, the members of these groups and outside specialists conducted a lengthy and detailed discussion of the Pliocene–Pleistocene boundary problem and the biostratigraphy, biochronology, magnetostratigraphy, paleoclimatology, radiometric ages, and sedimentology of the Vrica section. At the end of the meeting, an overwhelming majority of those present approved a formal proposal in favor of the Vrica stratotype.

The final wording of the proposal of the ICS working group on the Pliocene–Pleistocene boundary concerning the definition of the Pliocene–Pleistocene boundary-stratotype, as described in this volume, was edited during a joint meeting of the ICS working group on the Pliocene–Pleistocene boundary and the IGCP-41 working group in Madrid, May 23, 1983.

Location of the Pleistocene boundary-stratotype in the Vrica section

As is well known, the proposal by the temporary commission to advise on the question of the definition of the Pliocene–Pleistocene boundary, which was accepted as read by the council of the XVIII International Geological Congress in London, 1948, emphasized that the Pliocene–Pleistocene boundary-stratotype should be placed not only to coincide with the base of the Italian Calabrian Stage but also "at the horizon of the first indication of climatic deterioration in the Italian Neogene succession" and "should be based on changes in marine faunas."

Given the state of knowledge at the time and the composition of the temporary commission, there can be no question that the "first indication of climatic deterioration" must be understood as an allusion to the entrance of the first of the well-known "northern guests" into Mediterranean faunas. This is made doubly clear in the follow-up report organized by the Italian Geological Society at the XIX International Geological Congress, in Algiers, 1952, which was requested by the council of the London congress in order to select "a type locality for the precise definition of the boundary." In that report, four sections were described: the Monte Mario section near Rome (Blanc, Tongiorgi, and Trevisan, 1954), the Castell'Arquato section near Piacenza (Di Napoli, 1954), the Santerno section near Imola (Ruggieri, 1954), and the Val Musone section near Ancona (Selli, 1954); see also Azzaroli et al. (Chapter 11, this volume). In that report, the Pliocene–Pleistocene boundary was placed to coincide in each section with the first occurrence of "northern guests," and each section was considered to be an example of the Calabrian Stage. It is therefore the appearance of the first "northern guest" in the Italian Plio–Pleistocene sections that is the *historical* criterion used to mark the Pliocene–Pleistocene boundary, in conformity with the intention of the London 1948 resolution.

In 1977, proposals for the location of the Pliocene–Pleistocene boundary-stratotype in the Vrica section were first advanced. Selli et al. (1977) proposed to locate the Pliocene–Pleistocene boundary within "Unit Y" of the Vrica section, an informal biostratigraphic unit proposed by Selli et al. (1977) that extends from the *Cytheropteron testudo* FA (between marker beds *e* and *f*) to the *Hyalinea baltica* FA (between marker beds *o* and *p*).

Accordingly, various authors, including Colalongo and Sartoni (1977), Nakagawa et al. (1980), Colalongo and Pasini (1977a, 1980a,b), Pelosio et al. (1980), Colalongo et al. (1980, 1981, 1982), and Nakagawa (1981), have proposed that the Pleistocene boundary-stratotype should be equated with different horizons located between marker beds *e* and *m* of the Vrica section.

In regard to these proposals, Colalongo and Pasini (1980a) and Colalongo et al. (1980) recommended that the Pleistocene boundary-stratotype be placed, in any event, within the *e–m* interval because that is where the most pronounced, if not precisely coincident, faunal changes in both planktic and benthic microfossils are recorded. That recommendation was approved at the joint meeting of IGCP-41 and INQUA Subcommission 1-a at the 1980 Paris International Geological Congress.

At that point, the problem was to select a physical horizon within the *e–m* interval that could serve as the definitive Pleistocene boundary-stratotype. The 1982 joint meeting of INQUA Subcommission 1-a and IGCP-41 in Moscow recommended the physical horizon immediately below the first occurrence of the cold-adapted ostracode *Cytheropteron testudo* (between marker beds *e* and *f*), but, as noted earlier, that was soon rendered unacceptable by the discovery that this ostracode also was present in the Mediterranean basin during the middle Pliocene.

Summary of the final proposal

After discussion with members of IGCP-41 and preliminary approval at the 1968 Moscow INQUA Congress, in 1983 the ICS Working Group on the Pliocene/Pleistocene Boundary presented its final report to the parent ICS Subcommission on Quaternary Stratigraphy, recommending "*the base of the claystone conformably overlying sapropelic marker bed e of the Vrica section . . . as the Pliocene/Pleistocene boundary-stratotype*."

The document, "Proposal of the ICS Working Group on the Pliocene/Pleistocene Boundary," prepared in 1983 by E. Aguirre and G. Pasini as chairman and secretary, respectively, of the IGCP-41 working group, made three important points in support of the Vrica section as the boundary-stratotype section for the base of the Pleistocene epoch. Those points, which are even more pertinent in view of more recent developments, are reviewed and amplified as follows:

Point 1. The base of the claystone overlying marker bed *e* of the Vrica section is an appropriate location for the boundary. In fact, the base of the claystone unit is close to calcareous microplankton events such as the *Gephyrocapsa oceanica s.l.* FA, *Globigerinoides obliquus extremus* LO, the beginning of dominant left-coiled *Neogloboquadrina pachyderma*, and the *Globigerina cariacoensis* FA, which in other Italian sections (e.g., Santerno, Stirone) have been identified close to the first appearance of the shallow-water mollusk *Arctica islandica* (Colalongo, 1968; Pelosio et al., 1980; Azzaroli et al., Chapter 11, this volume; Rio et al., Chapter 5, this volume). In particular, Rio et al. (Chapter 5, this volume) state that "the stratigraphically lowest level where *A. islandica* is represented seems to be in the Castell'Arquato and the Stirone sections. No reliable nannofossil data are available from the former section. In the Stirone section [near Parma], we have observed that the appearance of *A. islandica* slightly predates the first occurrence of *G. oceanica s.l.* . . . close to the top of the Olduvai subchron." In the Vrica section, the base of the claystone overlying marker bed *e* is similarly positioned, being only 99 k.y. older than the first occurrence of *G. oceanica s.l.* ("medium-sized" *Gephyrocapsa* of Lourens et al., 1994) and close to the top of the Olduvai subzone (Figure 2.18). We may consider that the base of the claystone overlying marker bed *e* is penecontemporaneous with the first appearance of *A. islandica* in Italy.

It is well known that the entrance of *A. islandica*, the most famous among the earliest "northern guests," into the Mediterranean has historically been the main criterion for recognizing the beginning of the Pleistocene in Italy. Thus, the designation of the base of the claystone overlying layer *e* as the Pleistocene boundary-stratotype places the boundary very close to its traditional statigraphic position, thereby avoiding serious upset of the geological literature and maps, in accordance with the recommendations of the *International Stratigraphic Guide* (Hedberg, 1976; Salvador, 1994). Furthermore, the paleomagnetic reversal at the top of the Olduvai subzone, very close to bed *e*, is clearly associated with evidence of major climatic deterioration in various stratigraphic sections from around the world (as exemplified by reports in this volume).

Point 2. Multiple, reinforcing criteria offer a sound basis for worldwide correlation of the proposed boundary horizon. The main calcareous nannofossil events associated with the base of the claystone bed overlying marker bed *e* in the Vrica section, such as the *Discoaster brouweri* LO, *Gephyrocapsa oceanica s.l.* FA, and *Calcidiscus macintyrei* LO (Figure 2.18), are found in the same order and in the same position with respect to the Olduvai subzone in oceanic deep-sea sediments as well (Rio, 1982; Backman and Shackleton, 1983; Backman et al., 1983; Rio et al., Chapter 5, this volume). Similarly, the base of the claystone bed overlying marker bed *e* is close to planktic foraminifera events such as the *Globigerinoides obliquus extremus* LO and the beginning of dominant left-coiling in *Neogloboquadrina pachyderma* (Figure 2.18) that have been widely used in long-range correlations. Pollen analysis of the Vrica section (Combourieu-Nebout et al., 1990) demonstrates a clear correlation with the major Plio–Pleistocene paleoclimatic phases in the continental environments of Europe and, by extension, other continental regions. In terms of magnetostratigraphy, the base of the claystone overlying marker bed *e* is a little below the top of the Olduvai subzone, and in high-resolution sections can be even more precisely correlated within the thin, distinctive reversed-polarity interval recognized in the Vrica section by Zijderveld et al. (1991) (Figure 2.18). As for the stable isotopes, according to Lourens et al. (1994), the Pleistocene boundary-stratotype is located at oxygen-isotope stage boundary 65/64. This determination unambiguously positions the boundary-stratotype with respect to the precisely calibrated record of orbitally forced climatic variation that has now been observed in all the world's oceans and is being extended to many land sections.

Considering these data, the proposed Pleistocene boundary-stratotype can be easily identified on micropaleontological, palynological, magnetostratigraphic, and cyclostratigraphic grounds in both marine and nonmarine Plio–Pleistocene sequences. Consequently, the base of the claystone overlying marker bed *e* satisfies the basic requirement for a chronostratigraphic boundary, that is, its suitability for worldwide recognition, as recommended in the *International Stratigraphic Guide* (Hedberg, 1976; Salvador, 1994).

Point 3. Marker bed *e* and its contact with the overlying claystone are very well exposed in the Vrica section and in the surrounding area, and abundant fossils are present to facilitate correlation, again in accordance with the recommendations of the *International Stratigrahic Guide* (Hedberg, 1976; Salvador, 1994).

Approval of the final proposal by the ICS and the IUGS

The final version of the document, "Proposal of the ICS Working Group on the Pliocene–Pleistocene Boundary," was presented for discussion at the ICS business meeting in Moscow on August 13, 1984. In a postal ballot, the 25 voting members of the commission approved the proposal by a vote of 20 to 1, with 4 abstentions. Following a report on behalf of the ICS by M. G. Bassett and J. W. Cowie to a meeting of the full IUGS Executive Committee in Rabat, Morocco, on February 10, 1985, the proposal was submitted in writing to the IUGS Executive for a postal ballot. On May 31, 1985, the secretary-general of IUGS, R. Sinding-Larsen, informed the ICS that the proposal had received majority support from the IUGS Executive. That announcement of support represented formal ratification of the proposal by the IUGS (Bassett, 1985, p. 91).

Dating the boundary

In order to calculate the age of the Pleistocene boundary-stratotype, we take into account only the astronomically calibrated ages of the two boundaries of the Olduvai subzone. In the time scale of Shackleton et al. (1990), the top of the Olduvai subchron is placed in isotope stage 63, dated to 1.77 Ma, and the lower boundary of the subchron in isotope stage 71, dated to 1.95 Ma. According to Hilgen (1991), the top of the Olduvai falls within stage 64 and is dated to 1.79 Ma, and the base is dated to 1.95 Ma. Tiedemann, Sarnthein, and Shackleton (1994) recorded the upper boundary of the Olduvai in isotope stage 63, dated to 1.78 Ma, and the lower boundary in stage 71, dated to 1.94 Ma. Lourens et al. (1994) located the upper boundary within oxygen-isotope stage 64, dated to 1.79 Ma, and the base within stage 72, dated to 1.94 Ma. In the time scale of Shackleton et al. (1995a), the upper and lower boundaries of the Olduvai subchron are dated to 1.770 Ma and 1.950 Ma, respectively. The age values of Shackleton et al. (1995a) were also adopted by Cande and Kent (1995) in the new GPTS (geomagnetic polarity time scale). We have taken into account the magnetostratigraphy of Zijderveld et al. (1991) (Figure 2.18) and have calculated the age of the base of the claystone conformably overlying marker bed *e*, the Pleistocene boundary-stratotype, by linear interpolation between the ages of the bottom and the top of the Olduvai subzone (Hilgen, 1991). With reference to the ages of these boundaries proposed by Shackleton et al. (1990), Hilgen (1991), Tiedemann et al. (1994), Lourens et al. (1994), and Shackleton et al. (1995a), we obtain for the Pleistocene boundary-stratotype the respective ages of 1.796, 1.813, 1.803, 1.811, and 1.796 Ma; the last value is the same as in the revised GPTS of Cande and

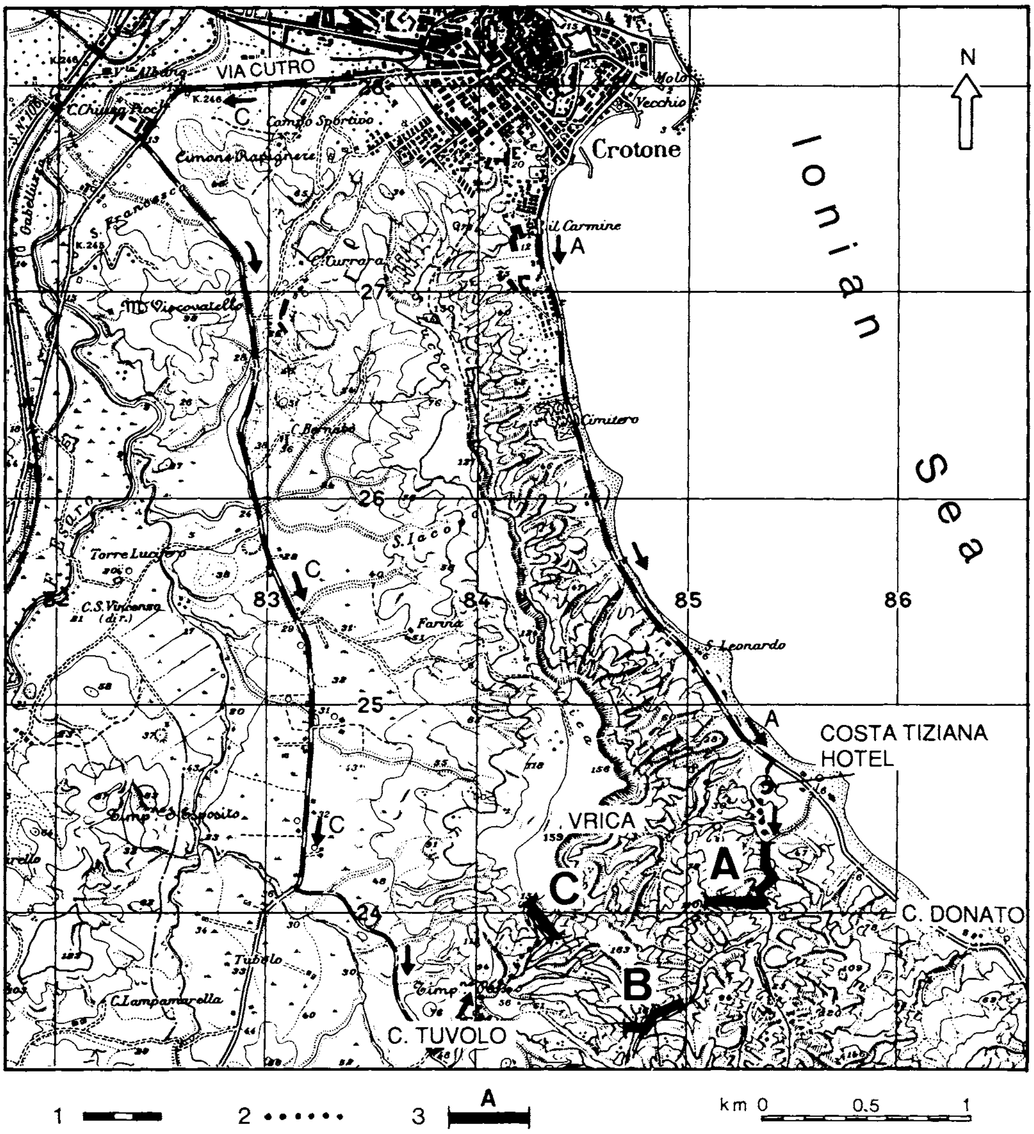

Figure 2.19. Field-guide itineraries for component-sections A and C of the Vrica section (Figures 2.2 and 2.3), indicated by arrows marked with *A* and *C*, respectively; 1, roads suitable for normal cars; 2, impassable to cars; 3, the component-sections. (Reproduced with permission of the Istituto Geografico Militare, authorization no. 2568 of April 1, 1987. Dai tipi dell'Istituto Geografico Militare, autorizzazione n. 2568 in data 1/4/1987.)

Kent (1995). All of these ages are the same, at 1.8 Ma, when rounded up to one decimal place.

Appendix: accessibility of the Vrica section

The Vrica section is located about 4 km south of the town of Crotone (Figures 2.1 and 2.19).

The component-sections A, B, and C in the Vrica section (Figures 2.2 and 2.3) can be easily reached by automobile from Crotone. Good boots are the only climbing equipment required to traverse the exposures. The outcrops of the Vrica section are being granted the status of nature preserves, with legal protection against intentional damage.

Component-section A (Figure 2.19)

This section can be reached from the parking lot at the Costa Tiziana Hotel, 3 km south of Crotone. This hotel is located a few tens of meters west (inland) of the coastal road, via a short driveway.

The base of component-section A is located about 250 m south of the hotel, at the confluence of two normally dry ravines. The trail from the hotel to this point is not well marked but may be followed easily with the aid of the map shown in Figure 2.19.

The traverse from the base of component-section A to the "sapropel–clay layer" (Figure 2.3) follows up the valley that extends southward from this confluence. The stratigraphy of this lower part, measuring about 45 m in thickness, has been reconstructed by geometric and lithologic correlations of scattered outcrops on both sides of the valley. About 100 m south-southwest of the base of the section is a small earthen check-dam, and some 200 m farther, on the west side of the seasonally flooded plain behind the dam, is the mouth of an east–west gully (Figure 2.5). Just at the mouth of the gully, about 2 m above the plain, a clearly laminated shale layer about 30 cm thick crops

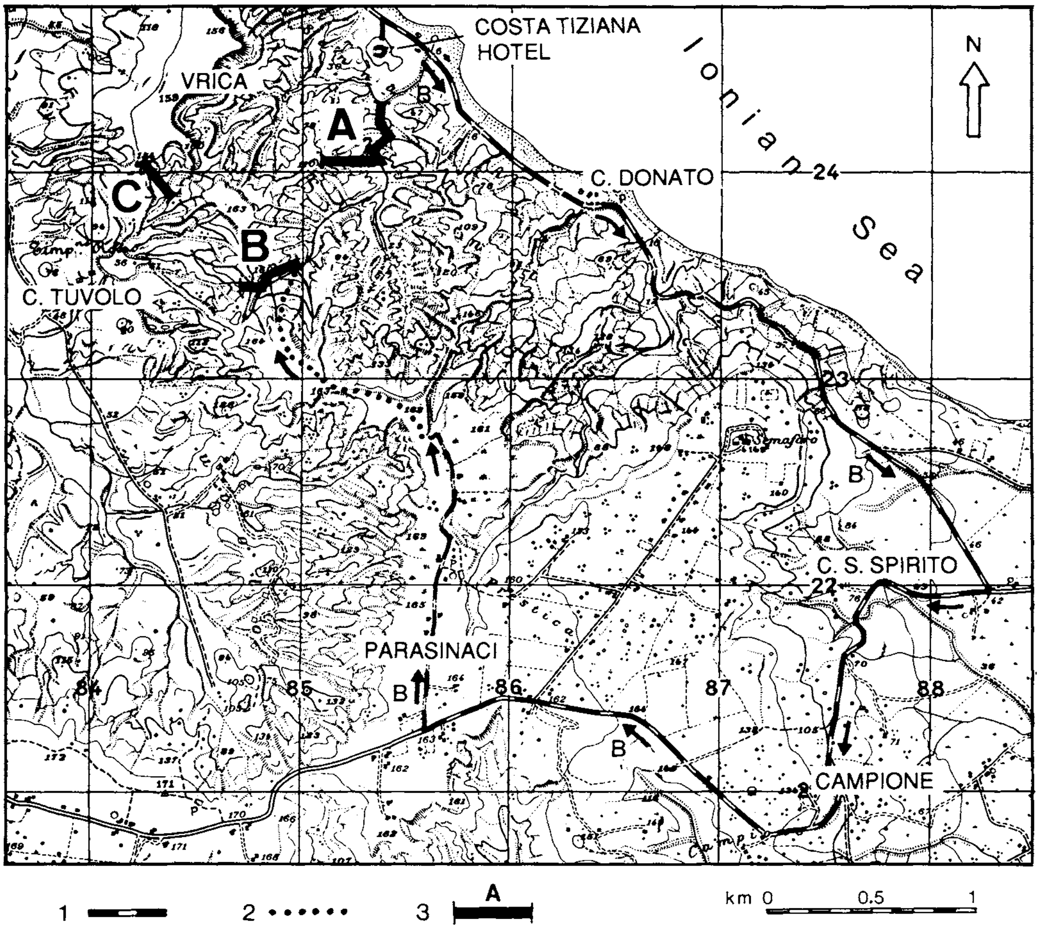

Figure 2.20. Field-guide itinerary for component-section B of the Vrica section (Figures 2.2 and 2.3), indicated by arrows marked with *B*. The road from Crotone to the Costa Tiziana Hotel is shown in Figure 2.19; symbols as in Figure 2.19. (Reproduced with permission of the Istituto Geografico Militare, authorization no. 2568 of April 1, 1987. Dai tipi dell'Istituto Geografico Militare, autorizzazione n. 2568 in data 1/4/1987.)

out. This is the "sapropel–clay layer," 45 m above the base of the Vrica section (Figure 2.3). This shale layer is normally deeply weathered and not conspicuous, but it can easily be located with the help of a rock hammer.

From this point, the section is followed westward up the side gully for several tens of meters. The first conspicuous sapropelic shale layer, cropping out near the head of the gully, is marker bed *a* (Figure 2.3).

By climbing to the ridge above the gully (Figure 2.5) and following the ridge line to the southwest for a few tens of meters, the sapropelic marker beds *b*, *c*, *d*, and *e* can be recognized according to the thicknesses of the respective layers and the interbedded claystones, as described previously (Figure 2.3).

Component-section B (Figure 2.20)

This section may be reached by continuing past the mouth of the east–west gully in which the upper portion of section A crops out and proceeding south-southwest up the dry ravine that feeds into the impounded plain.

An easier way, however (Figure 2.20), is to continue driving past the Costa Tiziana Hotel and Cà Donato on the coastal road, to a crossroads about 8 km from Crotone. There, turn right and drive past Cà Santo Spirito and Campione, and about 2.5 km past Campione turn right again and go about 1.5 km to the end of the country road that crosses the locality Parasinaci. Park and walk northwest along the northern or northeastern margin of marine terrace remnants. About 1 km northwest of the end of the country road a northeast-trending ravine cuts sharply into the terrace surface; the cliff on the opposite side of this ravine exposes component-section B (Figure 2.6). To reach the base of component-section B, go downhill (north-northeast) along a gentle slope, as far as the normally dry bed of the ravine.

Marker bed *a*, at the base of component-section B (Figure 2.3), crops out in the ravine near a curve about 40 m downstream from a major confluence. In some seasons it may be covered by alluvium and will not be easy to find. All the other marker beds of component-section B, however, are always clearly recognizable and can be identified (as outlined in the earlier section "Sedimentology and paleoecology") by their thicknesses and internal spacing (Figure 2.3). To traverse component-section B starting from the base at marker *a*, follow the ravine upstream to the point where it crosses the tuff layer of marker bed *m*. From this point onward, the section must be traversed on the steep western slope of the ravine. The uppermost recognizable marker bed is sapropelic layer *p*, cropping out a few meters below the crest of the the slope.

Component-section C (Figure 2.19)

This section can be reached from the top of component-section B by following the map shown in Figure 2.19. An easier route is to go by car on the Via Cutro from the center of Crotone (Figure 2.19), turning left about 1.8 km from town on a country road that is about 300 m beyond the first leftward curve. Follow this road for 3.8 km, as far as a gentle bend to the right some 100 m before reaching a bridge. Immediately beyond the bend, turn left on the road to Cà Tuvolo (Figure 2.19), a little house near the road. At Cà Tuvolo turn again to the left and follow an uneven road for about 300 m, to the base of an earth dam. From this point walk along the western edge of the pond behind the dam to its northernmost point (Figure 2.19), where a gully and a small valley oriented northeast–southwest are found. About 300 m up this valley from the edge of the pond, a gully in the western side of the valley exposes component-section C (Figure 2.7).

Marker bed *o*, at the base of component-section C (Figure 2.3), crops out a few meters above the point where the gully joins the main streambed. Within the gully, all the other marker beds can be recognized, with the marker bed *t* a few meters below the upper end.

The topmost part of the continuous Plio–Pleistocene marine sequence consists of about 2 m of grayish claystones, overlying marker bed *t*. Weathered red sands (not shown in the columnar sections of this chapter) overlie these claystones. These sands are part of a marine terrace, a large remnant of which is preserved at the top of the gully.

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Postscriptum

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3

The magnetostratigraphy of the Vrica section, Italy, and its correlation with the Plio–Pleistocene of the Boso Peninsula, Japan

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Introduction

During the XI INQUA Congress in Moscow in 1982, the INQUA Subcommission 1-d, "Pliocene–Pleistocene Boundary," and the working group of IGCP Project 41, "Neogene/Quaternary Boundary," jointly proposed that the section at Vrica in Calabria, southern Italy, be adopted as the Pliocene–Pleistocene boundary-stratotype. With the acceptance of that proposal by the IUGS International Commission on Stratigraphy (Nikiforova and Alekseev, Chapter 1, this volume), the type horizon designated at Vrica has been extended to Plio–Pleistocene sections around the world.

On the other side of the globe from Vrica, many marine sedimentary sections that are continuous through the Pliocene–Pleistocene transition are well exposed along the Pacific coast. Among these the Boso Peninsula in central Japan is one of the best places to study the Pliocene–Pleistocene transition (Itihara et al., Chapter 24, this volume). In recent years we have worked to correlate the Plio–Pleistocene sections of Calabria with those of the Boso Peninsula on the basis of geomagnetic reversals and planktonic microfossil events. We began our study of the Vrica section in 1976, shortly after the initial reports from the Italian team led by Raimondo Selli (1975; Selli et al., 1977). In 1977, in meetings at Birmingham, England, and at Dushanbe, Tadjikistan, USSR, we responded to the urgent request of the IGCP-41 working group and reported on a preliminary study of the paleomagnetic polarity sequence of the Vrica section, although the sampling interval had been inadequate to satisfactorily resolve the magnetostratigraphy (Nakagawa et al., 1980). Subsequently, we reported a revised Calabrian Plio–Pleistocene polarity sequence at a field conference on the Neogene–Quaternary boundary held at Chandigarh, India, in 1979 (Nakagawa, 1981) and in a symposium on the lower boundary of the Quaternary, held during the XI INQUA Congress in Moscow in 1982 (Nakagawa, 1982).

Since that time, we have repeated our paleomagnetic analyses with new equipment, having revisited Vrica in 1983 for further field work and sampling. As a result we were able to improve significantly on our preliminary reports, particularly in regard to biostratigraphic and lithostratigraphic correlations to the paleomagnetic record. The magnetostratigraphic data presented here (Figures 3.4, 3.5, 3.7, 3.8, 3.10) are calibrated according to the time scale of Berggren, Kent, and Van Couvering (1985); the same data using orbitally tuned calibration (Zijderveld et al., 1991) give ages approximately 6% older.

Vrica section

In the vicinity of the Vrica triangulation station south of Crotone, in the type area of the classic Calabrian Stage (Figure 3.1), fossiliferous marine Plio–Pleistocene strata crop out along the valleys cut into the "Milazzian" and younger terraces and on the sea-cliffs (Pasini and Colalongo, Chapter 2, this volume). Figure 3.2 shows the current status of our field observations in this area, with the locations and horizons of samples taken for laboratory analysis of paleomagnetism and microfossils.

The section consists mostly of clayey siltstone, but includes many layers of sand and sapropelic clays and a few layers of volcanic ash, which are useful marker horizons. Some of these layers are interbedded at short intervals, forming distinctive striped zones of which the intervals defined by markers *S9–S9t*, *S4–S8*, *S1–S3*, *b–e*, *f′–m2*, and *q–t* are the most prominent. The proposed boundary-stratotype horizon is at the base of the shale layer overlying sapropelic marker layer *e*, in the notation of Pasini and Colalongo (Chapter 2, this volume). The distinctive stratigraphy of the *b–e* and *f′–m2* zones helps to locate the boundary in the field. Principal distinguishing characteristics include the fact that the distances between markers *b, c, d,* and *e* are almost equal, that *b* and *d* are bluish gray and *c* and *d* are yellow from a distance, and that *f* and *h* are gray and thicker than the other layers in the zone *f′–m2*.

The geologic structure of the Vrica section is simple (Figures 3.2 and 3.3). In general the beds dip gently, about 10° west-southwest. A normal fault strikes N40W near sample localities 31 and 47, with a downthrow to the west of about 10 m. The section at Vrica lies on the eastern limb of a syncline, the western limb of which exposes the Plio–Pleistocene section at Le Castella, about 20 km farther to the south.

Figure 3.4 shows the geomagnetic polarity sequence of the Vrica section together with magnetic measurements of the

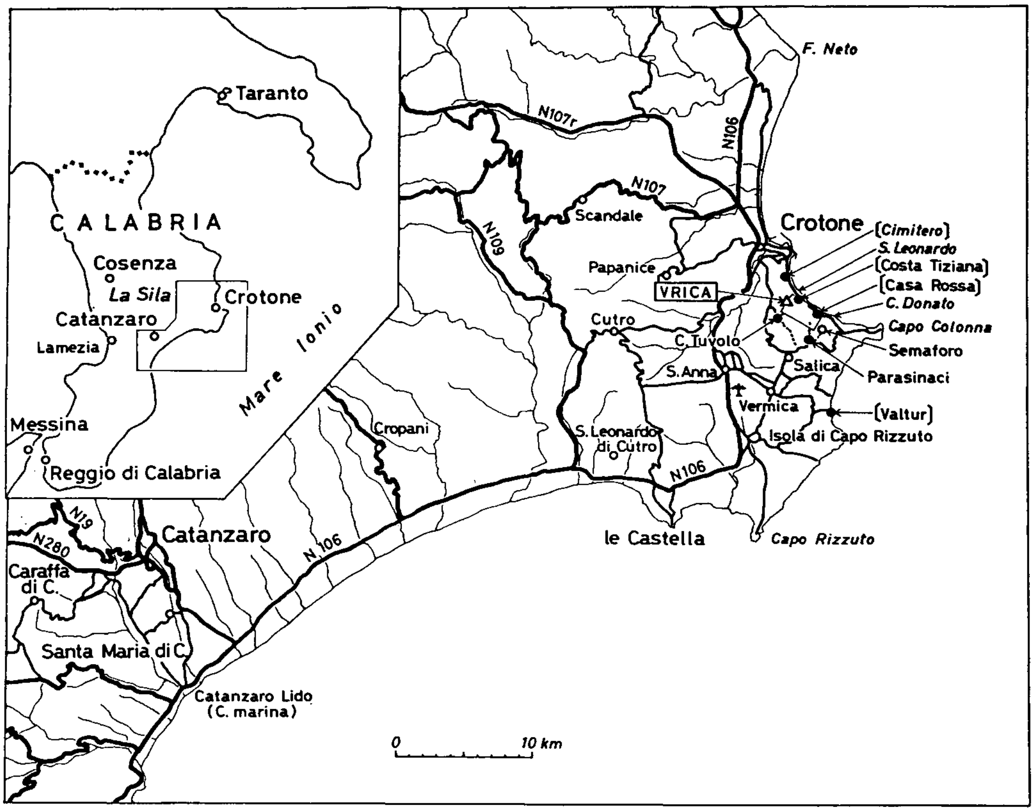


Figure 3.1. Index map of Calabria.

samples. After initial magnetic examination, we applied thermal demagnetization at 200°C and alternating field demagnetization at 15 mT to all samples. Intensity and direction of the remanent magnetization were measured with a ring-core magnetometer. The polarity sequence is based on the six most reliable measurements at each sampled horizon. Stable reversed polarity is dominant in the section, but a normal-polarity zone with a short reversed intercalation occurs in the middle part. Intermediate polarity values are found in a horizon in the lower part of the section and in a thicker zone in the upper part. The normal-polarity zone is correlated with the Olduvai normal-polarity subchron by the fossil evidence described later; additional normal polarities, observed in a thin zone just above this by Zijderveld et al. (1991), are probably also part of the Olduvai as observed in more condensed marine sections.

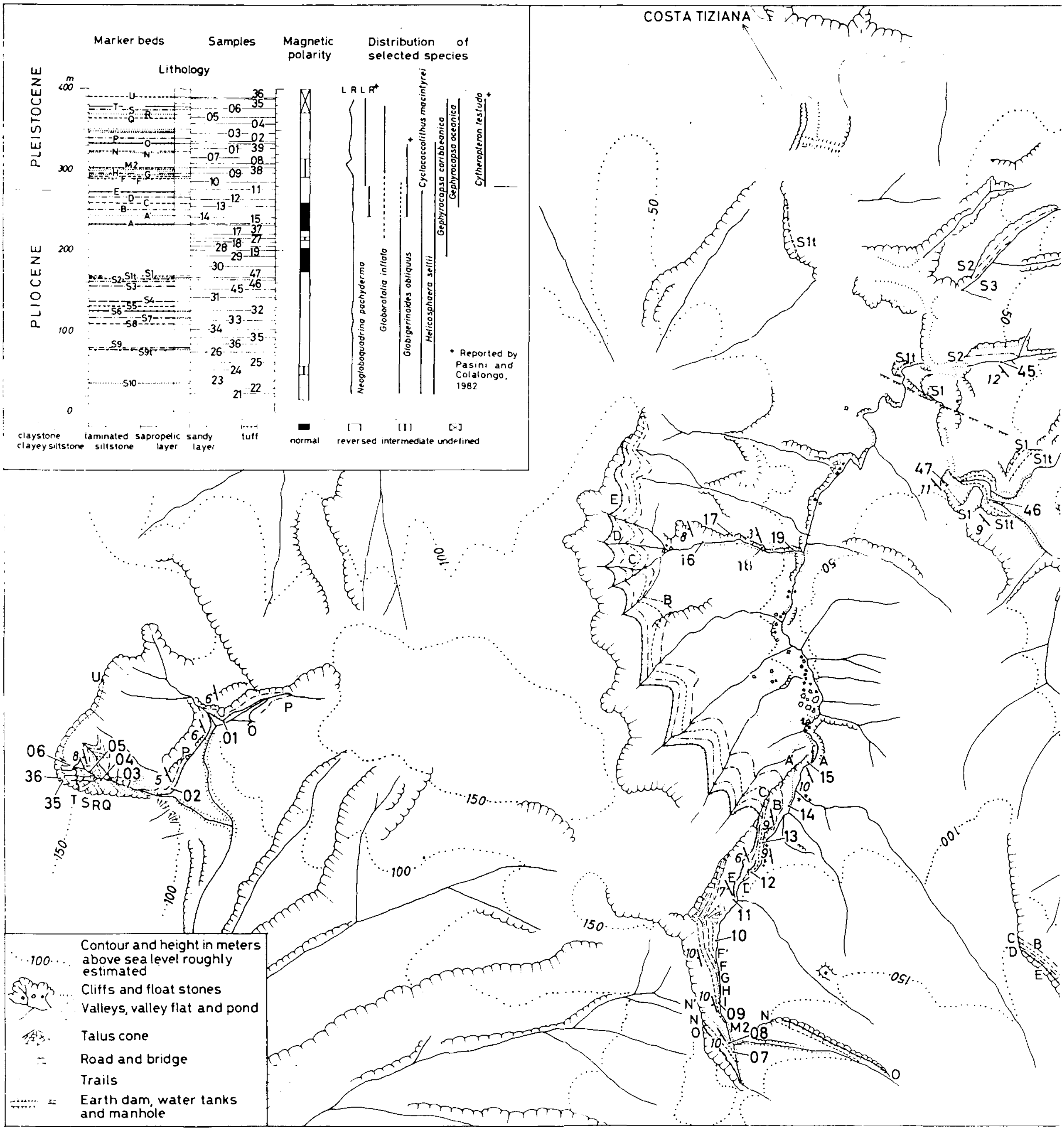
Figure 3.5 shows the stratigraphic distribution of selected planktonic microfossils, pollen, and spore grains in the same section. The index horizons for regional correlation are (1) the lowest horizon of *Gephyrocapsa aperta*, (2) the lowest horizon of *Gephyrocapsa caribbeanica*, (3) the highest horizon of *Globigerinoides obliquus*, (4) the highest horizon of *Cyclococcolithus* [= *Calcidiscus*] *macintyrei*, (5) the horizon of D-S (dextral-to-sinistral) coiling change in *Neogloboquadrina pachyderma*, and (6) the highest horizon of *Helicosphaera sellii*.

No remarkable changes are recognized in pollen flora, but increases in *Abies*, *Picea*, and *Tsuga* and simultaneous decreases in *Podocarpus* and Taxodiaceae in the upper part indicate gradual cooling, which coincides with the cooling of sea water indicated by the coiling change in *N. pachyderma*.

Colalongo et al. (1981, 1982) and Pasini and Colalongo (1982) reported the appearance of the "cold guest" *Cytheropteron testudo* between the sapropelic layers *e* and *f* (8 m above *e* and 12 m below *f*, near sample localities 10–12). Pasini and Colalongo (Chapter 2, this volume) noted that although this event is no longer considered as a reliable index to the moment at which a glacial climate first affected the Mediterranean, there are further reasons to correlate this approximate level at Vrica to the initiation of Pleistocene conditions in temperate latitudes. Therefore, the recommendations of INQUA Subcommission 1-d and the IGCP-41 working group that designated the Pliocene–Pleistocene boundary at the top of the marker layer *e* at Vrica (Nikiforova and Alekseev, Chapter 1, this volume; Pasini and Colalongo, Chapter 2, this volume) remain valid.

Boso Peninsula

The Boso Peninsula is situated to the southeast of Tokyo (Figure 3.6). A fossiliferous marine Plio–Pleistocene section is well exposed in the river valleys cut into the uplands and on the sea-cliffs in the central to northern part of the peninsula. The sediments are mostly sandstone and siltstone, interbedded with many layers of volcanic ash, which are valuable correlation markers (Figure 3.7) (Itihara et al., Chapter 24, this volume).


COSTA TIZIANA
Marker beds
Samples
Magnetic polarity
Distribution of selected species
Lithology
PLEISTOCENE
PLIOCENE
m
400
300
200
100
0
L R L R*
Neogloboquadrina pachyderma
Globorotalia inflata
Globigerinoides obliquus
Helicosphaera sellii
Cyclococcolithus macintyrei
Gephyrocapsa caribbeanica
Gephyrocapsa oceanica
Cytheropteron testudo
* Reported by Pasini and Colalongo, 1982
claystone
clayey siltstone
laminated siltstone
sapropelic layer
sandy layer
tuff
normal
reversed
intermediate
undefined
Contour and height in meters above sea level roughly estimated
Cliffs and float stones
Valleys, valley flat and pond
Talus cone
Road and bridge
Trails
Earth dam, water tanks and manhole
50
100
150
S1
S1t
S2
S3
45
46
47
17
19
16
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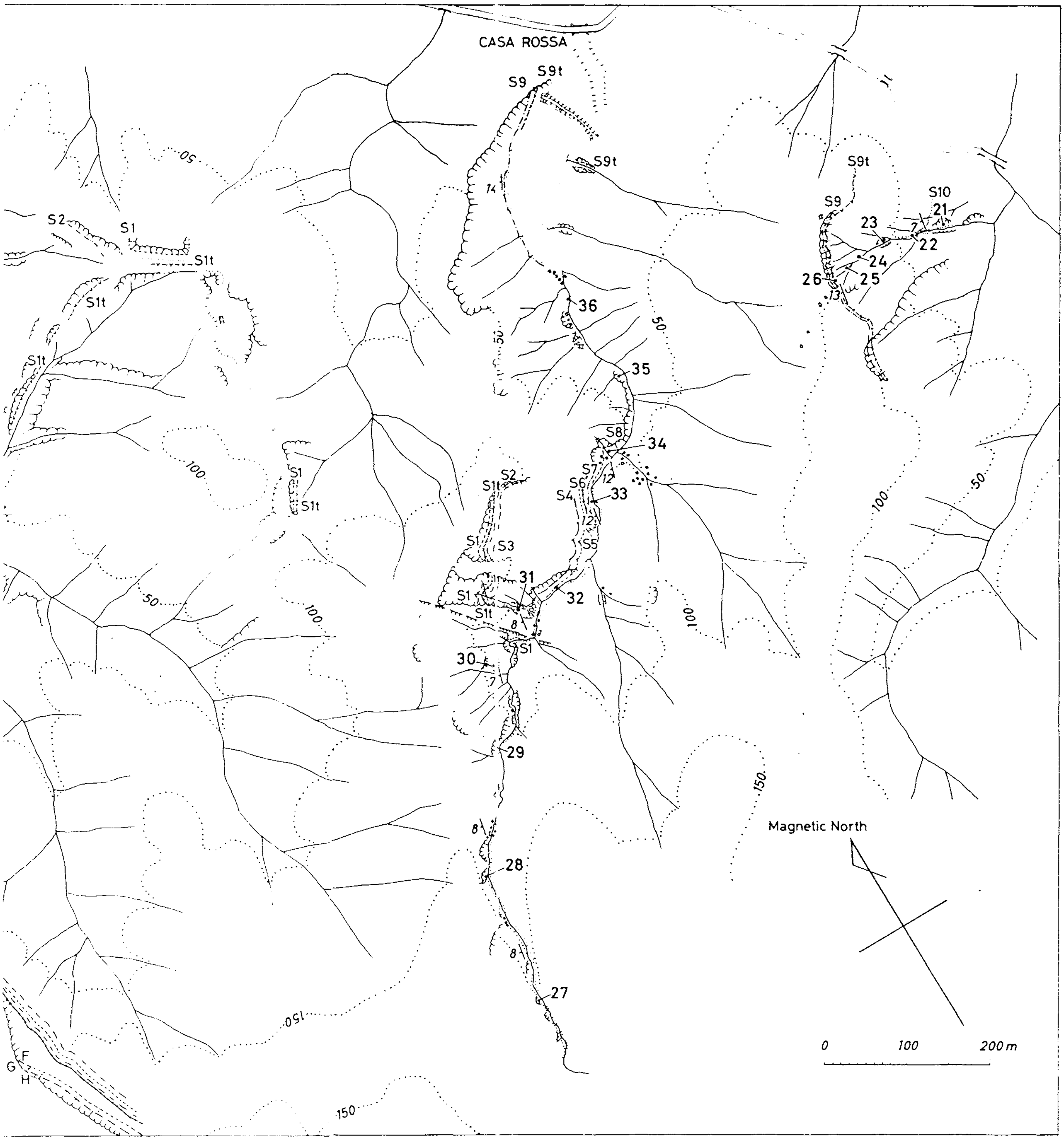
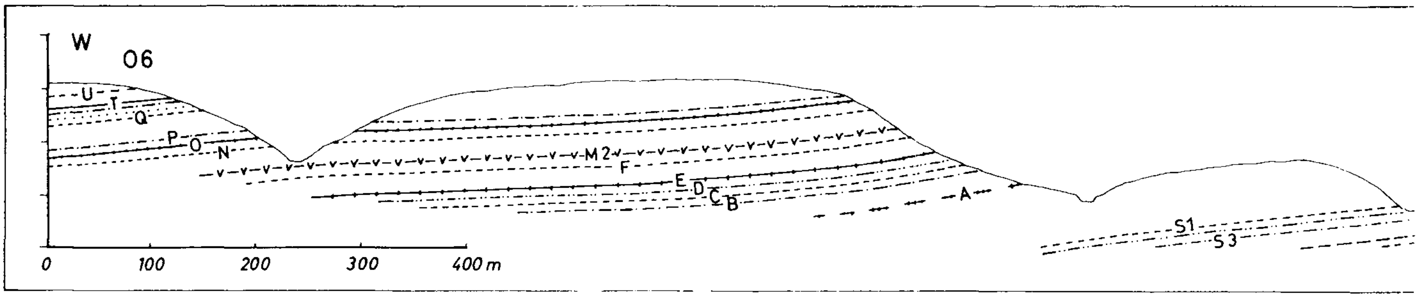


Figure 3.2. Geologic route map and columnar section for the Vrica area. The marker horizons *a–u* are indicated by capital letters *A–U* for clarity (see also Figures 3.4 and 3.5). Neogene–Quarternary boundary stratotype.



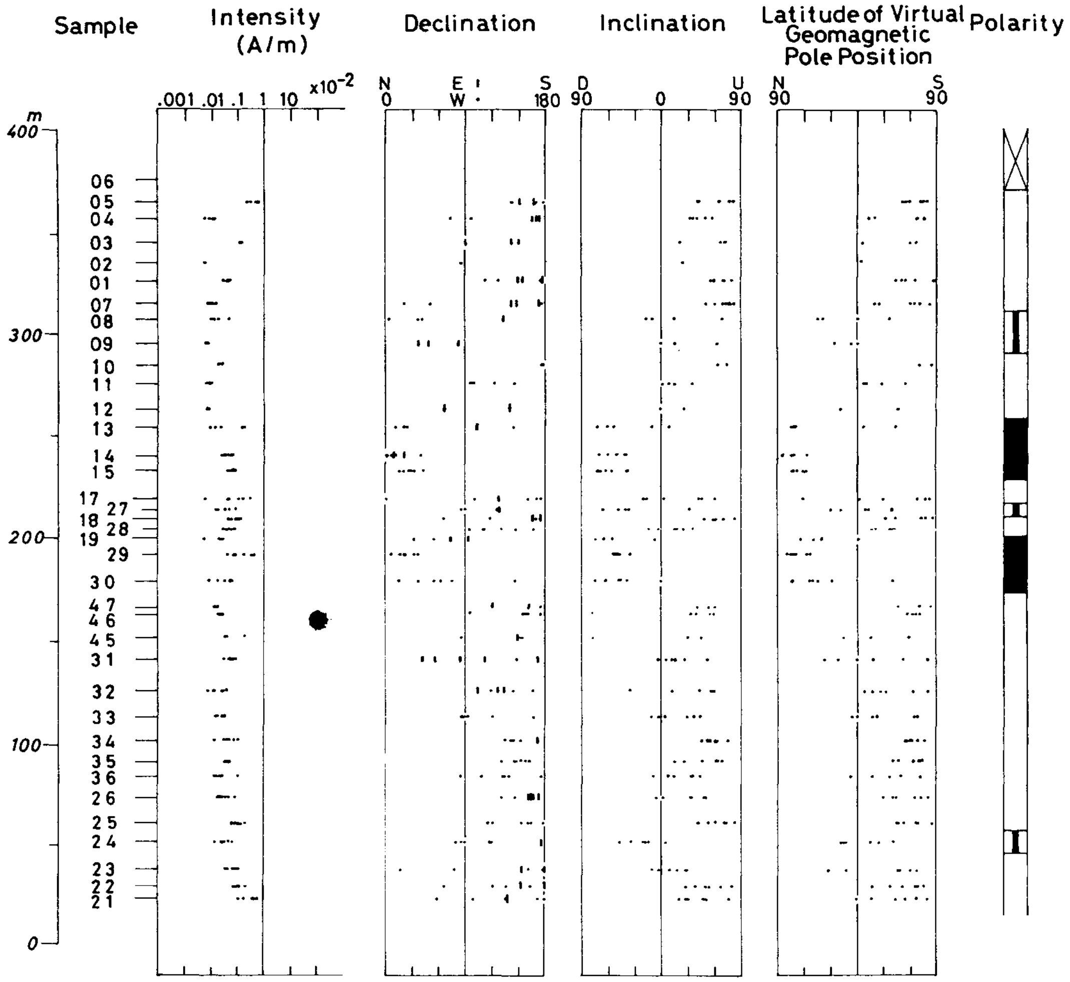


Figure 3.4. Magnetostratigraphy of the Vrica section.

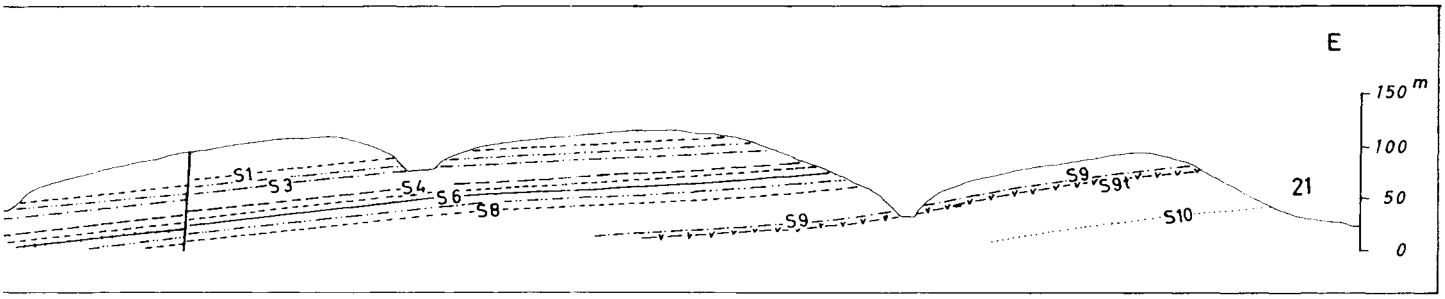


Figure 3.3. Projected cross section of the Vrica area.

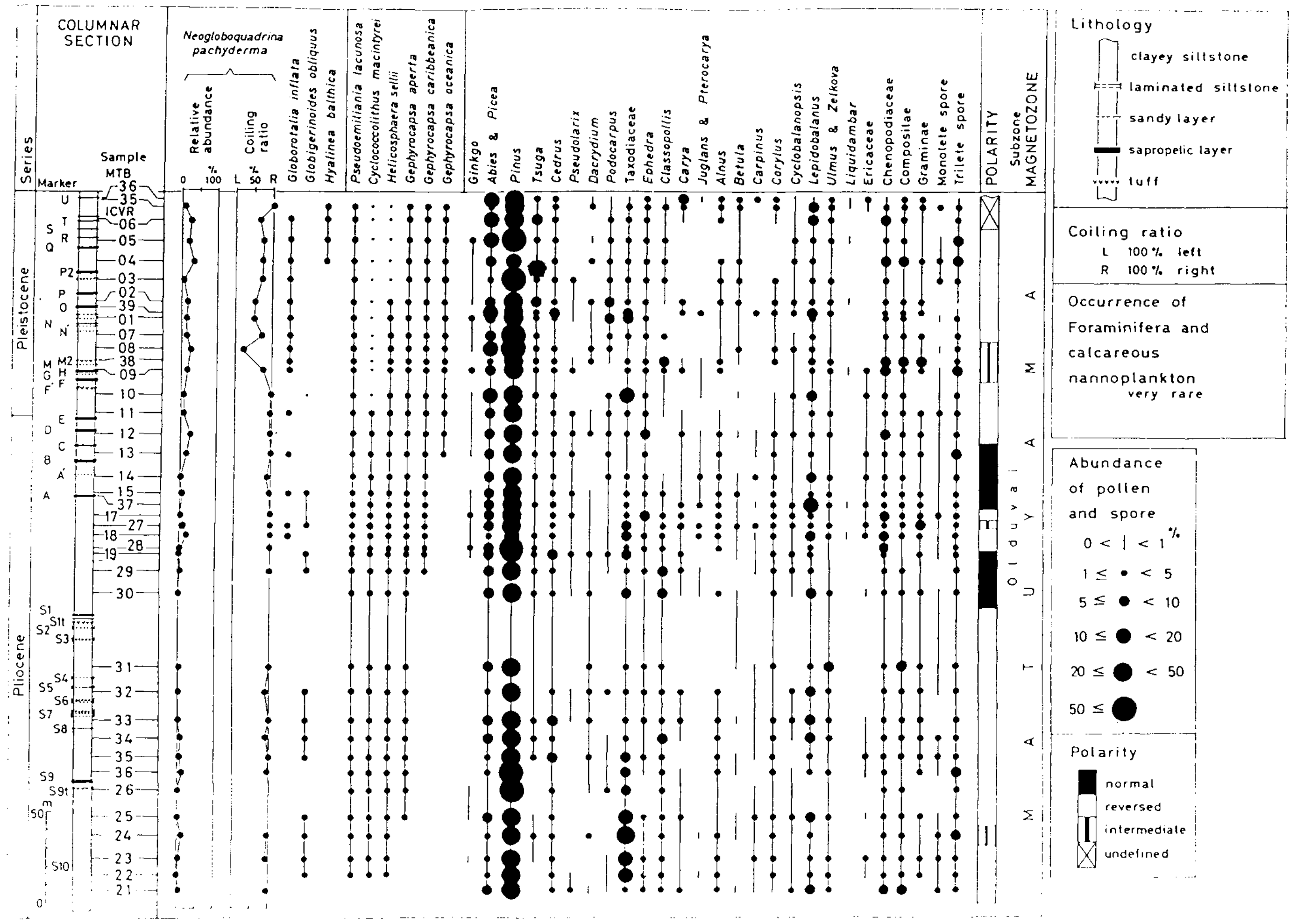


Figure 3.5. Magnetostratigraphy and biostratigraphy of the Vrica section.

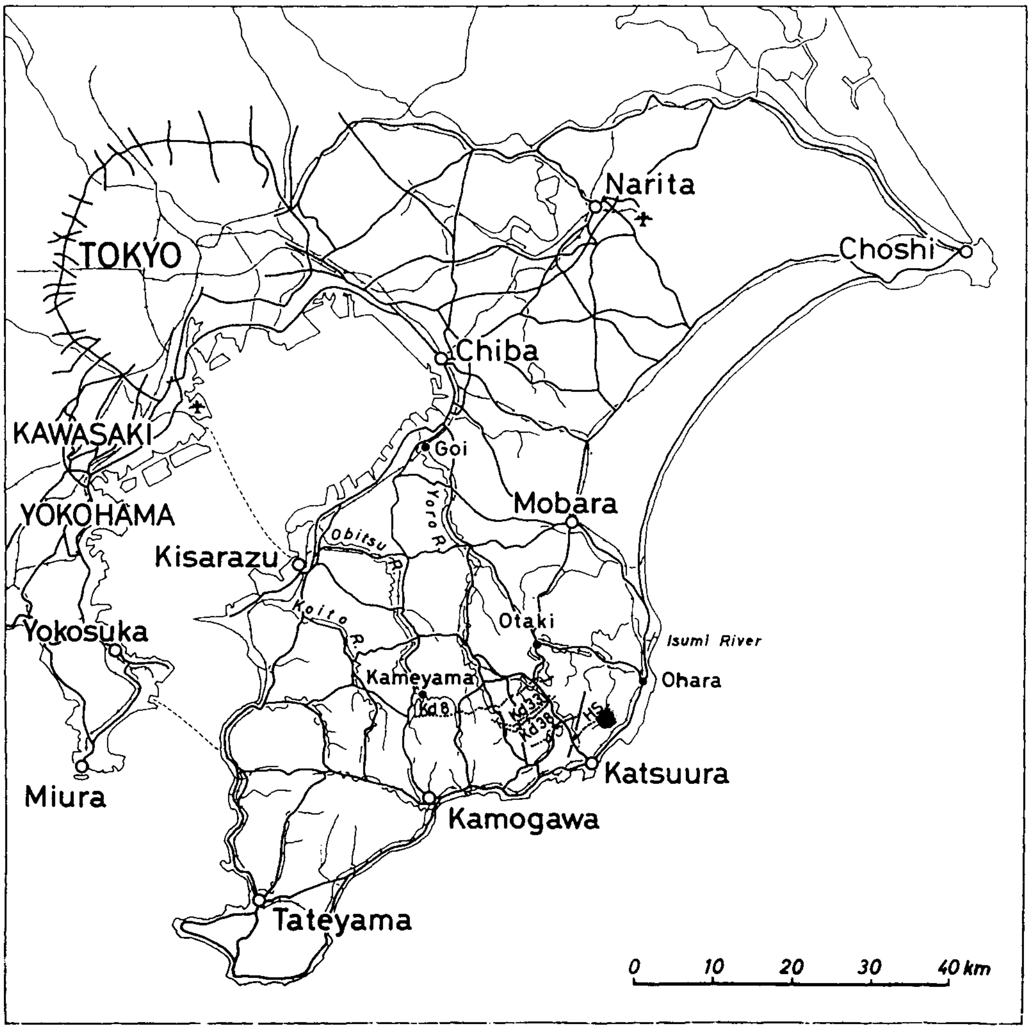


Figure 3.6. Index map of the Boso Peninsula.

In the geomagnetic polarity sequence of the Boso Peninsula, reversed polarity is dominant in the Plio–Pleistocene section, but there are three zones, plus several thinner horizons, of normal polarity (Niitsuma, 1976; Nakagawa and Niitsuma, 1977). In these same strata, planktonic microfossil index horizons remarkable for their use in long-distance correlation include (1) the lowest horizon of *Globorotalia tosaensis*, (2) the lowest horizon of *Globorotalia truncatulinoides*, (3) the highest horizon of *Globigerinoides obliquus*, (4) the highest horizon of *Discoaster brouweri*, (5) the lowest horizon of *Gephyrocapsa caribbeanica*, (6) the lowest horizon of *Eucyrtidium matuyamai*, (7) the horizon of S-D (sinistral-to-dextral) coiling change of *Pulleniatina* spp., (8) the lowest horizon of *Gephyrocapsa oceanica*, (9) the highest horizon of *Helicosphaera sellii*, and (10) the highest horizon of *Eucyrtidium matuyamai*.

Benthic foraminiferan and molluscan faunas indicate changes in sea-water temperatures. The indicated changes were moderate in the transition from Pliocene to Pleistocene, although the general trend was toward gradual cooling in the upper part of the section.

The transitional zone from Pliocene to Pleistocene is exposed along the upper streams of the Obitsu, Yoro, and Isumi rivers in the southeastern part of the peninsula (Figure 3.6). The area is near the Pacific coast, and modern geological maps have been published (Mitsunashi et al., 1961, 1976b), as has an excursion guidebook in English (Mitsunashi, Nakagawa, and Suzuki, 1976a).

Correlation between the Vrica area and the Boso Peninsula

The Vrica and Boso sections have yielded only a few planktonic microfossil species in common, but the sequence of index horizons noted earlier has been correlated to magnetostratigraphic and biostratigraphic sequences in many deep-sea sediment cores. Referring to the evidence in deep-sea sections, we can correlate the microfossil and geomagnetic horizons between Calabria and the Boso Peninsula as indicated in Figure 3.8. In this interpretation the normal-polarity zone in the lower part of the Kiwada Formation of the Boso Peninsula is correlated with the Olduvai

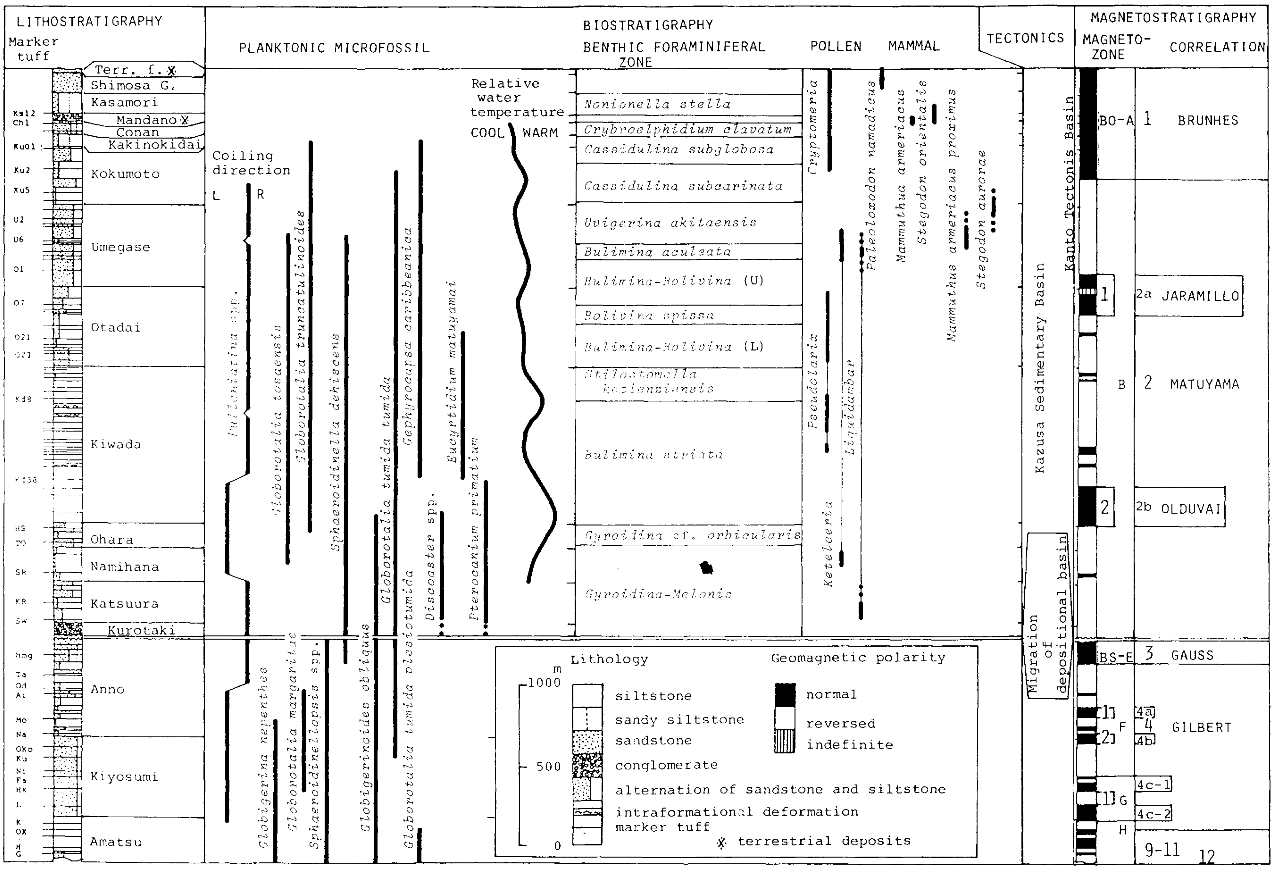
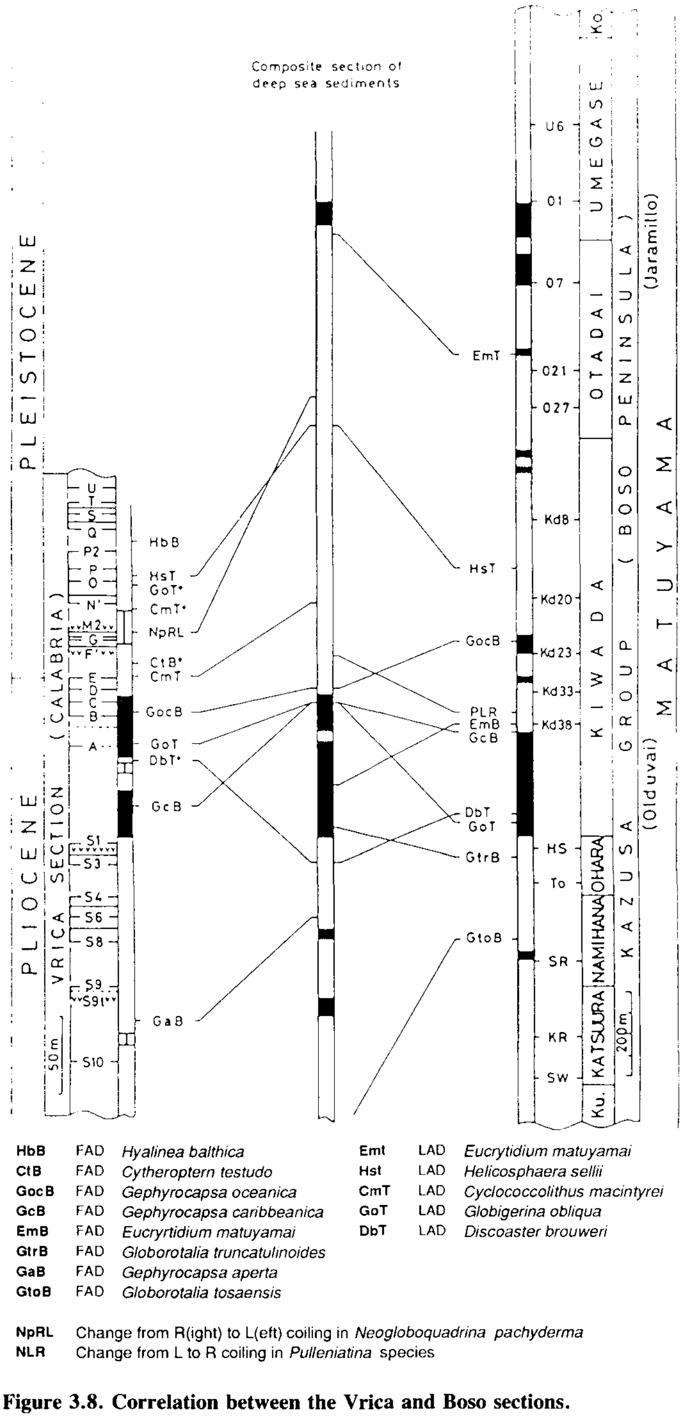


Figure 3.7. Magnetostratigraphy and biostratigraphy of the Plio–Pleistocene of the Boso Peninsula.



HbB	FAD	*Hyalinea balthica*	**Emt**	LAD	*Eucrytidium matuyamai*
CtB	FAD	*Cytheroptern testudo*	**Hst**	LAD	*Helicosphaera sellii*
GocB	FAD	*Gephyrocapsa oceanica*	**CmT**	LAD	*Cyclococcolithus macintyrei*
GcB	FAD	*Gephyrocapsa caribbeanica*	**GoT**	LAD	*Globigerina obliqua*
EmB	FAD	*Eucryrtidium matuyamai*	**DbT**	LAD	*Discoaster brouweri*
GtrB	FAD	*Globorotalia truncatulinoides*			
GaB	FAD	*Gephyrocapsa aperta*			
GtoB	FAD	*Globorotalia tosaensis*			

NpRL Change from R(ight) to L(eft) coiling in *Neogloboquadrina pachyderma*
NLR Change from L to R coiling in *Pulleniatina* species

Figure 3.8. Correlation between the Vrica and Boso sections.

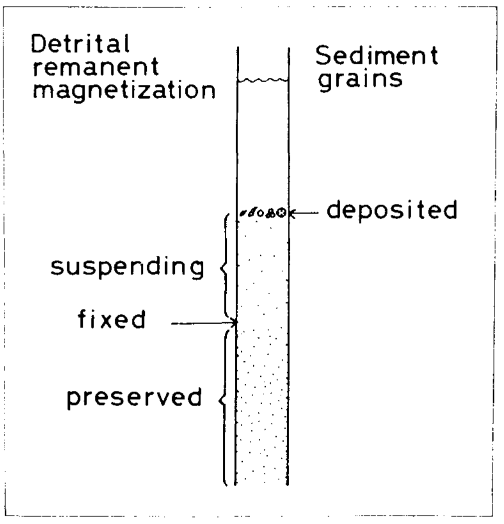


Figure 3.9. Illustration of time difference between sediment grains and detrital remanent magnetization.

normal subchron of the Matuyama reversed chron in the standard PMTS (paleomagnetic time scale) (Berggren et al., 1985) applied in deep-sea cores, and also with the normal-polarity zone identified to this paleomagnetic unit in the middle part of the Vrica section, just below the marker bed *e*.

The full sequence of microfossil index horizons linked with the paleomagnetic reversal sequence in deep-sea sediments is in rough agreement with the somewhat less complete records of the Boso and Vrica sections. The relationships of microfossil biochronologic events to geomagnetic reversals differ in detail, however, between the deep-sea synthesis and these two exposed sections. Undoubtedly, geographic, oceanographic, and ecological factors influenced local occurrences of fossils. Besides, if the age or equivalent horizon is measured using the geomagnetic reversal sequence as a standard, some time difference is a physical necessity. This is because the fixation of detrital geomagnetic orientation in marine deposits does not take place until the material is buried to about 40 cm (Niitsuma and Ku, 1977). The time required for this depth of burial is, of course, much longer in slowly deposited sediments. The age of a sediment grain, or a microfossil specimen, is therefore synchronous not with the age of the remanent magnetization in the horizon in which it occurs but with that of a horizon about 40 cm farther below it (Figure 3.9). Thus the apparent microfossil "date" of a paleomagnetic horizon is always somewhat younger in deep-sea sediments than in the more rapidly deposited shallow-marine equivalents.

With these factors taken into account, we can more accurately correlate the Plio–Pleistocene sections of Calabria to those of the Boso Peninsula. The Pliocene–Pleistocene boundary designated in the Vrica section, at the top of the sapropelite *e*, corresponds in age to a horizon between key tuffs Kd38 and Kd20 of the Kiwada Formation in the Boso Peninsula (Figure 3.8). The fission-track age of the Kd23 tuff in this interval is 1.6 ±0.2 Ma (Kasuya, 1990). Figure 3.10 shows further correlation of the Plio–Pleistocene sections in the Japanese islands and Taiwan. Nakagawa (1981) presented a correlation of magnetostratigraphic zones between Vrica and some other sections in Italy, according to the previously published biostratigraphy of Italian authors. Azzaroli et al. (Chapter 11, this volume) have reinterpreted the paleomagnetic findings according to new micropaleontological evidence.

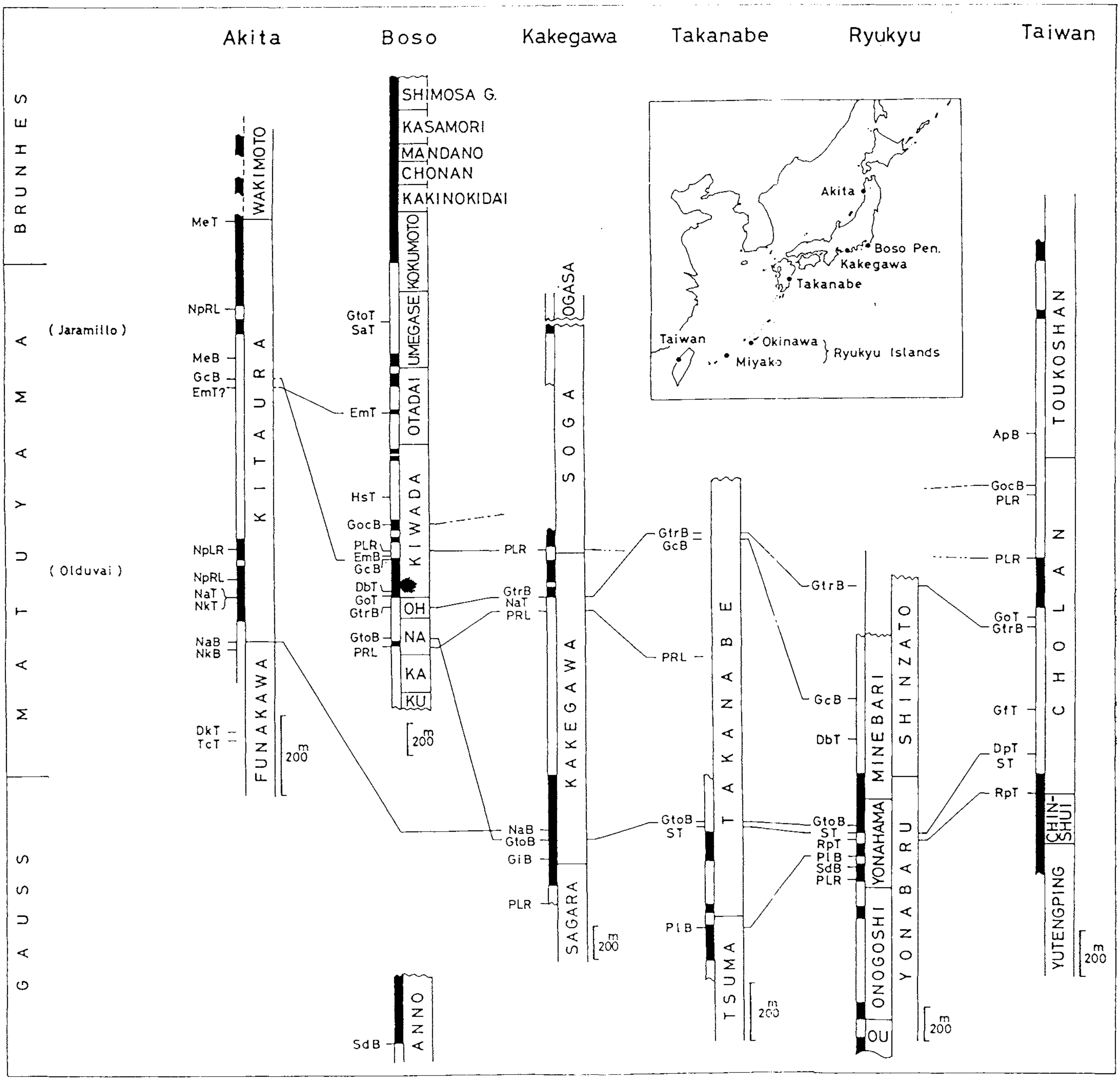


Figure 3.10. Correlation of the Plio–Pleistocene sections in the Japanese islands and Taiwan. (Courtesy of Dr. T.-Y. Huang)

ApB	FAD	*Asterorotalia pulchella*	**CmT**	LAD	*Cyclococcolithus macintyrei*
EmB	FAD	*Eucyrtidium matuyamai*	**DbT**	LAD	*Discoaster brouweri*
GaB	FAD	*Gephyrocapsa aperta*	**DkT**	LAD	*Denticulopsis kamtschatica*
GcB	FAD	*Gephyrocapsa caribbeanica*	**DpT**	LAD	*Discoaster pentaradiatus*
GlB	FAD	*Globorotalia inflata*	**EmT**	LAD	*Eucyrtidium matuyamai*
GocB	FAD	*Gephyrocapsa oceanica*	**GfT**	LAD	*Globigerinoides fistulosus*
GtoB	FAD	*Globorotalia tosaensis*	**GoT**	LAD	*Globigerinoides obliquus*
GtrB	FAD	*Globorotalia truncatulinoides*	**GtoT**	LAD	*Globorotalia tosaensis*
MeB	FAD	*Mesocena elliptica*	**HsT**	LAD	*Helicosphaera sellii*
NaB	FAD	*Neogloboquadrina asanoi*	**MeT**	LAD	*Mesocena elliptica*
NkB	FAD	*Neogloboquadrina kagaensis*	**NaT**	LAD	*Neogloboquadrina asanoi*
PlB	FAD	*Pseudoemiliana lacunosa*	**NkT**	LAD	*Neogloboquadrina kagaensis*
SdB	FAD	*Sphaeroidinella dehiscens*	**RpT**	LAD	*Reticulofenestra pseudoumbilicata*
			SdT	LAD	*Sphaeroidinella dehiscens*
			ST	LAD	*Sphaeroidinellopsis spp.*
			TcT	LAD	*Thalassiosira convexa*

NpRL	Horizon of R(ight) to L(eft) coiling change in *Neogloboquadrina pachyderma*
PLR	Horizon of L to R coiling change in *Pulleniatina* sp.
PRL	Horizon of R to L coiling change in *Pulleniatina* sp.

KA	Katsuura	**KU**	Kurotaki	**NA**	Namihana
OH	Ohara	**OU**	Oura		

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4 Comparison of the laminated units at Vrica and deep-sea sapropels from the eastern Mediterranean

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Introduction and background

Sapropels – highly organic, oxygen-reduced layers – in deep-water strata are similar in origin to organic-rich laminite units in upper-slope sediments and are clear manifestations of major changes in the climatic-oceanographic regime of the Mediterranean region. The periodic deposition of organic-rich facies at Vrica may have been related to the intensification of climatic oscillations in the late Pliocene, as seen in the deep-sea record. In this chapter, we review the conditions of sapropelic deposition and the relationship between the upper-slope and deep-sea events in the Ionian Basin during the later Pliocene and early Pleistocene.

Laminated sedimentary units are common features in marine sequences from Miocene to Pleistocene age in the marginal circum-Mediterranean region. For example, laminated sediments have been described from Algeria (Anderson, 1933), Spain (Geel, 1978), Sicily (Ogniben, 1957; Brolsma, 1978; Meulenkamp et al., 1978; McKenzie, Jenkyns, and Bennet, 1979; Gersonde, 1980), southern Italy (Martina et al., 1979), northern Italy (Sturani and Sampo, 1973), Morocco (Bizon, Muller, and Vergnaud-Grazzini, 1979), Cyprus (Bizon et al., 1979), and the Ionian Islands of Greece (Heimann, Just, and Muller, 1979; Thomas, 1980; Spaak, 1983). Many, but not all, of these laminated layers are rich in biogenic silica because of high abundances of diatoms. These diatomaceous laminites are often collectively referred to as "tripoli."

A number of different mechanisms have been proposed to explain the formation of Neogene laminites in marginal settings. The diatom-rich laminites of Morocco (Bizon et al., 1979) and Sicily (McKenzie et al., 1979; Gersonde, 1980; Van der Zwaan, 1982) have been attributed to increased productivity brought on by increased upwelling of nutrient-rich waters. Likewise, Van der Zwaan (1979) also concluded that increased productivity was responsible for the late Miocene diatomites in the Falconara section of Sicily. However, he attributed the increased productivity to an increase in the runoff of nutrient-rich continental water. In addition, it has generally been assumed that these laminated units were deposited in oxygen-deficient or anoxic bottom waters (Van der Zwaan, 1982, 1983; Spaak, 1983).

Whereas Neogene laminites have been found in marginal basins in both the eastern Mediterranean and the western Mediterranean, laminated deep-water sediments of equivalent age are found only in the eastern basin of the deep Mediterranean (Kidd, Cita, and Ryan, 1978; Cita and Grignani, 1982; Thunell, Williams, and Belyea, 1984). These "deep-sea laminites" are black, organic-rich units, commonly referred to as "sapropels." By definition, sapropels contain more than 2% organic carbon by weight (Olausson, 1961; Kidd et al., 1978). Lithologically similar units containing only 0.5% to 2.0% organic carbon are called "sapropelic layers."

Sapropel formation in the deep eastern Mediterranean is most commonly ascribed to the periodic development of a low-salinity surface layer that produced a density stratification and caused bottom waters to become oxygen-deficient (Olausson, 1961; Ryan, 1972; Thunell, Williams, and Kennett, 1977; Stanley and Blanpied, 1980; Rossignol-Strick et al., 1982; Thunell, Williams, and Cita, 1983). Calvert (1983) and Sutherland, Calvert, and Morris (1984) have argued convincingly that density stratification alone cannot account for the high organic-carbon content of deep-sea sapropels found in the eastern Mediterranean. Those authors suggest that the fresh-water runoff that formed low-salinity surface layers was also very rich in nutrients, and that caused a significant increase in productivity coincident with, and contributing to, oxygen minima below the euphotic zone. Evidence for increased productivity has been found in carbon-isotope studies of late Pleistocene sapropels (Thunell and Williams, 1982).

The question arises as to what relationship, if any, exists between these deep-sea sapropels and the laminites from marginal basins of the Mediterranean. It is obvious that the real distributions of these two facies are different, since laminites are found in both the eastern basin and the western basin. Also, as Spaak (1983) has pointed out, laminite deposition in marginal areas occurred much more frequently than sapropel formation in the deep eastern basin. While there have been a number of geochemical studies of deep-sea sapropels (Sigl et al., 1978; Calvert, 1983; Sutherland et al., 1984), very little is known about the organic geochemistry of laminites, making it very difficult to compare the two. In fact, virtually no information is available on the organic-carbon content of Neogene laminites from around the Mediterranean.

The objective of the study reported here was to make direct comparisons between the laminated units found in the Vrica section and the deep-sea sapropels recovered from DSDP site 125 on the Mediterranean Ridge. We attempted to evaluate the stratigraphic relationship of these facies, as well as to make preliminary comparisons of these units from the viewpoint of their organic geochemistry. By doing so, we should be able to determine whether or not it is appropriate to refer to the laminites at Vrica as "sapropels" (Selli et al., 1977; Colalongo et al., 1981, 1982).

Laminated layers in the Vrica section

In the Vrica section, 14 finely laminated layers are intercalated in the silty and marly limestone succession (Figure 4.1). These laminated units, together with sandy beds and volcanic ash layers, have been used as marker horizons for reconstructing the three component sections that make up the entire sequence at Vrica (Pasini et al., 1975; Selli et al., 1977). Biostratigraphic and chronostratigraphic studies indicate that there are five Upper Pliocene laminites and nine Lower Pleistocene laminites (Figure 4.1).

The laminated units at Vrica are considered to have been deposited in an oxygen-deficient, stagnant environment and have been referred to as sapropels (Pasini and Colalongo, 1982). Unlike many of the other Neogene laminated units found throughout the marginal Mediterranean, the laminites at Vrica contain very few diatoms and radiolaria (Pasini and Colalongo, 1982), although they do contain dwarfed benthic foraminifera assemblages, as compared with the abundant and diverse fauna found in the surrounding claystone. In addition, more than 700 well-preserved specimens of fish have been identified from these layers (Landini and Menesini, 1977, 1978). The benthic foraminifera assemblages, and also the fossil fish, suggest a depositional depth of more than 500 m for this sequence. The undisturbed condition of the fish and the sedimentary laminae is indicative of an abiotic, and thus anoxic, bottom environment.

Biostratigraphic correlation of laminites and sapropels

DSDP site 125 is the only eastern Mediterranean deep-sea succession in which Lower Pleistocene sapropels have been recovered within a rather complete sedimentary sequence. Despite some anomalous sedimentation patterns above and below these Lower Pleistocene sapropels (Raffi and Sprovieri, 1984), the nannofossil data for site 125 and the Vrica section establish a good biostratigraphic correlation of the two sequences (Figure 4.2).

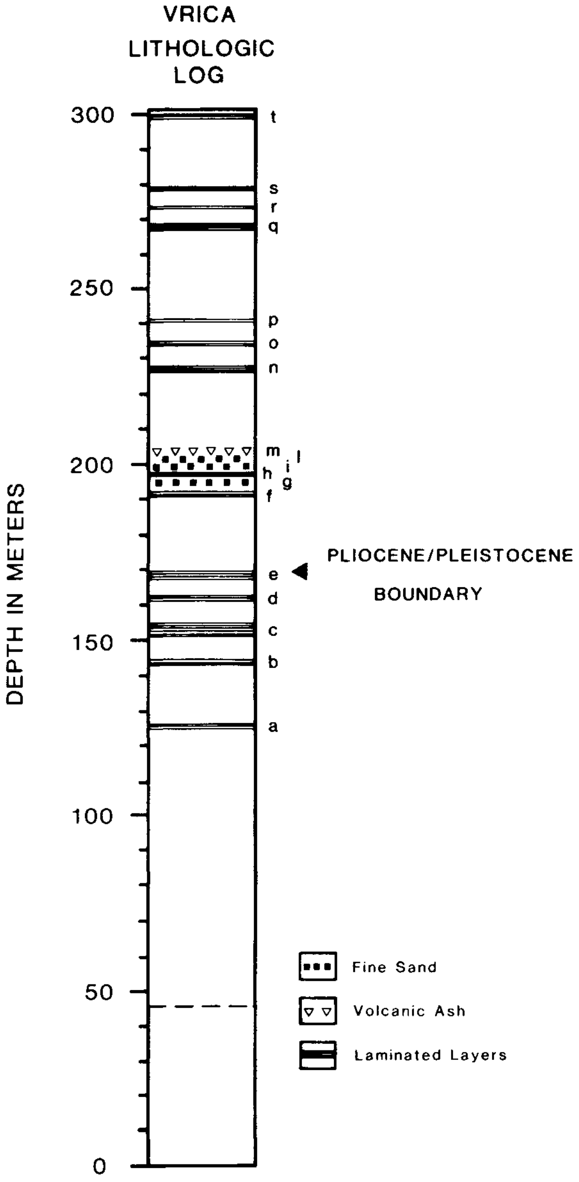

Figure 4.1. Stratigraphic column for the Vrica section, showing the positions of various lithologic marker beds and the position of the Pliocene–Pleistocene boundary (top of laminated layer *e*).

In the Vrica section (Rio et al., Chapter 5, this volume), laminated layers *a* through *f* were deposited after the last occurrence of *Discoaster brouweri* and before the first appearance of *Gephyrocapsa oceanica s.l.* (i.e., the *Crenalithus doronicoides* zone). In this part of the sequence, laminate *e* has been proposed as the physical basis for the Pliocene–Pleistocene boundary-stratotype at Vrica (Pasini and Colalongo, Chapter 2, this volume). Unfortunately, no sapropels are present in the equivalent stratigraphic interval at site 125 (Figure 4.2), where the *C. doronicoides* zone is represented by a very condensed section.

Laminite *h* at Vrica occurs within the Lower Pleistocene *Calcidiscus macintyrei* zone, and again there are no stratigraphically equivalent sapropels at site 125 (Figure 4.2). The remaining seven laminated units at Vrica (layers *n*–*t*) are dispersed throughout the upper 73 m of the section, within the

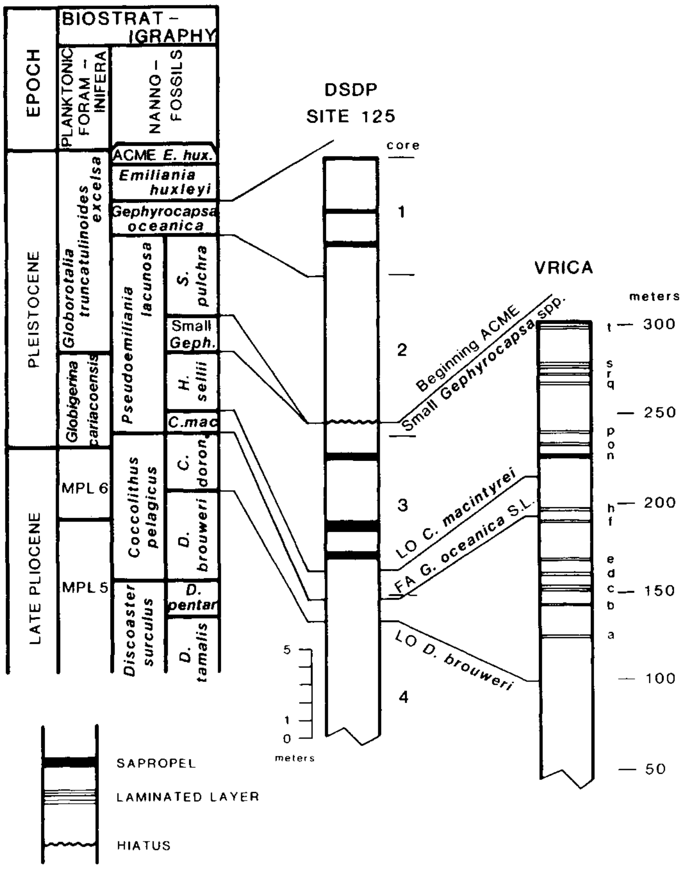

Figure 4.2. Biostratigraphic correlation of sapropel layers in DSDP site 125 and laminated layers in the Vrica section. The biostratigraphic data for site 125 are from Raffi and Sprovieri (1984), and those for Vrica are from Rio et al. (1984) and Rio et al. (Chapter 5, this volume).

Helicosphaera sellii zone (Figure 4.2). It is within this Lower Pleistocene interval that sapropels are also found at site 125.

Geochemical comparison of laminites and sapropels

One Upper Pliocene (layer *e*) and two Pleistocene laminites (layers *f* and *h*) were selected for geochemical study. A series of samples was collected across each of these layers, and the total organic-carbon and carbonate contents were measured. The findings from the determinations of organic-carbon and carbonate contents are presented in Figure 4.3. A consistent and systematic decrease in carbonate content and an increase in organic-carbon content were found within each of the laminites (Figure 4.3). Both the basic patterns and the absolute values were similar across the layers. In sediments enclosing the laminites, carbonate values ranged from 16% to 20%, while within the laminites those values decreased to between 12% and 14%. Within 50 cm above the top of each laminite, the carbonate content returned to the value typically found just below the layer.

The organic-carbon content, both above and below each of the laminites, generally was in the range of 0.4–0.5% (Figure 4.3). In contrast, the organic-carbon content was nearly doubled within the laminite layers, reaching values of 0.8–1.0%. In accordance with the current definition, the three studied laminites cannot be considered true sapropels, because they do not contain more than 2% organic carbon by weight. However, if organic-carbon content is the only criterion considered, then these three laminites can be classed as sapropelic layers.

The relationship between the organic carbon in the Vrica sediments and the C/N (carbon/nitrogen) ratios is illustrated in Figure 4.4. These data clearly indicate that the C/N ratios increase with increasing organic-carbon content (i.e., that the laminated sediments typically have higher C/N ratios). This is the same basic trend observed by Calvert (1983) in a study of late Pleistocene deep-sea sapropels.

The observed trends in C/N ratios may be explainable in a number of ways. Most simply, they may be reflecting two different sources of organic matter, since high C/N ratios are considered to be characteristic of terrestrial, rather than marine, organic matter (Trask, 1953; Bordovskiy, 1965). Using that criterion, Sigl et al. (1978) concluded that there was a large terrestrial component in the organic matter of Pliocene and Pleistocene sapropels from the eastern Mediterranean. However, as pointed out by Calvert (1983), C/N ratios can also be artifacts of diagenesis. As organic matter decomposes with burial, nitrogen is preferentially lost relative to carbon, and this process is accentuated in environments characterized by a high accumulation rate (Muller, 1977). That may have been the situation at Vrica, where sediments accumulated at rates of up to 44 cm/1,000 years (Rio et al., Chapter 5, this volume). Thus, without determining the extent of alteration of the organic matter, it is not possible to conclude unequivocally whether the organic matter in the laminites is primarily from a marine or terrestrial source.

Previous geochemical and sedimentological studies of deep-sea sapropels (Kidd et al., 1978; Muller, 1978; Sigl et al., 1978; Cita and Grignani, 1982; Calvert, 1983) provided the basis for comparison with the laminites we examined from Vrica. The biggest difference was in the organic-carbon content of the laminites, which fell within the range for sapropelic sediments (0.5–2.0% organic C), as opposed to the much higher values in the sapropels. Despite that difference, the "sapropelic laminites" and the deep-sea sapropels were similar in that both were characterized by high C/N ratios (Sigl et al., 1978; Calvert, 1983).

Another similarity between the laminites and the sapropels was that the carbonate contents in both units were significantly lower than in the surrounding sediment (Calvert, 1983). Such a decrease in carbonate content could be due to (1) a decrease in the production of calcareous plankton, (2) an increase in carbonate dissolution, and/or (3) an increase in the input of noncalcareous sediments. With regard to the first explanation, a decrease in productivity during the deposition of either laminites or sapropels would be at odds with the models that have been proposed for their formation. In particular, stable-isotope

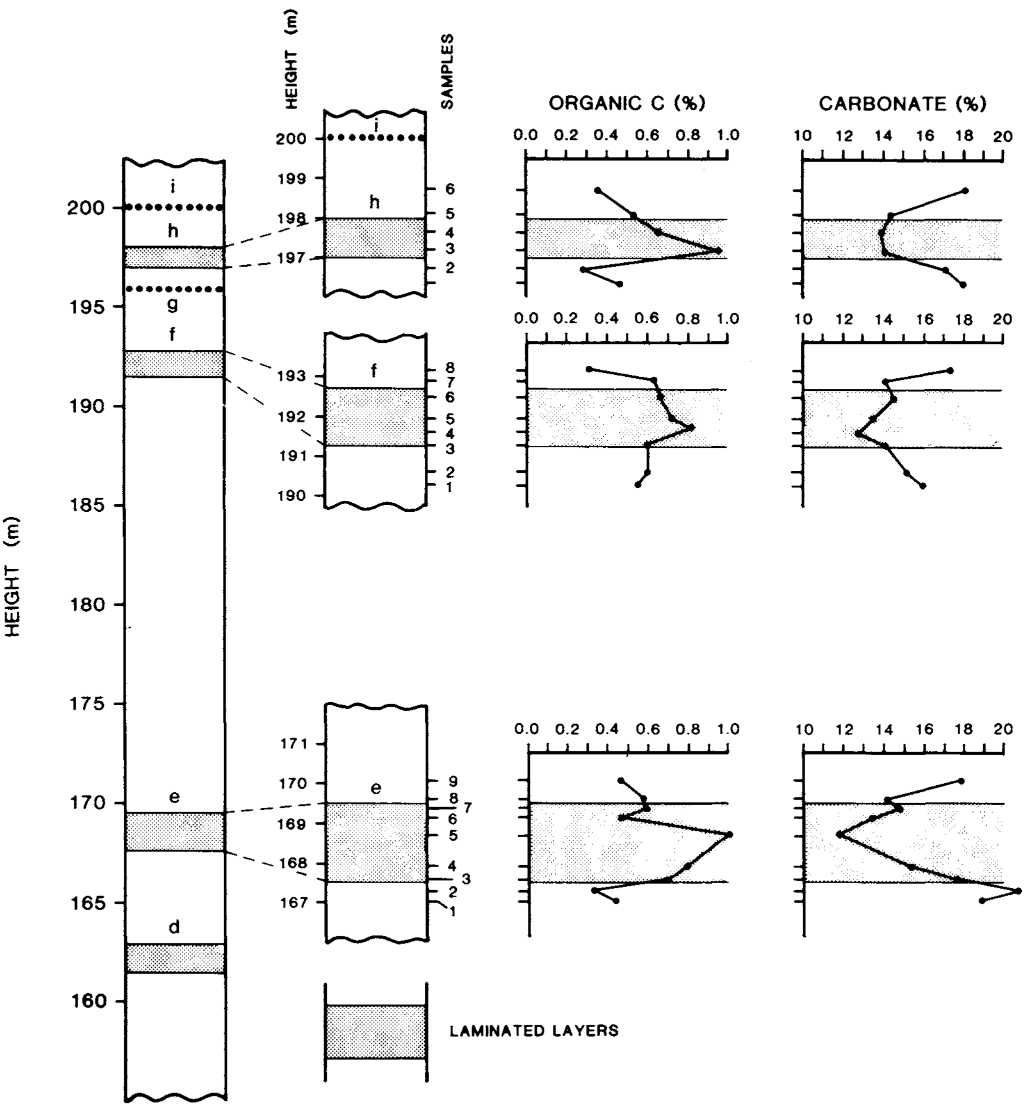

Figure 4.3. Vrica laminates. Results of analyses of organic-carbon and carbonate contents across laminated layers *e*, *f*, and *h* at Vrica.

studies (Thunell and Williams, 1982) and geochemical studies (Calvert, 1983) indicate a general increase in productivity during the deposition of late Pleistocene sapropels. The second explanation, that of dissolution, is inconsistent with the observation that calcareous microfossils in both types of facies are not appreciably less well preserved within these layers than in surrounding sediments (Sigl et al., 1978; Thunell and Williams, 1982). We consider, therefore, that the decrease in carbonate content within laminites and sapropels may have been due to dilution by noncalcareous sediments. Increased clastic input associated with enhanced runoff from the continents could be responsible for the observed decrease in carbonate, particularly in the marginal settings where most sapropelites (= laminites) are found. Increased runoff also would have been a source for the additional nutrients needed to raise productivity and increase the flux of organic carbon to the seafloor. An increase in runoff also may have resulted in the input of higher levels of terrestrial organic matter, which could be at least a minor component of the relatively high total organic carbon in the laminated layers.

Discussion

The deposition of laminites in marginal settings and of sapropels in the deep eastern basins undoubtedly was related to major changes in the climatic-oceanographic regime of the Mediterranean region. Thunell et al. (1984) have demonstrated that the latest Pliocene and early Pleistocene sapropels were fundamentally different from earlier Miocene and Pliocene sapropels and that the formation of the more recent sapropels was climatically modulated. In particular, the alternating glacial–interglacial climatic conditions that began in the late Pliocene (Thunell and Williams, 1983; Shackleton et al., 1984, 1995) may have been the triggering mechanism for both laminate and sapropel formation. These glacial–interglacial climatic changes most certainly had an

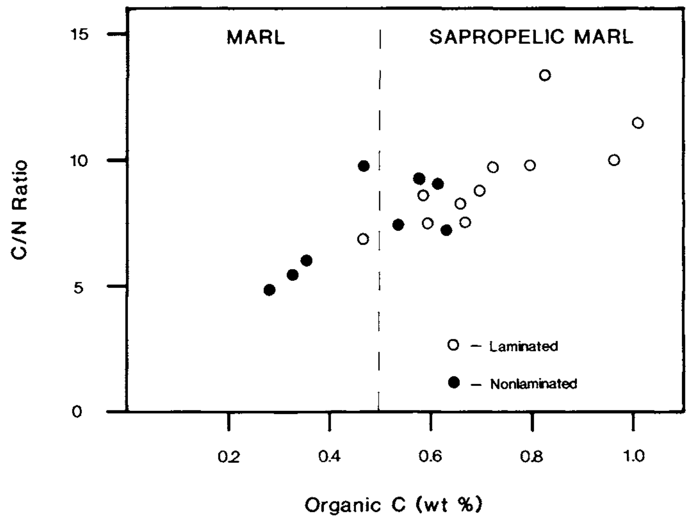

Figure 4.4. Relationship between organic-carbon content and C/N ratio in laminated and nonlaminated sediments from Vrica. The distinction between marls and sapropelic marls is based on the definition that sapropelic marls contain between 0.5% and 2.0% organic carbon (Kidd et al., 1978).

impact on the discharge of fresh water into the Mediterranean. It has been suggested that increased discharge from the Nile, due to increased precipitation, periodically altered the surface-water conditions in the eastern Mediterranean (Rossignol-Strick et al., 1982). Increased runoff would serve to establish sapropel-producing conditions in two ways: (1) The surface-water productivity would be increased because of increased nutrient input from the continents, and that, in turn, would increase the flux of organic matter to the seafloor. (2) The reduction of surface-water salinities would inhibit vertical overturn of the water column, resulting in oxygen depletion in the bottom waters and an environment favorable for preservation of organic matter.

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5 Calcareous nannofossil biochronology and the Pliocene–Pleistocene boundary

DOMENICO RIO, ISABELLA RAFFI, and JAN BACKMAN

Introduction

Calcareous nannofossils are microscopic skeletal elements of calcite produced by planktic unicellular algae, occurring abundantly in marine sediments from the Jurassic to the present. The rapid evolution of the nannofossils through time has been used to establish highly resolved biostratigraphic zonations, particularly for the Cretaceous and Cenozoic eras (e.g., Martini, 1971; Thierstein, 1976; Roth, 1978; Bukry, 1978). The zonal boundaries are defined by extinctions or first appearances of individual species, and such datum events have been found to have occurred synchronously on a global scale, when calibrated against oxygen-isotope stratigraphy (e.g., Thierstein et al., 1977) or against magnetostratigraphy (e.g., Backman and Shackleton, 1983).

Nannofossils are well represented in the Pliocene and Lower Pleistocene marine deposits in Italy, and therefore they provide a means by which this regional geologic record can be correlated with extra-Mediterranean areas (Figure 5.1). This possibility of correlation is important because the type regions of the Pliocene and the Pleistocene are located in Italy (Lyell, 1833; Berggren, 1972).

In this chapter we shall briefly review the development of nannofossil biostratigraphy at the Pliocene–Pleistocene transition, as derived from deep-sea sediments and regions in Italy. Second, we shall present the biochronology of critical nannofossil datum events. Third, we shall insert the regional Italian lithostratigraphic record into the established nannofossil biostratigraphic-biochronologic framework and estimate the age of the physical horizon that has been proposed by INQUA Subcommission 1-d, "Pliocene/Pleistocene Boundary," to be used as the definition of the Pliocene–Pleistocene boundary in the Vrica section, Calabria.

This chapter was submitted essentially in its present form in 1985, when two of us (D.R. and I.R.) were at the Istituto di Geologia, Università di Parma. Although much work has subsequently been done on Italian Plio–Pleistocene stratigraphy (Nikiforova, Foreword, this volume), only a few changes have been necessary to update the biostratigraphic data herein.

Nannofossil biostratigraphy at the Pliocene–Pleistocene transition

Biostratigraphic studies of calcareous nannofossils originated from the work of Bramlette and Riedel (1954). Within a decade, the development of Pliocene and Pleistocene nannofossil biostratigraphy was widely recognized. Ericson, Ewing, and Wollin (1963) and Riedel, Bramlette, and Parker (1963) characterized the extinction of *Discoaster brouweri* as a useful datum event in the deep-sea sediments of different oceans. Akers (1965) observed that the evidence of that extinction event in the northern Gulf of Mexico had been preceded by the closely spaced extinctions of *Discoaster surculus*, *D. pentaradiatus*, and *D. variabilis*. Most notably, however, Ericson et al. (1963) correlated the extinction level of *Discoaster brouweri* to the Pliocene–Pleistocene boundary. Although that correlation was immediately questioned (Riedel et al., 1963; Akers, 1965), it has been widely used since then to separate the Pliocene from the Pleistocene in deep-sea sediments. A study by McIntyre, Bé, and Pretiskas (1967) indicated that the first appearances of species of the genus *Gephyrocapsa* followed shortly after the extinction of *D. brouweri* and that those gephyrocapsid events could also be used for approximate determination of the Pliocene–Pleistocene boundary.

The first tentative nannofossil zonations for the Upper Pliocene and the Lower Pleistocene were proposed by Hay et al. (1967) and Boudreaux and Hay (1967, 1969). Gartner (1969) proposed a zonation that later was incorporated in the Cenozoic "standard zonation" of Martini (1971). In a comprehensive zonal scheme for the entire Cenozoic, Bukry (1973) utilized the *Gephyrocapsa* group to refine biostratigraphic subdivisions in the interval of the Pliocene–Pleistocene boundary, while also noting that *Calcidiscus macintyrei* disappeared shortly after the *D. brouweri* extinction. The revised Pleistocene zonation of Gartner (1977) disregarded the *Gephyrocapsa* events and instead subdivided the Lower Pleistocene interval, after the extinction of *D. brouweri*, in the ascending order of (1) the extinction of *C. macintyrei*, (2) the extinction of *Helicosphaera sellii*, and (3) the upper limit of a short stratigraphic interval strongly dominated by small *Gephyrocapsa* spp.

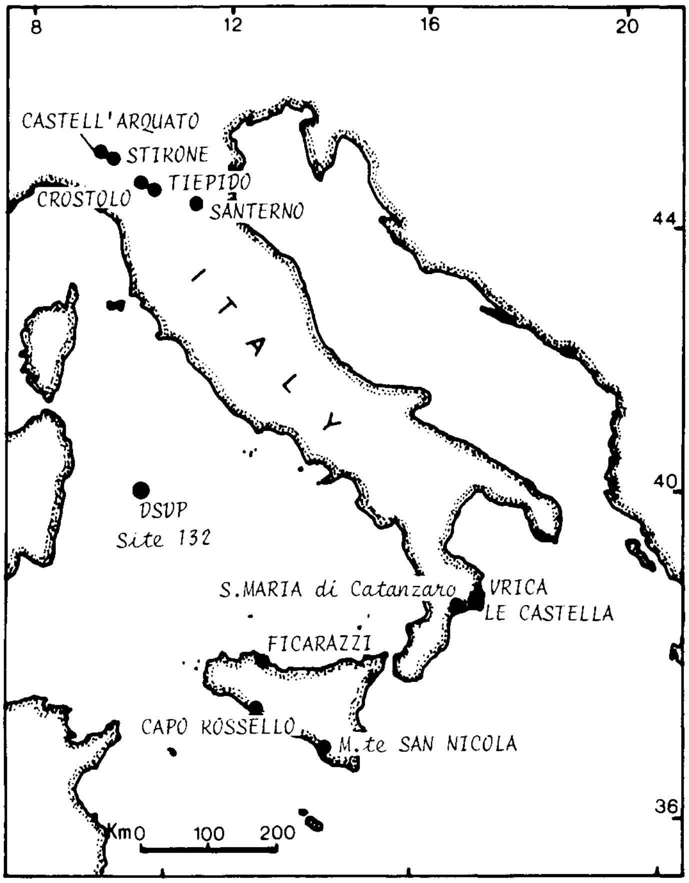


Figure 5.1. Italian sections discussed in the text.

Gartner's (1977) set of events (including the *D. brouweri* extinction) can be readily identified in Mediterranean deep-sea sediments (Raffi and Rio, 1979). In Italian land sections, however, consistent identification of these datum events has been difficult because of severe sediment reworking and, particularly in northern Italian sections, because of the fact that both *C. macintyrei* and *D. brouweri* declined to low abundances prior to their extinctions (Rio et al., 1982). Quantitative approaches can be used, however, to overcome most of the reworking problem (Backman and Shackleton, 1983). On the other hand, the use of first appearances, rather than extinctions, minimizes the influence of reworking on the biostratigraphic resolution. This led Haq, Berggren, and Van Couvering (1977), Gartner (1977), and Raffi and Rio (1979) to make renewed studies of the biostratigraphic distribution of the *Gephyrocapsa* group of species, which, although beset by problems of taxonomy and nomenclature, are abundant in Italian land sections. This group also underwent marked evolutionary changes in the critical time interval, thus offering the opportunity for high-level biostratigraphic control. Rio (1982), in summarizing the work on Lower Pleistocene gephyrocapsids, indicated four biostratigraphically useful events, in ascending order: (1) the first appearance of *Gephyrocapsa oceanica s.l.*, (2) the first appearance of *Gephyrocapsa* spp. larger than 5.5 μm, (3) the beginning of dominance of small *Gephyrocapsa* spp., and (4) the end of dominance of small *Gephyrocapsa* spp.

Finally, Backman and Shackleton (1983) demonstrated that the increase in abundance of *Discoaster brouweri triradiatus* relative to *D. brouweri* represents a distinct biostratigraphic signal in sediments of late Pliocene age. They also showed that these *Discoaster* forms had virtually the same extinction times. *Discoaster brouweri triradiatus* shows rare and scattered occurrences in the Pliocene, except during the last few hundred thousand years of its range, and therefore is less likely to have been reworked.

The zonations and the set of events discussed are summarized in Figure 5.2.

Nannofossil biochronology

The biochronologic resolution that can be achieved by means of calcareous nannofossils at the transition from the Pliocene to the Pleistocene is determined by eight of nine datum events reported in Figure 5.2. All but *H. sellii* provide datum events usable over wide geographic areas, including the Mediterranean region. Backman and Shackleton (1983) have shown that *H. sellii* is virtually valueless for long-distance correlations, and this species is therefore excluded from further biochronologic discussion.

The remaining eight datum events have been determined in sedimentary sequences that have established magnetostratigraphies, allowing their ages to be derived from interpolations of sediment accumulation rates in reference to the magnetic-reversal boundaries. Therefore, the precision of each age estimate depends partly on the accuracy of the time scale used and partly on uncertainties in estimates of sediment accumulation rates, which increase as a function of increasing distance from the nearest magnetic-reversal boundary. The error in our age estimates probably does not exceed ±0.05 Ma in seven of the eight cases.

In a series of studies, Backman and Shackleton (1983), Backman, Shackleton, and Tauxe (1983), and Backman et al. (1984) worked out a reliable nannofossil biochronology in which a sequence of datum events can be shown to follow a predictable pattern in numerous deep-sea cores for which there are paleomagnetic controls, and which represent a wide range of geographic locations.

The first datum event is defined as the increase in abundance of *D. brouweri triradiatus* relative to *D. brouweri* to values in excess of 20%. The higher proportion of the former lasted uninterruptedly until the simultaneous extinctions of both forms. The duration of this signal may range between approximately 0.1 and 0.2 m.y., since the available data do not allow a better age determination of its beginning. The synchronous extinctions of *D. brouweri* and *D. brouweri triradiatus* occurred immediately prior to the Olduvai subchron. These events took place at 1.89 Ma, according to the time scale of Berggren, Kent, and Van Couvering (1985) used in preparing this chapter, corrected to 2.04 Ma in the most recently revised orbitally tuned time scale (Shackleton et al., 1995). The available data indicate that these *Discoaster* events

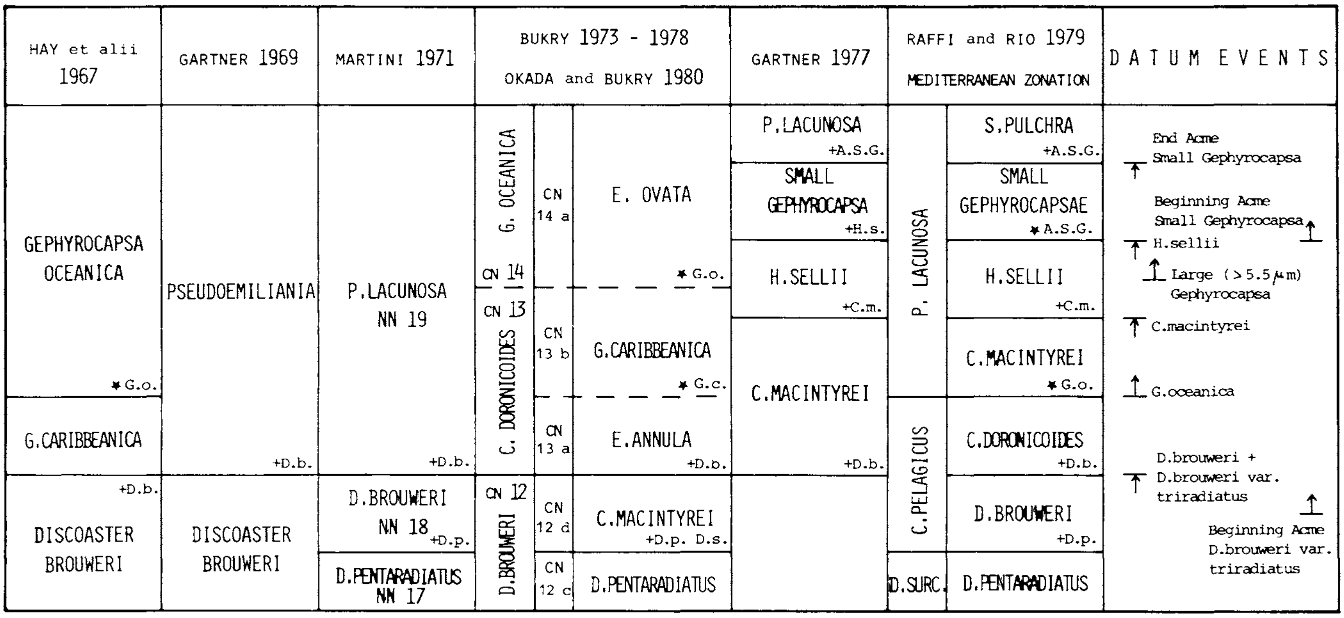


Figure 5.2. Principal nannofossil zonation schemes proposed for the Upper Pliocene and Lower Pleistocene. Note that the range of *Helicosphaera sellii* is valid only for Mediterranean sections. Abbreviations in Figure 5.2 refer to binomens of index taxa in right-hand column. Other symbols are as follows: A.S.G., acme of small gephyrocapsids; *, first appearance; +, last appearance or extinction.

occurred at the same time in all low-latitude and mid-latitude oceans, and in the higher latitudes of the North Atlantic as well. Such was also the case with the extinction of *Calcidiscus macintyrei*, although that event occurred later than the Olduvai subchron. These three datum events are easily identified from a taxonomic point of view, adding to their value as biochronologic indices. We have therefore used the extinction events of these forms for deducing sediment accumulation rates in the sequence provided by DSDP site 132 in the Tyrrhenian Sea, in which magnetostratigraphy is lacking, in order to derive age estimates for the *Gephyrocapsa* group in that sequence. The *C. macintyrei* event has also been used in analyzing core V12–18 and DSDP site 502B in order to account for the accumulation rate changes occurring between the Olduvai and Jaramillo subchrons.

The biochronology of datum events provided by the genus *Gephyrocapsa* is complicated by the fact that unambiguous identification of gephyrocapsid species is difficult because of profound intra-species and inter-species morphologic variations. Only by applying rigorous morphometric criteria (Rio, 1982) has it been possible to consistently distinguish certain transition events (as discussed earlier) in the group's evolution using light-microscope techniques. In order to estimate ages for the four *Gephyrocapsa* events in sequences with established magnetostratigraphies, we have studied their stratigraphic distributions in core V28–239 from the western equatorial Pacific (Backman and Shackleton, 1983), in core V12–18 from the southwestern Atlantic (Backman and Shackleton, 1983), and in DSDP site 502B in the Caribbean (Prell et al., 1982).

The findings from core V28–239 are presented in Figure 5.3. Of all the material studied, this core provides the best sequence as far as continuity in deposition and control of sedimentation rates are concerned. The first appearance of *G. oceanica s.l.* occurs over a 30-cm interval just above the Olduvai subchron, and the first appearance of *Gephyrocapsa* spp. larger than 5.5 μm occurs at a level about halfway between the Olduvai and Jaramillo (1.34 Ma uncorrected; 1.44 corrected). Shortly below the base of the Jaramillo subchron, a profound change is recorded in the *Gephyrocapsa* assemblage, characterized by disappearance of large and normal-sized *Gephyrocapsa* spp. This event marks the beginning of an interval in which, virtually, only *Gephyrocapsa* spp. smaller than 3.5 μm are present. This dominance of small *Gephyrocapsa* spp. comes to an end close to the top of the Jaramillo subchron, being just within it at low latitudes, and just above it at middle and higher latitudes in the Atlantic, as well as in the Mediterranean (Castradori, 1993). Gartner (1977) first observed this interval and introduced a "small *Gephyrocapsa* Zone," which he defined as "the interval from the highest occurrence of *Helicosphaera sellii* to the highest level of dominantly small *Gephyrocapsa*." Gartner considered that the uppper limit of the stratigraphic interval characterized by the acme of small *Gephyrocapsa* spp. was "very sharp," and its lower limit less sharp "but still readily identifiable." Rio's (1982) results indicate, however, that the lower limit of the small *Gephyrocapsa* spp. acme is no less distinct than its upper limit. The data sets obtained from core V12–18 and from DSDP site 502B (Figure 5.4) provide age estimates for the gephyrocapsid datum events that are, without exception, in good agreement with those obtained from V28–239, indicating synchronous occurrences in the western equatorial Pacific, the southwestern Atlantic, and the Caribbean Sea.

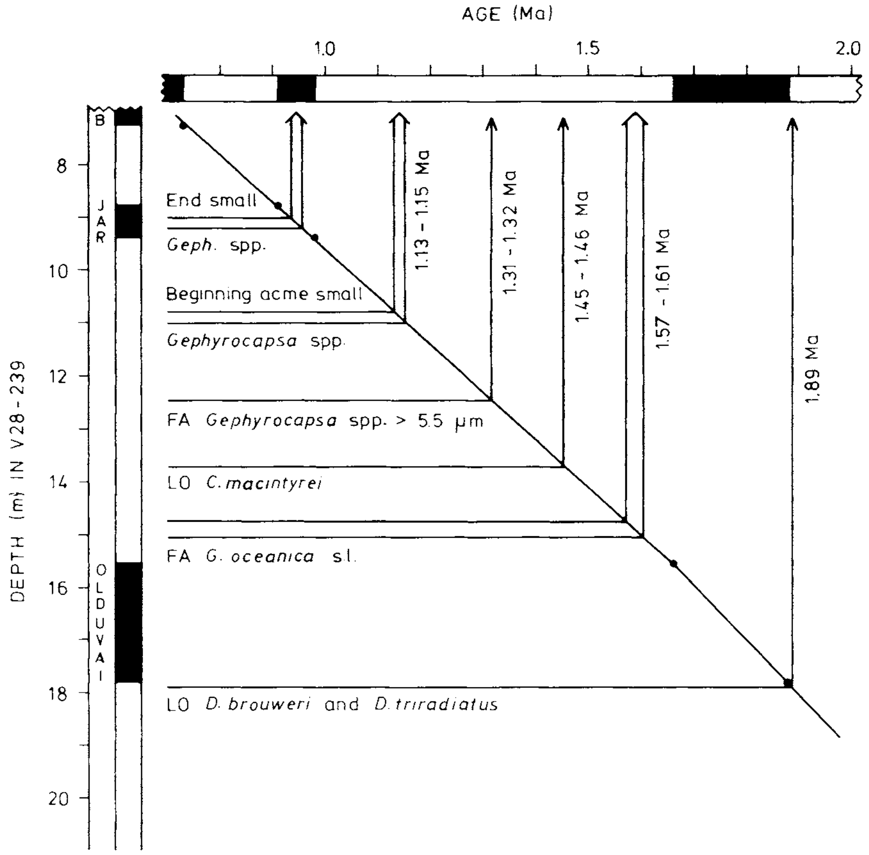


Figure 5.3. Sediment accumulation rate for core V28–239, plotted against nannofossil-datum chronology.

Age estimates for *Gephyrocapsa* events were derived also from DSDP site 132 by linear interpolation among the extinction events for *Pseudoemiliana lacunosa* (Thierstein et al., 1977; Raffi and Rio, 1979), *C. macintyrei*, and *D. brouweri/D. brouweri triradiatus* (depth positions used for the three latter events are those reported by this study). At DSDP site 132, the end of dominance of small *Gephyrocapsa* spp. could not be determined, because it occurs in an unrecovered interval (Raffi and Rio, 1979). Recently, it has been widely observed in piston cores from the Ionian Sea and Levantine Basin (Castradori, 1993). The ages obtained for the remaining *Gephyrocapsa* events are in agreement with those obtained from the extra-Mediterranean sequences (Table 5.1).

Nannofossil biostratigraphy of the Italian surface exposures

A correlation chart for classical Italian land sections (including stratotypes) and DSDP site 132 is presented in Figure 5.5. This figure shows the critical biostratigraphic and lithostratigraphic characters on which the chronostratigraphy of the Pliocene and the Pleistocene is based. Two topics that deserve attention are, first, the timing of the appearance of the "northern guests" in the Mediterranean and, second, the prior lithostratigraphic definitions of the Pliocene–Pleistocene boundary.

The "northern guests" and the definition of the Pliocene–Pleistocene boundary

The definition of the Pliocene–Pleistocene boundary (its assignment to a physical horizon) in the stratotype section had to be such that, besides being amenable to long-distance correlation, it would respect the historical concepts of the Pliocene and the Pleistocene in order to maintain stratigraphic stability. As discussed by Pelosio, Raffi, and Rio (1980), the Pliocene–Pleistocene boundary traditionally had been placed at a level at which marine faunal elements that at present are restricted to the boreal bioprovince began to occur as "northern guests" in the Mediterranean (specifically the Italian) geologic record. Indeed, that concept was the main reason for a concentration of studies of the Calabrian Stage (the traditionally accepted first stage of the Pleistocene) at Santa Maria di Catanzaro (Gignoux, 1913; Selli, 1971), of the top of the Piacenzian (the generally accepted stage for the Upper Pliocene) at Castell'Arquato (Barbieri, 1971; Colalongo, Elmi, and Sartoni, 1974), of the proposed Pliocene–Pleistocene boundary-stratotype at Le Castella (Pelosio et al., 1980), and of the subsequently proposed boundary-stratotype in the Vrica section (Colalongo et al., 1982; Pasini and Colalongo, Chapter 2, this volume). It is important to note that Lyell's (1833) original concept of the Pleistocene, based on fossil molluscan faunas in Italian strata with no less than 70% of

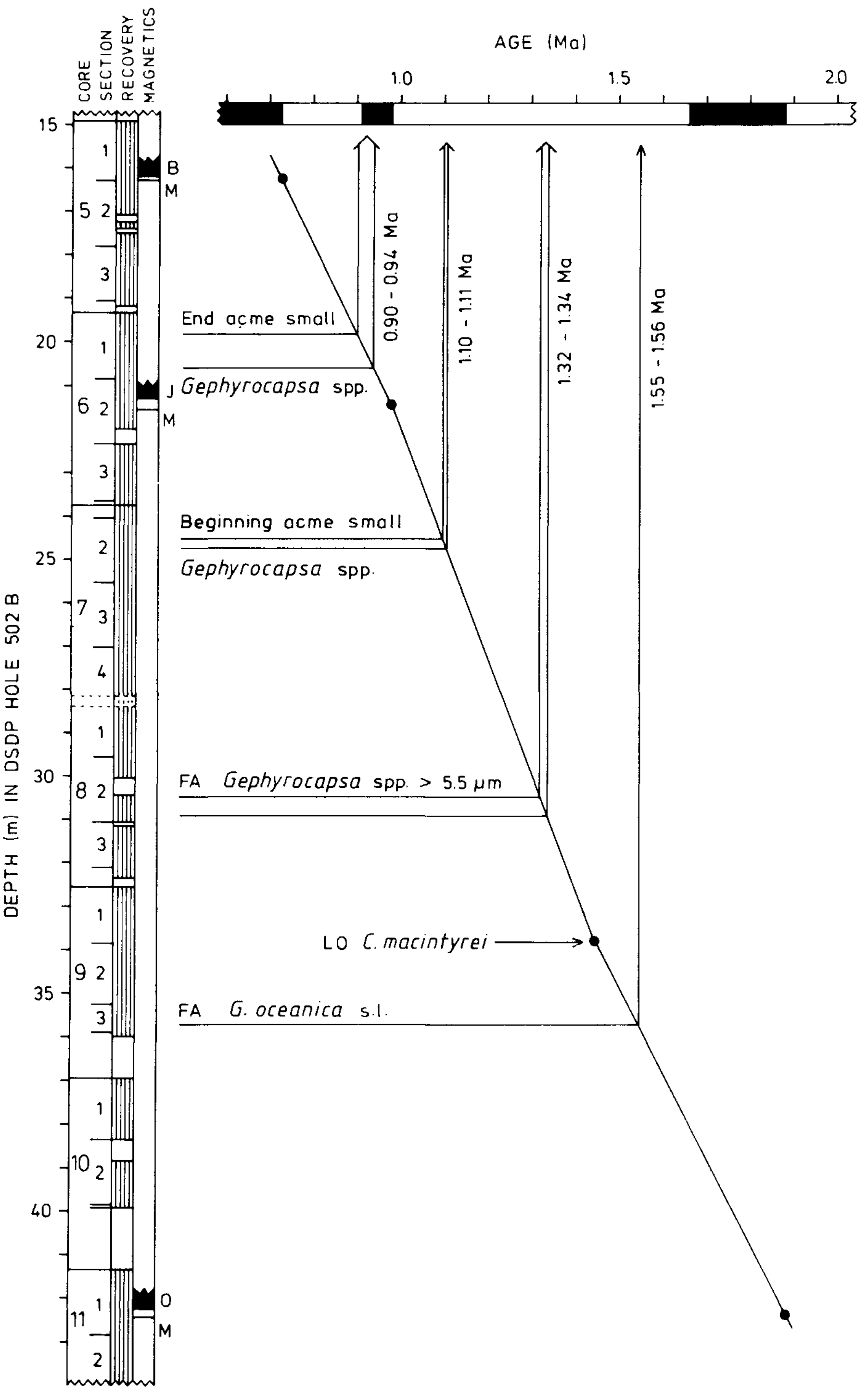
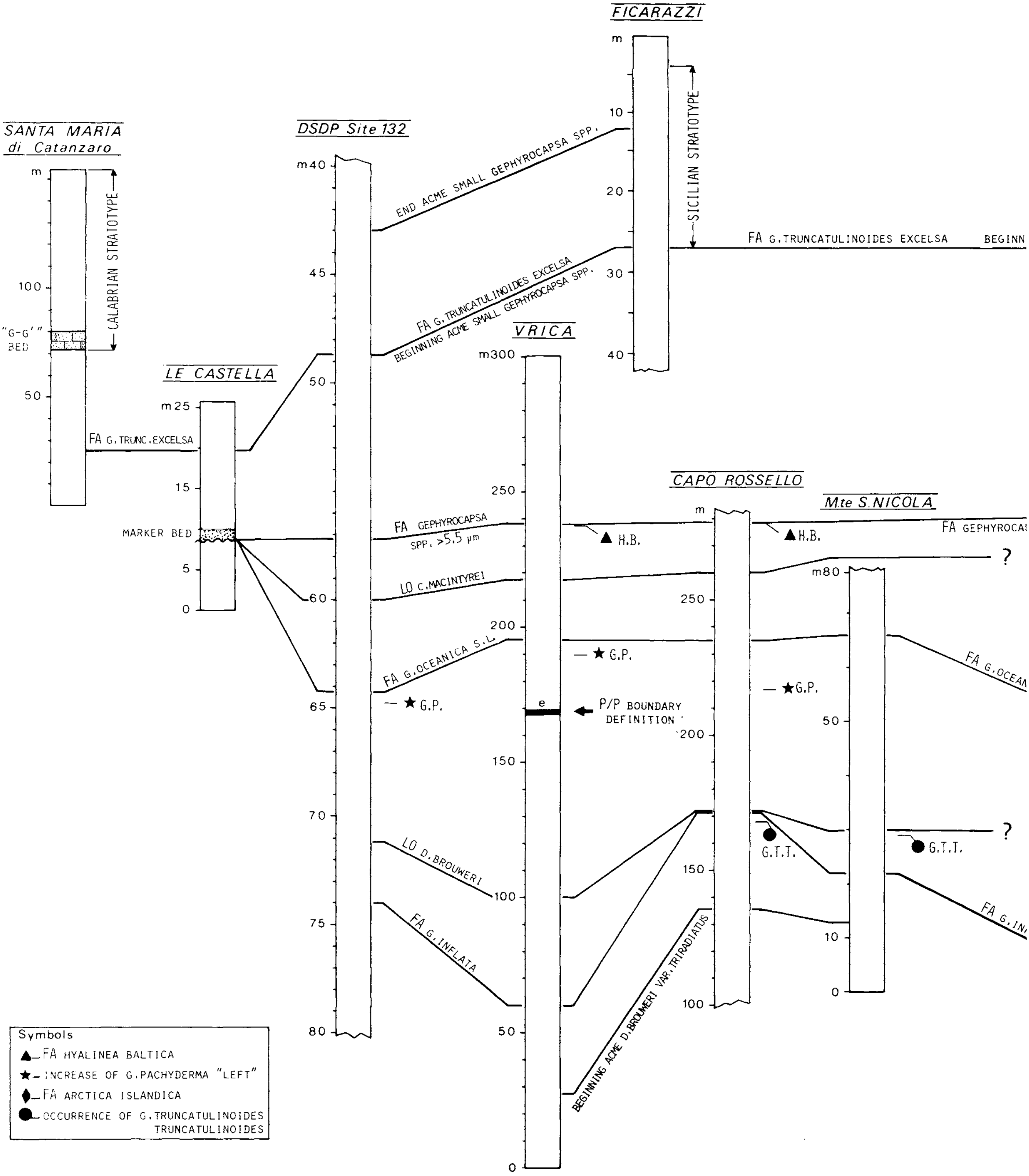


Figure 5.4. Sediment accumulation rate for DSDP site 502B plotted against nannofossil-datum chronology (uncorrected time scale).

species living in the region to this day, actually corresponds fairly well to the situation at the time of arrival of the "northern guests" in the Mediterranean (Pelosio et al., 1980).

The first faunal element that was proposed as a "northern guest," and indeed the most "popular" one, was the shallow-water pelecypod *Arctica islandica*. That form today has specific ecologic demands in that it thrives only in water depths less than 50–60 m, and consequently its fossils have only a limited stratigraphic applicability. For that reason stratigraphers were forced to use other criteria in order to recognize the base of the Pleistocene in sections where *A. islandica* was missing. As a substitute, they used benthic forms like the foraminifer *Hyalinea baltica* (Trevisan and Di Napoli, 1938) and the ostracode *Cytheropteron testudo* (Ruggieri, 1950, 1973), which retained the


FICARAZZI
m
10
20
30
40
SICILIAN STRATOTYPE
FA G.TRUNCATULINOIDES EXCELSA
BEGINN
SANTA MARIA di Catanzaro
m
100
"G-G'" BED
50
CALABRIAN STRATOTYPE
FA G.TRUNC.EXCELSA
DSDP Site 132
m40
45
50
60
65
70
75
80
END ACME SMALL GEPHYROCAPSA SPP.
FA G.TRUNCATULINOIDES EXCELSA
BEGINNING ACME SMALL GEPHYROCAPSA SPP.
LE CASTELLA
m25
15
MARKER BED
5
0
VRICA
m300
250
200
e
150
100
50
0
FA GEPHYROCAPSA SPP. >5.5 µm
LO C.MACINTYREI
FA G.OCEANICA S.L.
★ G.P.
P/P BOUNDARY DEFINITION
LO D.BROUWERI
FA G.INFLATA
BEGINNING ACME D.BROUWERI VAR.TRIRADIATUS
▲ H.B.
CAPO ROSSELLO
m
250
200
150
100
Mte S.NICOLA
m80
50
10
0
▲ H.B.
FA GEPHYROCA
?
★G.P.
FA G.OCEAN
● G.T.T.
● G.T.T.
?
FA G.IN
Symbols
▲_FA HYALINEA BALTICA
★_INCREASE OF G.PACHYDERMA "LEFT"
♦_FA ARCTICA ISLANDICA
●_OCCURRENCE OF G.TRUNCATULINOIDES TRUNCATULINOIDES

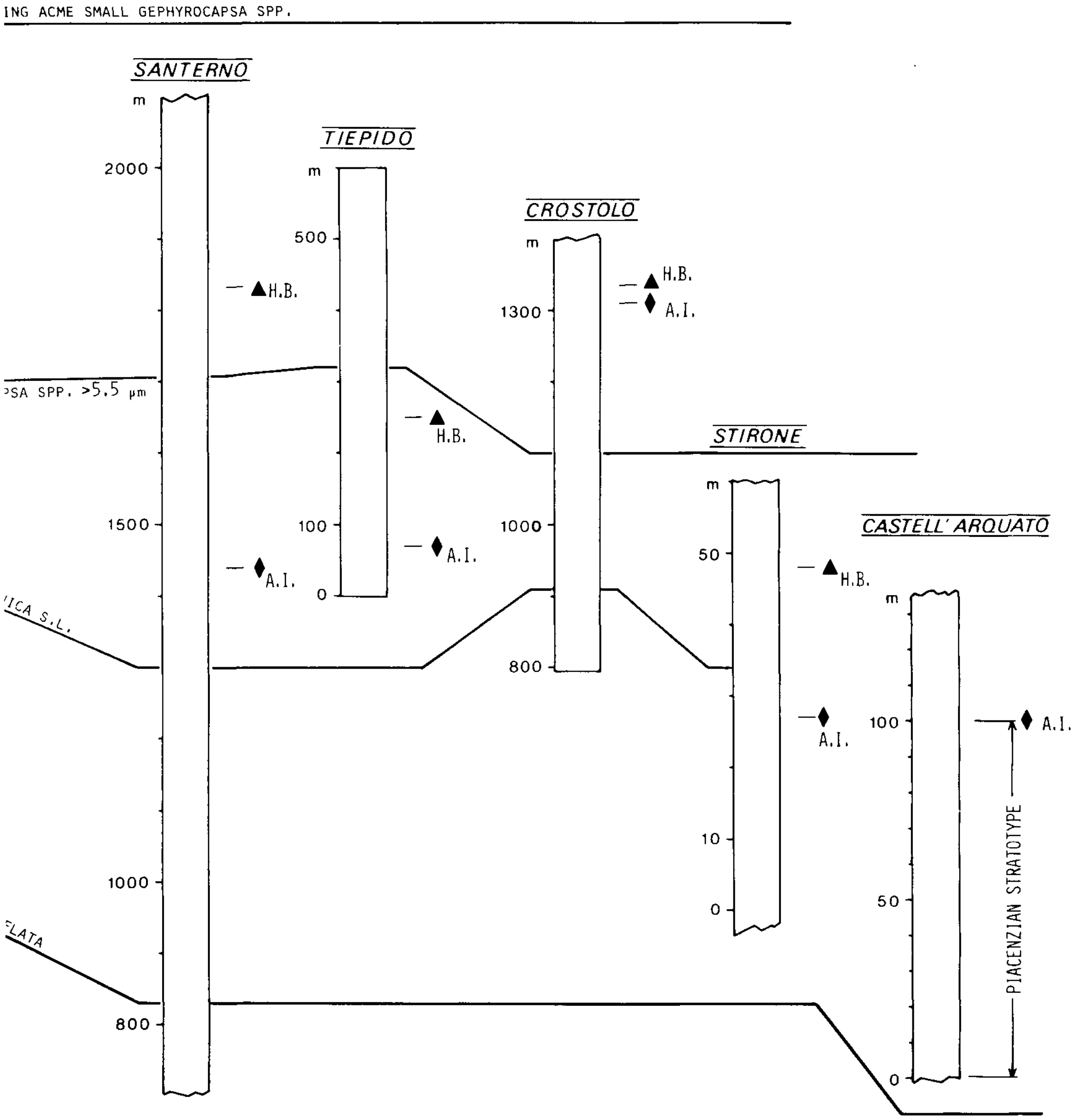


Figure 5.5. Correlation of classic Italian land sections to DSDP site 132 (note key to abbreviations). FA, first appearance.

Table 5.1. *Locations and previous studies of sections utilized for Gephyrocapsa spp. biochronology*

Section	Location	Lat.	Long.	Water depth (m)	Main studies
V28-239	Equatorial Pacific	3°15'N	159°11'E	3,490	Shackleton and Opdyke (1977), Backman and Shackleton (1983)
DSDP hole 502B	Colombia Basin, Caribbean Sea	11°29'N	79°23'W	3,051	Prell et al. (1982)
V12-18	Southwestern Atlantic	28°42'S	34°34'W	4,021	Backman and Shackleton (1983)
DSDP site 132	Mediterranean, Tyrrhenian Sea	40°15'N	11°25'E	2,813	Ryan and Hsu (1973), Raffi and Rio (1979)

"northern immigrant" concept. Evidence of climatic deterioration, such as the coiling change of *Neogloboquadrina pachyderma* (Dondi and Papetti, 1968) and the extinction of Taxodiaceae (Lona, 1962), also have been used as substitutes for the recognition of the Pliocene–Pleistocene boundary. The rationale underlying those choices was based on the misconcept that the earth passed from ice-free to glaciated conditions at the Pliocene–Pleistocene boundary. Such a threshold in the earth's history ought to have been detectable in every environment and therefore should have been correlatable as a time horizon. However, recent work has shown that the climatic history is much more complicated, and glacial conditions probably were brought on through a series of climatic steps (e.g., Shackleton and Kennett, 1975; Thunell and Williams, 1983). The appearance of the "northern guests" in the Mediterranean during the Pleistocene was gradual and discontinuous, occurring at different times for different taxa. In particular, the appearances of *A. islandica* and *H. baltica* did not represent a single event, and since both were environmentally controlled, these forms do not allow reliable correlations even within a small region. The correlations that are based on planktic forms (Figure 5.5) illustrate this point. Note the scattering of the appearances of *A. islandica* and *H. baltica* relative to the proposed correlations. It also appears from Figure 5.5 that the first appearance of *H. baltica* was clearly later than that of *A. islandica*. *H. baltica* occurs close to the appearance of large *Gephyrocapsa* spp. (larger than 5.5 μm), whereas *A. islandica* occurs below the first appearance of *G. oceanica s.l.*

If we want to adhere to the original concept that the base of the Pleistocene is to be tied to the first appearance of *A. islandica* in Italian sections, we need to locate, in a deeper-water boundary-stratotype section, a physical horizon that is time-equivalent with the first appearance of *A. islandica* in a shallow-water section. The stratigraphically lowest level where *A. islandica* is present seems to be represented in the Castell'Arquato and the Stirone sections. No reliable nannofossil data are available from the former section. In the Stirone section, we have observed that the appearance of *A. islandica* slightly predated the first occurrence of *G. oceanica s.l.* Below the levels containing the first *A. islandica*, abundant *Gephyrocapsa* spp. are present, including transitional forms indicating the first appearance of *G. oceanica s.l.* This suggests that the appearance of *A. islandica* is close to the top of the Olduvai subchron in this section. It is noteworthy that in both the Stirone and the Castell'Arquato sections the first appearance of *A. islandica* postdates the first occurrence of *Globorotalia inflata*. In Mediterranean sections the latter event occurs close to the first appearance of *Globorotalia truncatulinoides truncatulinoides*, which has been determined to have occurred just below the Olduvai subchron both in the Mediterranean (Rio et al., 1984a,b) and in the equatorial oceans (Rio, Fornaciari, and Raffi, 1990; Shackleton et al., 1995).

The INQUA Subcommission 1-d, "Pliocene/Pleistocene Boundary," adopted the proposal that the base of the shale bed overlying sapropelic layer *e* in the Vrica section should be used for definition of the Pliocene–Pleistocene boundary. This level is close to the top of the Olduvai subchron, according to magnetostratigraphy and biostratigraphy (as discussed later), suggesting that the boundary-definition proposal is appropriate in historical terms.

Prior to the introduction of the Vrica section as the location of the Pliocene–Pleistocene boundary-stratotype, two other definitions were in use: the base of the Calabrian Stage, provisionally defined in the Santa Maria di Catanzaro section (Selli, 1971), and the so-called marker bed in the Le Castella section (Berggren and Van Couvering, 1974; Haq et al., 1977).

The Santa Maria di Catanzaro section

Although there is a general agreement that the Santa Maria di Catanzaro section is unsuitable for definition of the base of the Pleistocene (Haq et al., 1977; Pelosio et al., 1980; Colalongo et al., 1982), this section needs some discussion because the Calabrian Stage was first introduced (Gignoux, 1913) with reference to this section, among many others. It was this section, however, and in particular the G-G' bed shown in a diagram by Gignoux (1913), that had been provisionally designated to represent the stratotype base of the Calabrian Stage (Selli, 1971; cf. Berggren and Van Couvering, 1979). However, the G-G' bed is well above the first Mediterranean occurrence of *Globorotalia truncatulinoides excelsa* (Colalongo, Pasini, and Sartoni, 1981), an event which occurred close to the beginning of dominance of small *Gephyrocapsa* spp. and which also marks the base of the Sicilian Stage (Di Stefano and Rio, 1981) or, preferably, substage (Ruggieri, Rio, and Sprovieri, 1984). The Calabrian, as exemplified at Santa Maria di Catanzaro, is thus completely overstepped by the Sicilian (Figure 5.5). Because the Sicilian was introduced earlier (Doederlein, 1872), it follows that the Calabrian is a junior synonym of the Sicilian and, as such, is an invalid stage (Ruggieri and Sprovieri, 1977). Note also in Figure 5.5 how the G-G′ bed represents a chronohorizon well above the base of the Pleistocene.

The Le Castella section

Many authors, according to the review by Rio et al. (1974), have used a sandy level, named "marker bed" by Emiliani, Mayeda, and Selli (1961), in the Le Castella section as defining the Pliocene–Pleistocene boundary. This level is unsuitable as a boundary definition because there is a hiatus immediately below it. The "marker bed" level is a key horizon associated with the first local occurrence of *H. baltica*, an event of somewhat uncertain value for recognizing the base of the Pleistocene (as discussed earlier). On biostratigraphic grounds, Haq et al. (1977) considered that the "marker bed" level in the Le Castella section is associated with the top of the Olduvai subchron and therefore proposed that this geomagnetic reversal should be considered as an index for the Pliocene–Pleistocene boundary. While this association has since been confirmed at Vrica (as discussed later), the data presented in Figure 5.5 and Table 5.3, together with the nannofossil biochronology previously discussed (Table 5.2), indicate that the age of the Le Castella "marker bed" is younger than 1.48–1.44 Ma, because of the presence of *Gephyrocapsa* spp. larger than 5.5 μm.

The Vrica section

Calcareous nannofossils have been studied in the Vrica section by Nakagawa (1977, 1981), Nakagawa et al. (1980), Cati and Borsetti (1980), Raffi and Rio (in Pasini and Calalongo, 1982), and Backman et al. (1983). The sample levels studied and the main findings reported by those authors are shown in Figure 5.6. Although the nannofossil assemblages in the Vrica section are variable in terms of abundance and preservation state, many samples contain abundant nannofossils showing good or excellent preservation. Reworked input appears to have been relatively stable (Backman and Shackleton, 1983), and the extinction levels of Pleistocene species can be accurately identified through the use of quantitative data and by taking into account the backgound "noise" of reworking.

The most abundant elements of the assemblages are represented by *Gephyrocapsa* spp., *Pseudoemiliana lacunosa* (with both circular and elliptical morphotypes), *Calcidiscus* spp., *Syracosphaera* spp., *Helicosphaera* spp., and *Coccolithus pelagicus*. Secondary elements are represented by *Cyclolithella annula, Rhabdosphaera* spp., *Umbilicosphaera* spp., *Pontosphaera* spp., and *Scyphosphaera* spp. The genus *Discoaster* is represented by rare reworked Tertiary specimens, and by rare *D. brouweri* and *D. brouweri triradiatus* in the lower part of the section. The genus *Ceratolithus* is practically absent from the section, as is the case in most Upper Pliocene and Lower Pleistocene sediments within the Mediterranean area.

Of the previously discussed nannofossil biochronologic indications, all except the interval of dominantly small *Gephyrocapsa* spp. are present in the Vrica section (Figure 5.6). Each of these is briefly discussed here.

Last occurrence of Discoaster brouweri. Except for Nakagawa and co-workers, all authors have considered the extinction of *D. brouweri* in their range charts. No discrepancy is noted in the positions of this datum event when the different sampling intervals are taken into account. Combining the Raffi and Rio data with those of Backman et al. (1983) results in a location between 101 and 105 m for this datum event in the Vrica section.

Discoaster brouweri triradiatus abundance interval. Backman et al. (1983) provided counts of the abundance of this form in the Vrica section and showed that its last occurrence coincides with that of *D. brouweri*, in agreement with the findings of Backman and Shackleton (1983) from extra-Mediterranean areas and the findings from other Mediterranean sections (Rio, 1982). The increase in abundance of *D. brouweri triradiatus* relative to *D. brouweri* in the Vrica section begins between 20 and 35 m, according to Raffi and Rio's sampling.

Last occurrence of Calcidiscus macintyrei. The findings of Raffi and Rio, on one hand, and Backman and co-workers, on the other, are in full agreement regarding the extinction level of *C. macintyrei* (considering the different sampling intervals). Cati and Borsetti suggest a slightly higher extinction level, and Nakagawa and co-workers report *C. macintyrei* throughout the Vrica section. These different ranges are best explained by reworking, as demonstrated by Backman et al. (1983). By combining the sampling levels of Backman and colleagues and those of Raffi and Rio, the last occurrence of *C. macintyrei* in the Vrica section can be located to 216 m.

Range of Helicosphaera sellii. All nannofossil biostratigraphers except Cati and Borsetti consider *H. sellii* to be present thoughout the Vrica section. Cati and Borsetti have suggested that this species is absent from the lowermost and uppermost parts of the section. The reason for this discrepancy in observation is unknown. Neverthless, we consider that the top of the Vrica section is below the regional last occurrence of *H. sellii*. In other Mediterranean sections the disappearance of *H. sellii* occurs together with the temporary disappearance of normal-sized *Gephyrocapsa* spp. (which is the beginning of the

Table 5.2. *Sections studied in this chapter*

Section	Field stratigraphic information	Nannofossil data	Other information
Castell'Arquato (Piacenza)	Pelosio et al. (1980)		Piacenzian stratotype section (Barbieri, 1971)
Stirone (Parma)	Bertolani Marchetti et al. (1979)	Pelosio et al. (1980), Rio et al. (this chapter)	
Crostolo (Reggio Emilia)	Barbieri and Petrucci (1967)	Raffi and Rio (1980a)	
Tiepido (Modena)	Annovi et al. (1979)	Raffi and Rio (1980b), Rio et al. (this chapter)	
Santerno (Bologna)	Colalongo et al. (1974)	Raffi and Rio (1980c), Rio et al. (this chapter)	Santernian and Emilian stratotype sections (Ruggieri and Sprovieri, 1977)
Monte San Nicola (southern Sicily)	Rio et al. (1984a)	Rio et al. (1984a)	
Capo Rossello (southern Sicily)	Rio et al. (1984b)	Rio et al. (1984b)	Rossellian superstage stratotype section (Cita and Decima, 1975)
Ficarazzi (Palermo, Sicily)	Di Stefano and Rio (1981)	Di Stefano and Rio (1981)	Sicilian stratotype section (Ruggieri and Sprovieri, 1977), proposed Lower–Middle Pleistocene boundary-stratotype section (Ruggieri et al., 1984)
Vrica (Calabria)	Selli et al. (1977)	Raffi and Rio (in Pasini and Colalongo, 1982), Backman and Shackleton (1983)	Pliocene–Pleistocene (P/P) boundary-stratotype section
Le Castella (Calabria)	Raffi and Rio (1980d), Colalongo et al. (1981)	Raffi and Rio (1980d)	Proposed P/P boundary-stratotype section (Pelosio et al., 1980)
Santa Maria di Catanzaro (Calabria)	Colalongo et al. (1981)		Calabrian stratotype section

interval of dominantly small *Gephyrocapsa* spp.) and the appearance of *Globorotalia truncatulinoides excelsa* (Raffi and Rio, 1979; Di Stefano and Rio, 1981). Neither of these two events is recorded in the Vrica section.

Gephyrocapsa group. Range charts for the *Gephyrocapsa* group have been used by all authors who have studied the Vrica section except Backman et al. (1983). Identifications in this group have been unstable because of taxonomy and nomenclature problems, and thus it is not surprising that conflicting findings have been reported by different authors. In order to apply consistent taxonomic concepts, we have adopted the informal (morphometric) taxonomy of Rio (1982). It should be emphasized that we have applied this informal taxonomy on all material studied, thus avoiding taxonomy and nomenclature problems when evaluating the biochronology of the datum events provided by the *Gephyrocapsa* group.

Nannofossil biochronology of the Vrica section and the age of the proposed Pliocene–Pleistocene boundary-stratotype

The INQUA Subcommission 1-d, "Pliocene/Pleistocene Boundary," meeting in Madrid in 1983, proposed that the base of the

Table 5.3. *Summary of field stratigraphic and biostratigraphic studies used in the compilation of Figure 5.5*

	Age estimates (Ma) for nannofossil datum events around the Plio–Pleistocene boundary			
Datum event	V28-239 (western eq. Pacific)	DSDP 502B (Caribbean Sea)	V12-18 (southwestern Atlantic)	DSDP 132 (Mediterranean Sea)
End of acme of small *Gephyrocapsa* spp.	0.93–0.95	0.90–0.94	–	–
Beginning of acme of small *Gephyrocapsa* spp.	1.13–1.15	1.10–1.11	–	1.08–1.14
First appearance of *Gephyrocapsa* spp. > 5.5 µm	1.31–1.32	1.32–1.34	1.29–1.31	1.32–1.36
First appearance of *Gephyrocapsa oceanica s.l.*	1.57–1.61	1.55–1.56	1.55–1.59	1.56–1.62
Last occurrence of *Calcidius macintyrei*	Synchronous worldwide: 1.45–1.46 Ma^{+}			
Last occurrence of *Discoaster brouweri*	Synchronous in low-latitude and midlatitude oceans, and high latitudes in North Atlantic: 1.89 Ma (Backman and Shackleton, 1983)			
Last occurrence of *D. brouweri* var. *triradiatus*	Synchronous in low-latitude and midlatitude oceans, and high latitudes in North Atlantic: 189 Ma (Backman and Shackleton, 1983)			
Increase in proportion of *D. brouweri* var. *triradiatus*	Occurs in all *Discoaster*-bearing sediments: 2.0–2.1 Ma (Backman and Shackleton, 1983)			

stratum overlying layer *e* in the Vrica section should be used for definition of the Pliocene–Pleistocene boundary. As noted by Nikiforova and Alekseev (Chapter 1, this volume), that proposal has now been adopted by the IUGS authorities. This level is only 8 m below the level where the cold-water ostracode *Cytheropteron testudo* is first recorded, which previously was proposed as a boundary definition (Colalongo et al., 1982; cf. Pasini and Colalongo, Chapter 2, this volume). Considering the high rates of sediment accumulation in the Vrica section, the two biochronohorizons are virtually indistinguishable (Figure 5.7).

The last occurrence of *D. brouweri* (and *D. brouweri triradiatus*) is associated with the bottom of the Olduvai subchron, as in all low-latitude and mid-latitude deep-sea cores. The first appearance of *G. oceanica s.l.* occurs immediately above the Olduvai subchron, which is closely followed by the extinction of *C. macintyrei* and the first appearance of *Gephyrocapsa* spp. larger than 5.5 μm. The sequencing of these datum events and their estimated ages, in terms of sedimentation rates relative to the magnetostratigraphy at Vrica, correspond precisely with the findings from the Pacific, the Caribbean, and the Atlantic (Shipboard Scientific Party, 1992), giving the marker bed *e* an unarguable biostratigraphic, biochronologic, and magnetostratigraphic context. From this context, the proposed Pliocene–Pleistocene boundary can be accurately identified in both marine and continental geologic records, within the Mediterranean as well as in extra-Mediterranean areas.

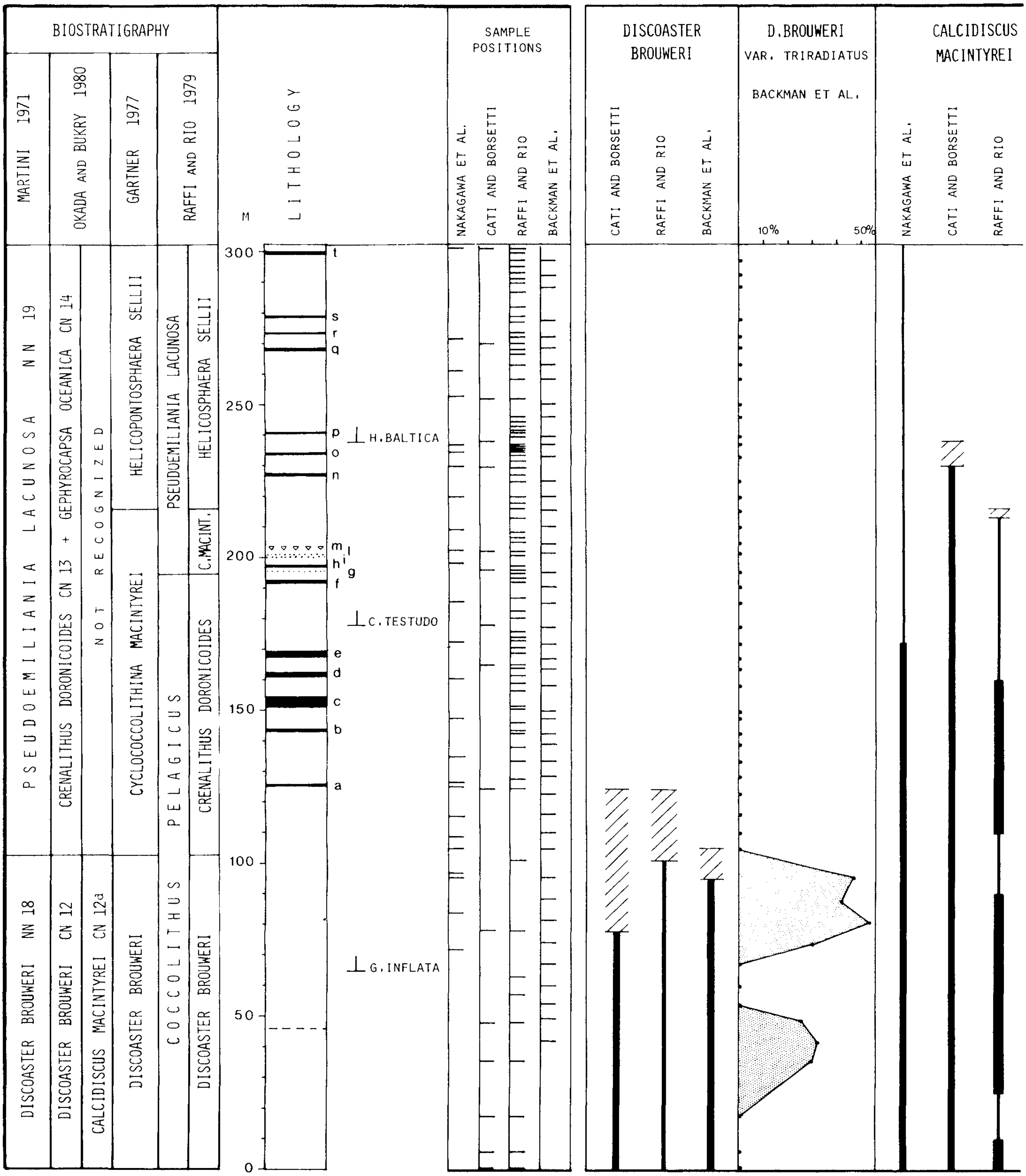
Conclusions

Several calcareous nannofossil datum events in Upper Pliocene and Lower Pleistocene marine sediments are well documented with regard to the magneostratigraphic time scale, from which their ages are estimated. All datum events observed in the investigated material occur without notable diachronism from the global standard.

A cursory investigation of the biostratigraphy and lithostratigraphy of the proposed Pliocene–Pleistocene boundary in the Vrica section affirms that it is historically appropriate in that it respects the concepts of the Pliocene and the Pleistocene as originally conceived by Lyell (1833).

As presently defined, the Calabrian Stage is a junior synonym for the Sicilian, and furthermore its base is unsuitable for definition of the base of the Pleistocene, according to the paleoclimatic concept that has guided the international effort begun at the London (1948) International Geological Congress.

The "marker bed" in the Le Castella section that had been proposed previously to be the definition of the Pliocene–Pleistocene boundary (Haq et al., 1977; Pelosio et al., 1980) cannot be used for this purpose because there is a hiatus immediately below it. Moreover, the only key event associated with the "marker bed" is the first (apparent) local occurrence of *Hyalinea baltica*, which is an unreliable and biostratigraphically variable event. Calcareous nannofossils occur abundantly in the proposed Vrica stratotype section, and all critical Upper Pliocene and Lower Pleistocene nannofossil datum events that


BIOSTRATIGRAPHY
MARTINI 1971
OKADA AND BUKRY 1980
GARTNER 1977
RAFFI AND RIO 1979
PSEUDOEMILIANIA LACUNOSA NN 19
DISCOASTER BROUWERI NN 18
CRENALITHUS DORONICOIDES CN 13 + GEPHYROCAPSA OCEANICA CN 14
NOT RECOGNIZED
DISCOASTER BROUWERI CN 12
CALCIDISCUS MACINTYREI CN 12a
HELICOPONTOSPHAERA SELLII
CYCLOCOCCOLITHINA MACINTYREI
DISCOASTER BROUWERI
PSEUDOEMILIANIA LACUNOSA
COCCOLITHUS PELAGICUS
HELICOSPHAERA SELLII
C.MACINT.
CRENALITHUS DORONICOIDES
DISCOASTER BROUWERI
M
LITHOLOGY
300
250
200
150
100
50
0
t
s
r
q
p
o
n
m
l
i
h
g
f
e
d
c
b
a
H.BALTICA
C.TESTUDO
G.INFLATA
SAMPLE POSITIONS
NAKAGAWA ET AL.
CATI AND BORSETTI
RAFFI AND RIO
BACKMAN ET AL.
DISCOASTER BROUWERI
CATI AND BORSETTI
RAFFI AND RIO
BACKMAN ET AL.
D.BROUWERI VAR. TRIRADIATUS
BACKMAN ET AL.
10%
50%
CALCIDISCUS MACINTYREI
NAKAGAWA ET AL.
CATI AND BORSETTI
RAFFI AND RIO

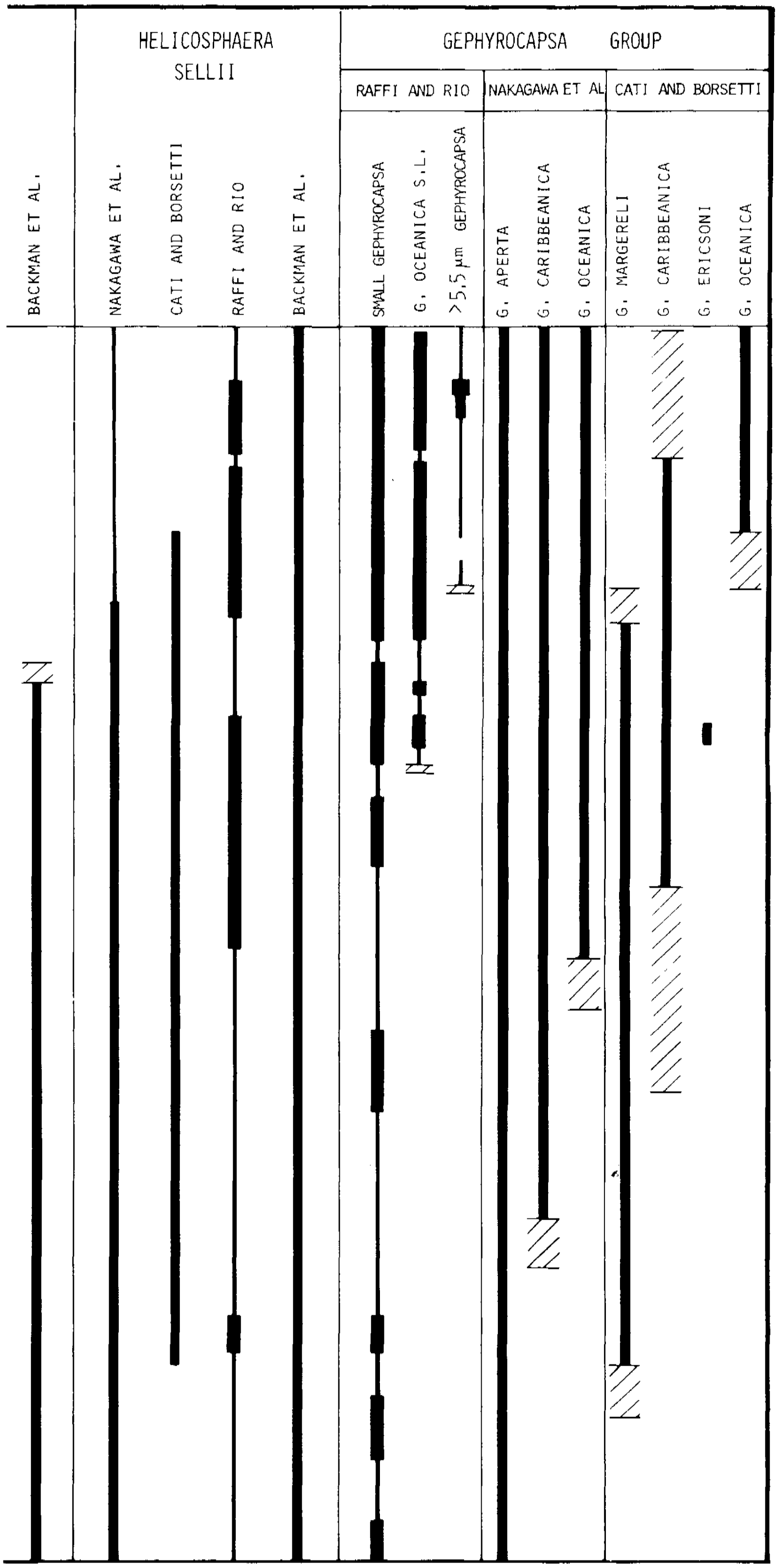


Figure 5.6. Summary of nannofossil biostratigraphy in the Vrica section.

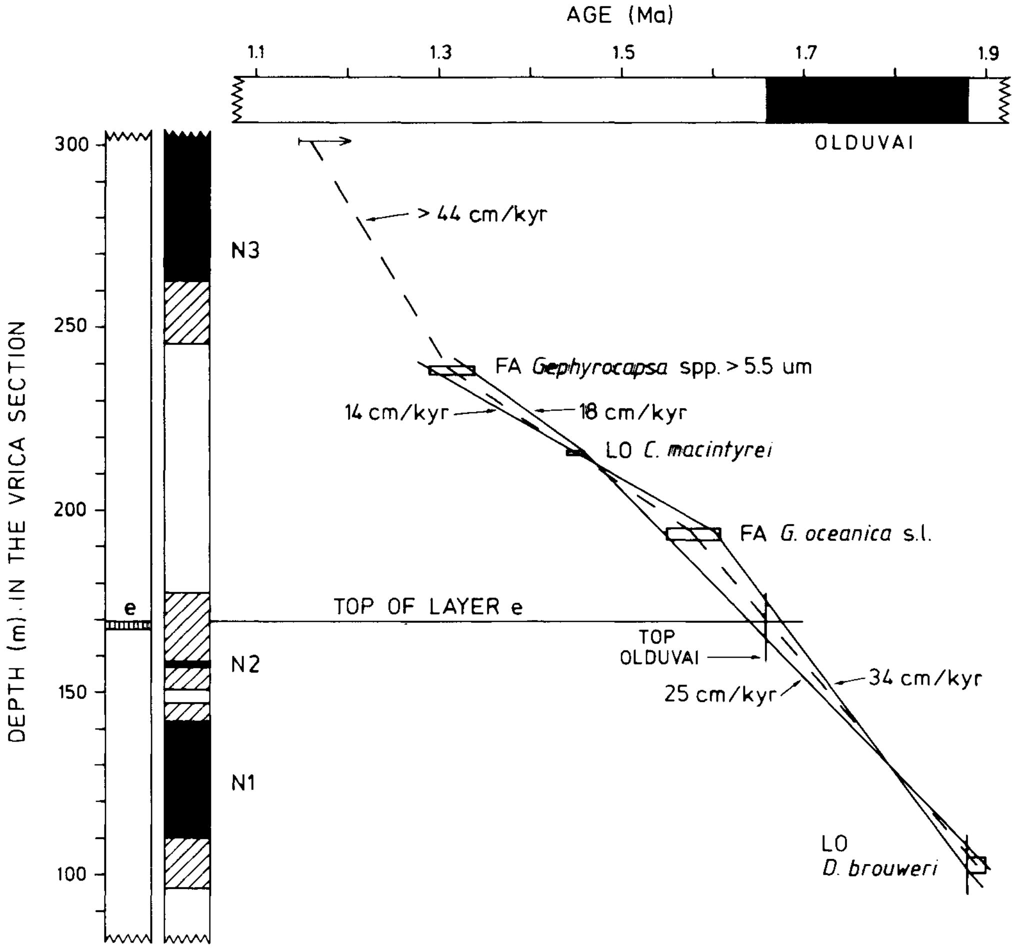


Figure 5.7. Biochronology of the Pliocene–Pleistocene boundary in the Vrica section (uncorrected time scale).

can be used as biochronologic indications are present. By combining these with the established magnetostratigraphy of the section, the Pliocene–Pleistocene boundary in the Vrica stratotype can be correlated to most other marine sections by means of biostratigraphy, biochronology, and magnetostratigraphy. Magnetostratigraphy also allows correlation to continental biochronology.

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6 Diatom microfossils from the Vrica section

ANDREW J. M. PALMER

Introduction

The marine strata exposed at Vrica, in Calabria, were proposed by Selli et al. (1977) as the Pliocene–Pleistocene boundary-stratotype section because those strata, originally deposited at depths between 500 and 800 m, offered the potential for correlation of oceanic microfossils with established datum levels. Backman, Shackleton, and Tauxe (1983) and Rio, Raffi, and Backman (Chapter 5, this volume) correlated the calcareous nannofossils from Vrica to those recovered from deep-sea piston core V28–239, taken at a depth of 3,490 m on the Solomon Rise in the western equatorial Pacific. In that biostratigraphy, supported by observations from other cores, the first appearance of *Gephyrocapsa s.l.* just above the Olduvai subchronozone and the global extinction of *Calcidiscus macintyrei* at a slightly higher level are the closest available approximations of the boundary, which is located within the uppermost part of the Olduvai at approximately 1.8 Ma (Pasini and Colalongo, Chapter 2, this volume), and not above the subchron, according to the data of Tauxe et al. (1983) (Figure 6.1).

Diatom assemblages of Plio–Pleistocene age have been reported from deep-sea sites around the globe, with emphasis on the Pacific Ocean (Burckle and Todd, 1976; Barron, 1980a,b; Koizumi, 1985). More detailed studies in the northwest Pacific (Koizumi, 1985, 1992), Kamchatka (Gladenkov, 1994), northeast and southern Pacific (Schrader, 1973, 1976), and the Southern Ocean (McCollum, 1975; Gombos, 1976) support the regional analysis. Research by Burckle and co-workers (e.g., Saito, Burckle, and Hays, 1975; Burckle and Opdyke, 1977; Burckle and Trainer, 1979) was based on integration of diatom datum levels and magnetostratigraphy. Burckle and Trainer (1979) noted four features in central Pacific diatom assemblages (Figure 6.2) which have proved to be of widespread significance (Shackleton et al., 1995a,b). These are as follows:

1. the marked decline of *Coscinodiscus nodulifer* var. *cyclopus* starting below the Olduvai subchronozone
2. the consistent abundance of *Rhizosolenia praebergonii* var. *robusta* beginning just below the base of the Olduvai subchronozone
3. within the Olduvai, the last appearance of *Rhizosolenia praebergonii* and the first appearance of *Pseudoeunotia doliolus*
4. the last appearance of *Rhizosolenia praebergonii* var. *robusta* just above the Olduvai subchronozone

In the study reported here, I examined sediment samples from the Vrica section for diatom assemblages that might be useful in correlating onshore (marine) sediments with the plankton stratigraphy established in offshore (marine) sediments, following the example set by Koizumi, Barron, and Harper (1980).

Sample material

Twenty-four sediment samples from three laminated sapropelic units (*e*, *f*, and *h*) (Raffi and Thunell, Chapter 4, this volume) of the Vrica section (Figure 6.3) were prepared for diatom analysis. The material was dispersed in hydrogen peroxide and dilute hydrochloric acid, washed by a simple settling technique to minimize loss of any opal, and subsequently examined in a high-refractive-index mounting medium under oil immersion at 1,000×. Two samples were found to be diatomaceous, both from level *e*. The lower level was distinguished here as *e*5, and the upper as *e*6. Fecal pellets were not common, and the samples were not difficult to disaggregate. Backman et al. (1983) refer to evidence that "the whole section is marked by very extensive . . . influx of [reworked] Palaeogene and Cretaceous nannofossils." Diatoms are subject to the same general taphonomic constraints as calcareous nannofossils, and the possibility of redeposited specimens was kept in mind during this study.

Observations of assemblage composition

Two sets of observations were made: a quantitative count of 186 specimens from sample *e*6 in a larger population randomly mounted upon a microscope slide, and a qualitative examination of correlative samples mounted on SEM (scanning electron microscope) stubs. Under light microscopy, fourteen taxa plus undifferentiated spores and "unknowns" collectively constituted more than 95% of the siliceous microfossil sample, as indicated in Table 6.1. All of the fossil diatoms belonged to species living in

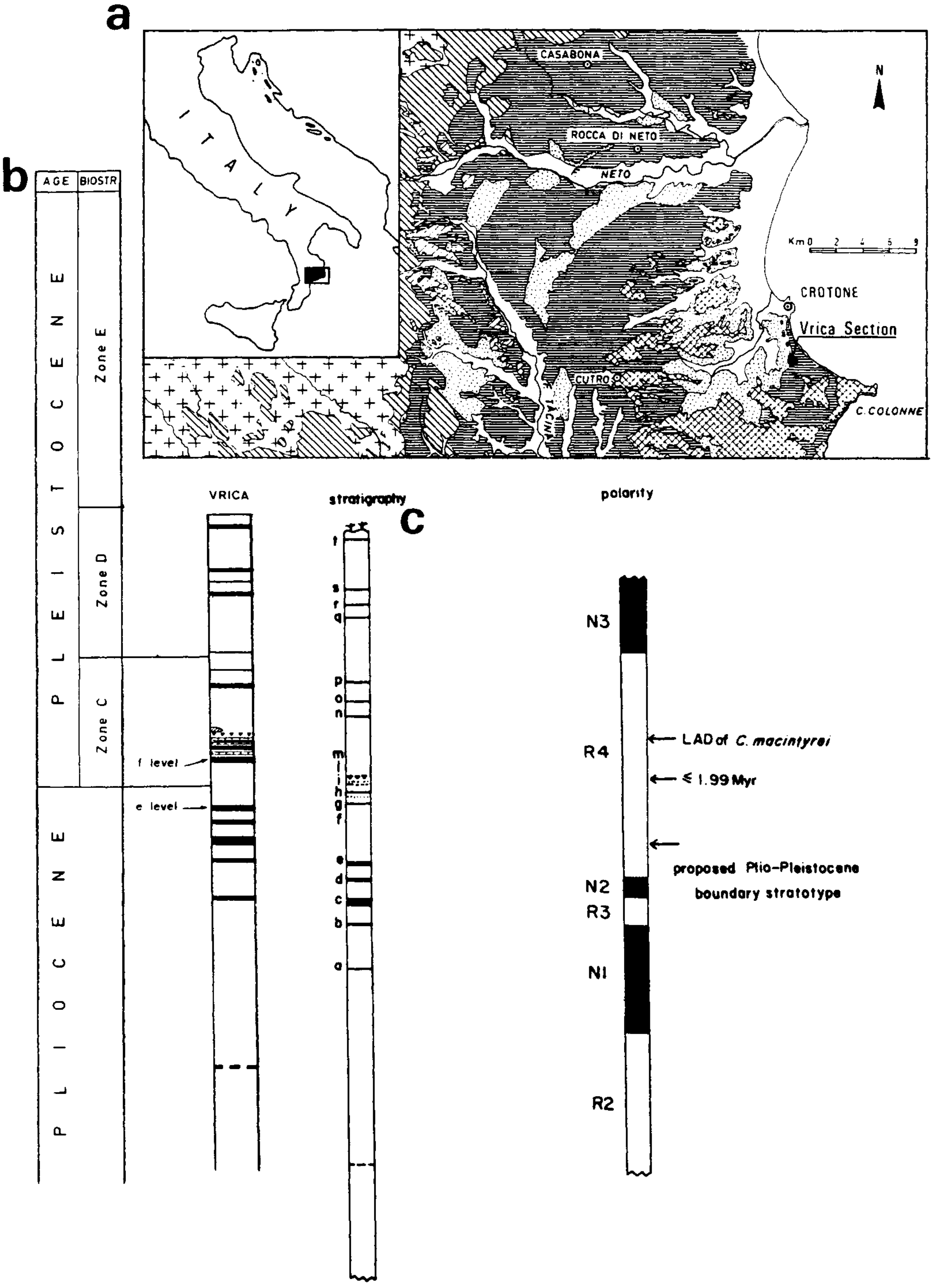

Figure 6.1. Geologic setting of the Vrica section: (a) location map and (b) stratigraphy, from Colalongo et al. (1982); (c) magnetostratigraphy of the Vrica section, from Tauxe et al. (1983). The boundary proposed in 1982 is shown at the first appearance of *Cytheropteron testudo*, whereas the boundary that was finally adopted is at the top of level *e*. The N1-N2 normal-polarity intervals are assigned to the Olduvai event (Tauxe et al., 1983; Nakagawa et al., Chapter 3, this volume).

modern marine environments, as noted in the table (Schrader and Matherne, 1981). Specimens of the colonial forms *Fragilaria* and *Melosira* probably originated in coastal environments and were transported offshore; their recovery in samples dominated by planktonic diatoms was not surprising (Hendey, 1964). The recovery of the colonial diatoms was remarkable, however, in that they were observed as intact chains, whereas adjacent samples were barren of any diatoms. Furthermore, there was no evidence of reworked pre–late Cenozoic diatoms in the prepared sample. Unfortunately, however, none of the taxa that have been considered here as key, magnetostratigraphically correlated markers were found. Several palynomorphs (including *Pinus* spp.) were observed, consistent with the details discussed by Selli et al. (1977).

SEM micrographs revealed that some specimens had suffered solution along exposed edges, particularly *Coscinodiscus*. While

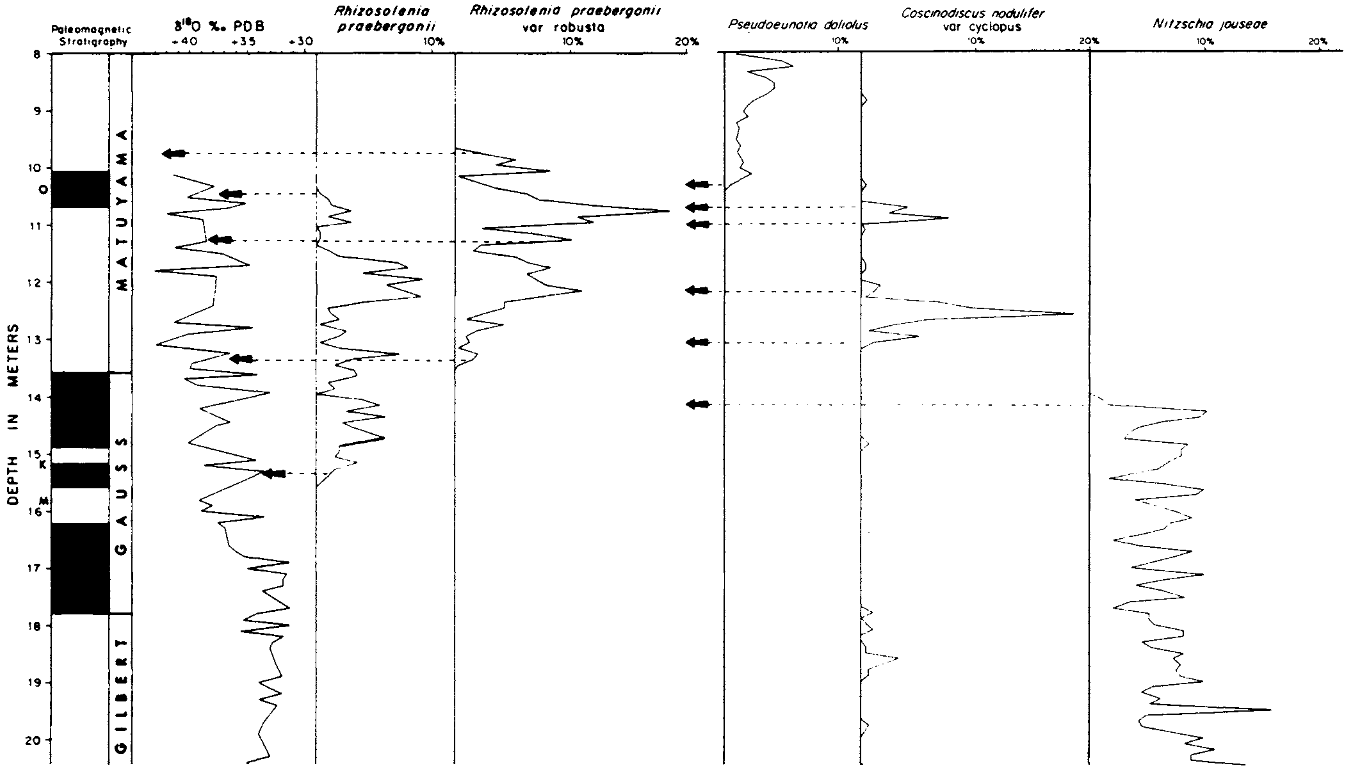

Figure 6.2. Integrated Plio–Pleistocene magnetostratigraphy and marine diatom stratigraphy from the equatorial Pacific. Note the level of the Olduvai normal polarities, marked with the letter O on the depth scale. (From Burckle and Trainer, 1979, with permission.)

Table 6.1. *Major and minor components of a diatom assemblage (n = 186) from Vrica level e6*

Taxon	Trophic Group	Count	Percentage[a]
Coscinodiscus obscurus	ocean plankton	62	33.3 ± 6.8%
Coscinodiscus radiatus	ocean plankton	23	12.4 ± 4.8%
(spores, undifferentiated)	plankton	18	9.7 ± 4.3%
Rhizosolenia hebetata f. *semispina*	temperate plankton	15	8.1 ± 3.9%
Thalassiosira eccentrica	neritic plankton	12	6.5 ± 3.5%
Coscinodiscus oculus-iridis	ocean plankton	6	3.2 ± 2.5%
Coscinodiscus normani	plankton	6	3.2 ± 2.5%
Melosira sp. (chains in girdle view)	sessile[b]	6	3.2 ± 2.5%
Fragilaria sp.	neritic plankton	5	2.7 ± 2.3%
Thalassiothrix mediterraneana	neritic plankton	5	2.7 ± 2.3%
Nitzschia sp.	ocean plankton	5	2.7 ± 2.3%
Coscinodiscus stellaris	ocean plankton	4	2.2 ± 2.1%
Actinoptychus senarius	neritic plankton	4	2.2 ± 2.1%
(unknowns)		4	2.2 ± 2.1%
Thalassiosira lineata	plankton	3	1.6 ± 1.8%
Hemidiscus cuneiformis	ocean plankton	3	1.6 ± 1.8%

[a] The calculated percentage is indicated at the 95% level of confidence.

[b] *Melosira* chains also found in plankton.

Source: Adapted from Galehouse (1971).

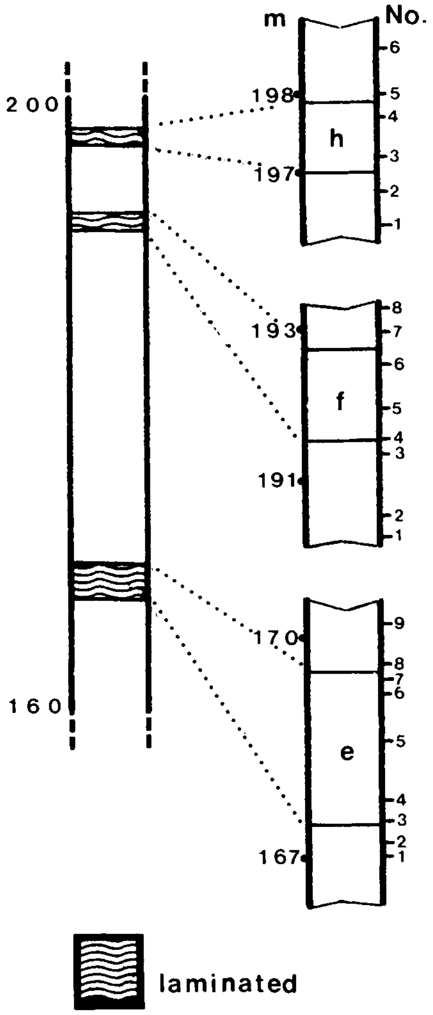

Figure 6.3. Stratigraphic distribution of diatom samples from the Vrica section. Thicknesses (in meters) accord with the section measured by Selli et al. (1977).

the overall dissolution was not severe, most of the sieve plates on the *Coscinodiscus* specimens were heavily eroded.

The diatom assemblage from sample *e*6 at Vrica is overwhelmingly dominated by oceanic and/or neritic plankton, with a minor contribution from coastal benthic diatoms. On the basis of these limited observations, it is not possible to reach conclusions about preferential dissolution. However, it can be assumed that the dissolution occurred either within the water column or at the sediment–water interface during sediment accumulation, rather than as a diagenetic change after burial. The generally good condition of the specimens, however, and the intact "chains" of colonial forms suggest that environmental conditions during accumulation of the *e*6 laminates were anomalous, in that the diatoms were relatively undisturbed before fossilization.

The diatom assemblage in sample *e*6 is dominated by *Coscinodiscus obscurus* (cf. Hustedt, 1971, figure 224), *Coscinodiscus radiatus* (cf. Hustedt, 1971, figure 225; Hendey, 1964, pl. 22), undifferentiated spores, *Rhizosolenia hebetata* forma *semispina*, and *Thalassiosira eccentrica*. Other diatoms noted are *Coscinodiscus asteromphalus* (cf. Hustedt, 1971, figure 250; Hendey, 1964, pl. 24) and *Coscinodiscus normani* (cf. Hendey, 1964, p. 80; Hustedt, 1971, as *C. rothi* var. *normani*, figure 213; Schmidt et al., 1972, as *Coscinodiscus fasciatus*). The assemblage does not appear to contain any of the equatorial Pacific taxa discussed by Burckle and Trainer (1979), such as *Coscinodiscus nodulifer* var. *cyclopus*, *Rhizosolenia praebergonii* (with the variety *robusta*), and *Pseudoeunotia doliolus*. Correlation between the diatom biostratigraphy associated with the Olduvai subchron in Pacific Ocean cores and that of the Vrica sample *e*6 is therefore impossible. The diatom assemblages in the deep-sea sapropels of the eastern Mediterranean examined by Schrader and Matherne (1981) were dominated by such taxa as *Rhizosolenia calcaravis*, *Thalassionema nitzschioides*, and *Thalassiosira oestrupii*. Again, there was a poor match between those elaborately quantified diatom assemblages and the diatoms recovered from the laminates of the Vrica section.

Conclusions

Although diatom biostratigraphic correlation of the Vrica section with other Plio–Pleistocene deep-sea sequences is severely limited by the absence of key taxa, some paleoenvironmental inferences are possible. At modern rates of surface production, diatoms are rarely preserved in the bottom sediments of the Mediterranean (Schrader and Matherne, 1981), apparently because the water is not sufficiently close to saturation with respect to opaline silica. It can be inferred that similar conditions prevailed during much of the time represented in the Vrica sediments. The anomalous recovery of diatom microfossils in Vrica samples *e*6 and *e*5 suggests two possible explanations: either an increase in surface productivity or a major change in the water chemistry of the Crotone Basin, either of which would have led to a transient saturation with dissolved silica. The first explanation is favored by a considerable body of information suggesting that surface production in the Mediterranean was sharply increased during deposition of the laminated sediments (Raffi and Thunell, Chapter 4, this volume), whereas there is no good evidence in favor of transient major changes in the global deep-sea silica budget during that interval.

The low diversity (not more than 15 distinct taxa in my count) indicates, however, that conditions apparently did not support very complex original populations. *Thalassionema nitzschioides* is a dissolution-resistant diatom (Schrader and Matherne, 1981) sometimes used informally as an indicator of upwelling (L. H. Burckle, personal communication, 1982). Its absence from Vrica sample *e*6 suggests, but of course does not prove, that an episode of upwelling was not the direct cause of the increased production indicated for this sample.

The interpretation of diatom microfossils from Vrica sample *e*6 is incomplete and probably will remain so until correlative recoveries are made. Nonetheless, it is fair to state that (1) there is no apparent component of redeposited Paleogene diatoms, (2) the dominance of marine plankton is consistent with a 500–800-m water column as inferred by Selli et al. (1977), (3) there is no immediate correlation with deep-sea diatom biostratigraphy, and (4) the assemblage represents a substantial short-term increase in latest Pliocene sea-surface productivity.

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Part III

The paleontological context of the Pleistocene boundary

7 The Pliocene–Pleistocene boundary in deep-sea sediments

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Introduction

This chapter is a summary of the major micropaleontological events that are represented in the stratigraphic record of deep-sea sediments over the past 2.5 m.y. The main focus is on those events associated with the Olduvai subchron, inasmuch as this is the time interval in which the Pliocene–Pleistocene boundary has now been located by formal action of the IUGS. Our discussion concentrates on three areas of particular interest: biostratigraphy, the history of oxygen-isotope variations, and carbonate stratigraphy.

Biostratigraphy

Boundary-stratotype section

Current studies on the proposed Vrica boundary-stratotype (Pasini and Colalongo, Chapter 2, this volume; Nakagawa et al., Chapter 3, this volume; Rio, Raffi, and Backman, Chapter 5, this volume; Palmer, Chapter 6, this volume) indicate that this definition places the Pliocene–Pleistocene boundary just above (or just within) the top of the Olduvai subchron of the Matuyama chron, with an estimated age of 1.81 Ma (Zijderveld et al., 1991; Pasini and Colalongo, Chapter 2, this volume), and that this level is bracketed below by the extinction, or LAD (last-appearance datum), of *Discoaster brouweri* and its triradiate variety, and above by the FAD (first-appearance datum) of *Gephyrocapsa oceanica s.l.* and the LAD of *Calcidiscus macintyrei*. The same relationships are seen in deep-sea cores (Figure 7.1). In foraminifera biostratigraphy, the boundary-stratotype is almost exactly coincident with the FAD of *Globigerina cariacoensis*. Other correlated events are the shift from dextral to sinistral coiling in most *Neogloboquadrina pachyderma*, the FAD of *Globorotalia umbilicata*, and (at a slightly earlier age) the last occurrence of *Neogloboquadrina atlantica* and first occurrence of *Globigerina digitata digitata*.

The deep sea

A number of biostratigraphic events have now been reliably correlated to the Cenozoic paleomagnetic stratigraphy in deep-sea cores over large latitudinal spans of the world's oceans. In Figure 7.2, we have plotted the better-documented datum levels that reflect changes in calcareous and siliceous plankton over the past 2.5 m.y. Owing to the sequential and relatively rapid changes that occurred in Pliocene paleogeography and climatology and, concomitantly, biogeography, some of these datum events are geographically restricted, whereas others are mondial in nature (Figure 7.2). It will be seen that the Olduvai subchron is bracketed by a number of biostratigraphic events in both calcareous and siliceous plankton.

In terms of planktonic foraminifera, the Olduvai subchron is bracketed by the LADs of *Globorotalia miocenica* and *Globorotalia exilis* (tropical Atlantic region only) in the late Pliocene, the FAD of *Globorotalia truncatulinoides* and the LAD of *Globorotalia limbata* just below the base of the Olduvai, the LAD of *Globigerinoides obliquus extremus* near the top of the Olduvai, and the LAD of *Globigerinoides fistulosus* coincident with or a short distance above the Olduvai (Figure 7.2). In North Atlantic sediments, the Pliocene–Pleistocene boundary, as recognized here, would appear to be within Ericson's Zone Q, which is characterized by the absence or low values of the *Globorotalia menardii–G. tumida* complex (Figure 7.3).

Coiling patterns in the genus *Pulleniatina* have been found useful in determining the biostratigraphy of Plio–Pleistocene deep-sea sequences (Saito, 1976). A continuous sequence can be followed from the late Miocene to the present day in the equatorial Pacific, whereas in the equatorial Atlantic the genus *Pulleniatina* was absent from about 3.8 Ma to 2.6 Ma, owing, no doubt, to the closure of the Panama seaway in mid-Pliocene time. In late Pliocene time there was general synchrony in coiling patterns between the Atlantic and Pacific forms (Figure 7.4), which is useful in interoceanic correlation. The Olduvai subchronozone (a time-rock unit, as distinct from the time terms "chron" and "subchron") is essentially bracketed by two peaks of sinistral coiling (L5 at the base and L4 at the top in the Pacific, and AL2 at the base and AL1 at the top in the Atlantic). The Pliocene–Pleistocene boundary, recognized here at the top of the Olduvai subchron, corresponds to a pronounced peak in dextral coiling (L4 and L3) in the equatorial Pacific and to the beginning

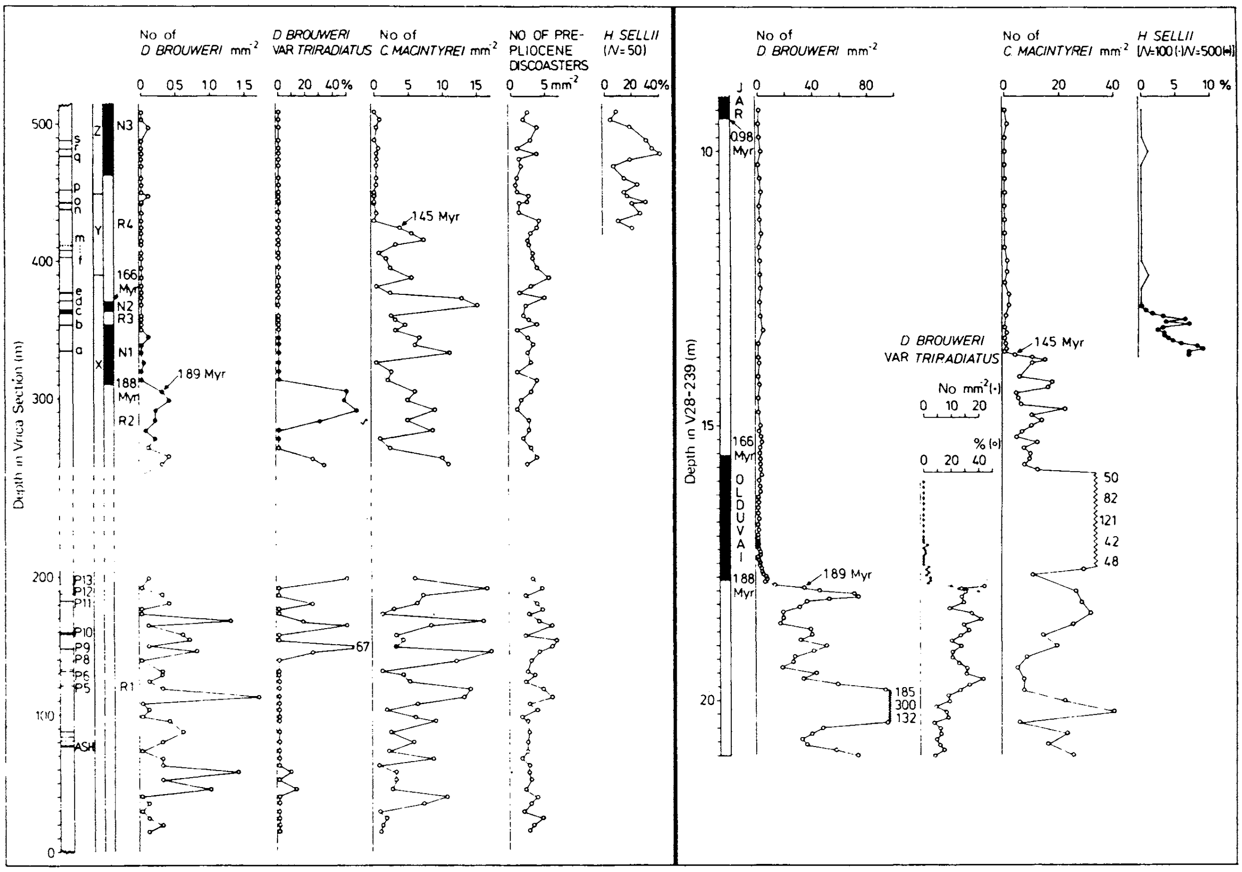


Figure 7.1. Quantitative nannofossil stratigraphy for the Vrica section and piston core V28–239 (western equatorial Pacific). On the left are data from the Vrica section, starting with the lithologic section of Selli et al. (1977) and the magnetostratigraphy of Tauxe et al. (1983). The abundances of *Discoaster brouweri*, *D. brouweri* var. *triradiatus* relative to all forms of *D. brouweri*, and *Calcidiscus macintyrei* each show a clear upper limit. Pre-Pliocene discoasters maintain a uniform reworked abundance throughout the section, and the proportion of *Helicosphaera sellii* relative to all helicosphaerids also does not show any drop in abundance in the upper part of the section. The filled circles between 270 and 345 m indicate samples that were prepared and counted in duplicate. On the right side, the sequence in piston core V28–239 is from Backman and Shackleton (1983), with magnetostratigraphy from Shackleton and Opdyke (1976). (From Backman et al., 1983, with permission of the editors of *Nature*.)

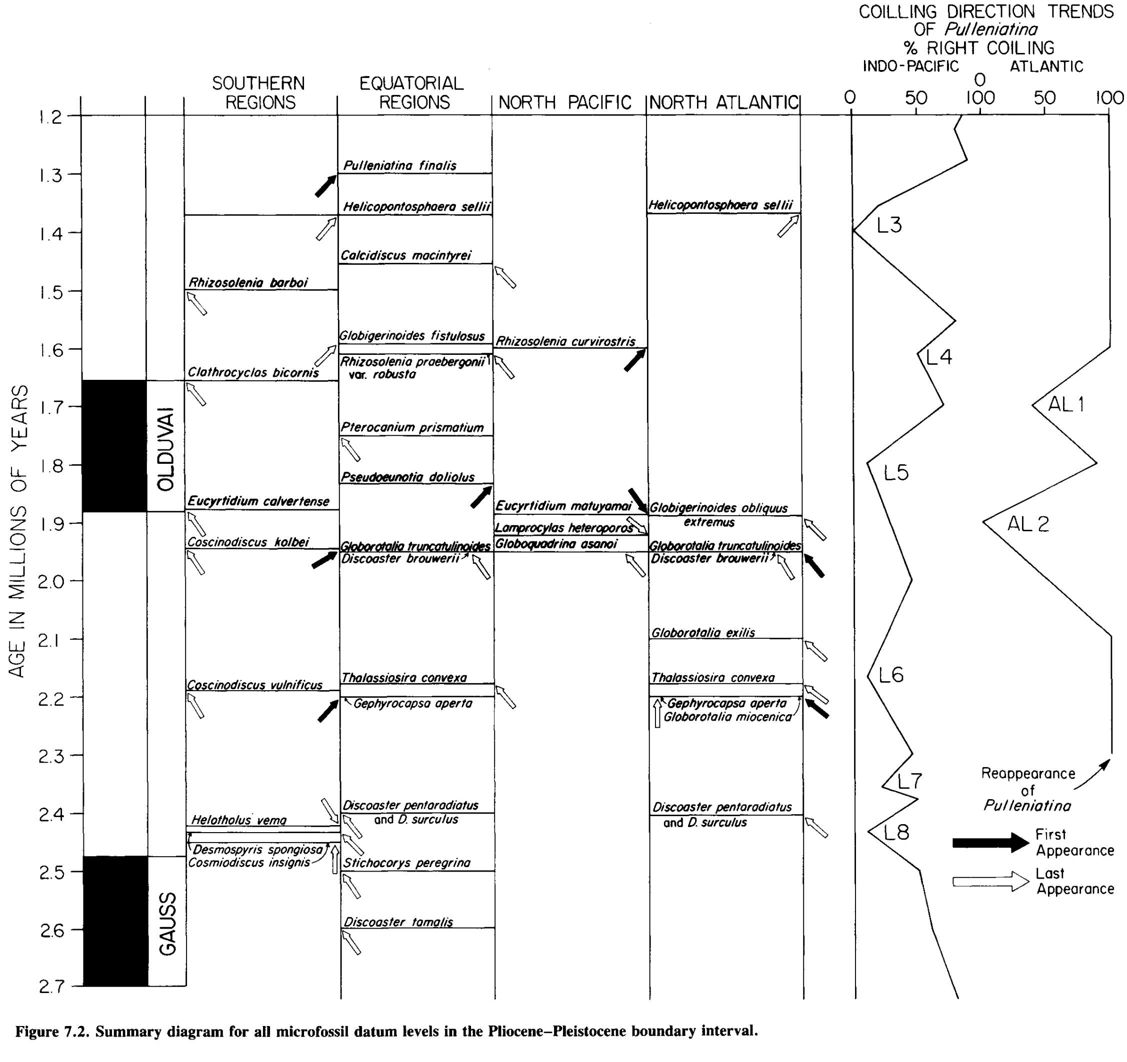


Figure 7.2. Summary diagram for all microfossil datum levels in the Pliocene–Pleistocene boundary interval.

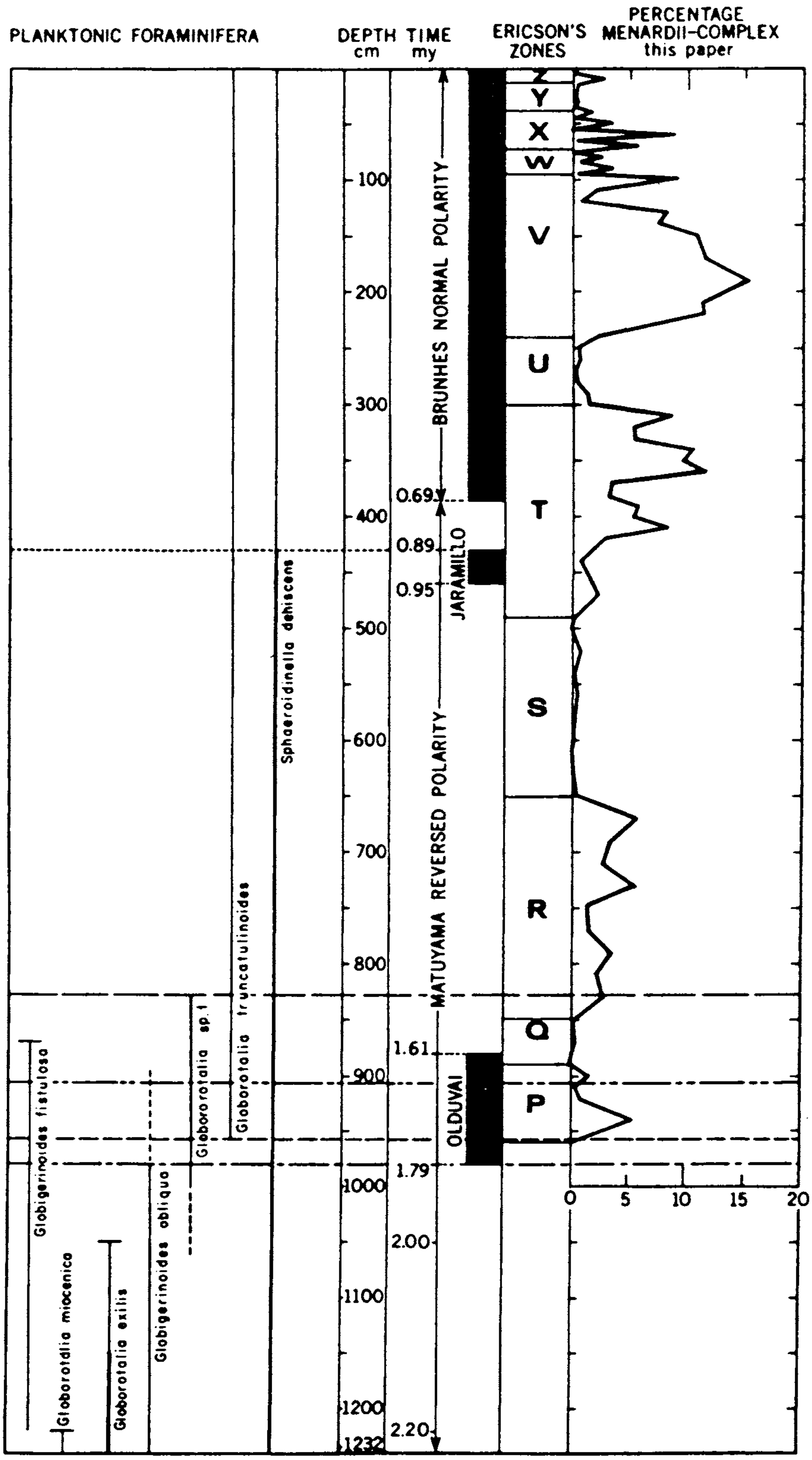


Figure 7.3. Magnetic and faunal stratigraphy. The solid curve is a plot of the percentage of *Globorotalia menardii* complex versus depth in centimeters. Ericson's zones are defined by the presence or absence of *G. menardii–G. tumida* species and subspecies. The paleomagnetic normal-polarity events are represented by solid black, and the reversed-polarity events by white. The appearances and extinctions of planktonic foraminifera species are represented by the labeled vertical lines. The horizontal dashed lines emphasize key biostratigraphic horizons. Between 800 cm and 1,000 cm, the dashed lines are the key faunal boundaries that define the Pliocene–Pleistocene boundary. The line through the top of the P-zone marks the *Discoaster* extinction boundary. The P-zone, as originally defined (Briskin and Berggren, 1975, fig. 1) by the first abundant *G. menardii–tumida* peak in the Olduvai subchron, is actually of latest Pliocene age in terms of the chronology presented here. (From Briskin and Berggren, 1975, courtesy of Micropaleontology Press.)

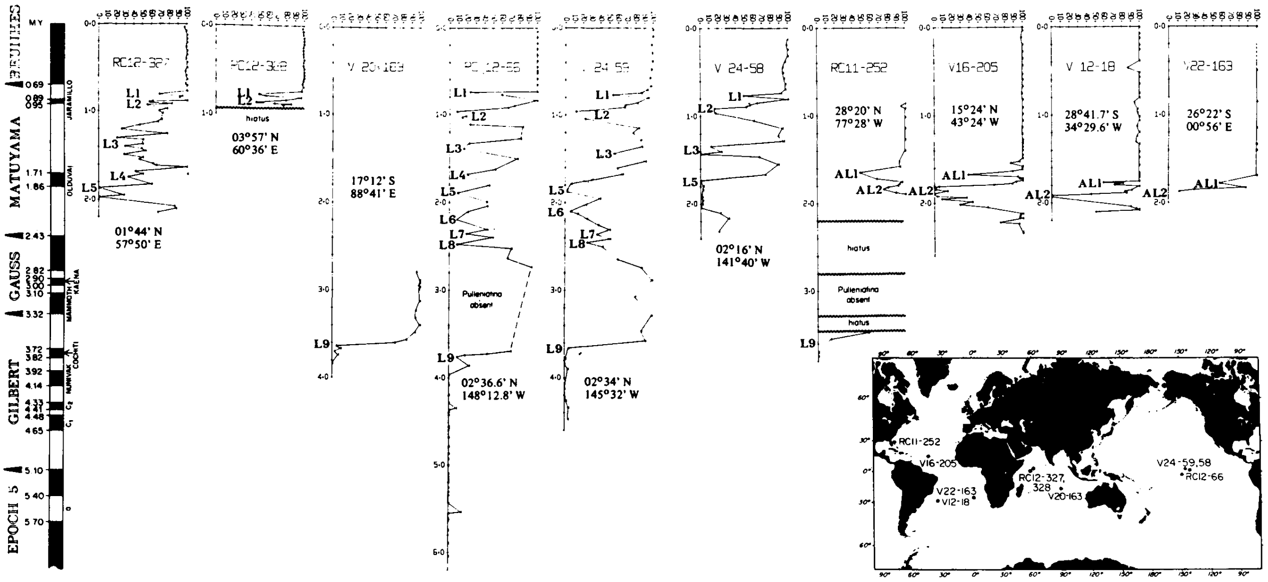


Figure 7.4. Time (m.y.) versus percentage of right-coiling individuals in *Pulleniatina* populations from 10 low-latitude deep-sea cores. Time lines are drawn according to the paleomagnetic-reversal sequence established for each core. Prominent left-coiling intervals are numbered L1–L9, AL1, and AL2. The upper Gilbert and Gauss intervals in core RC12–66 are low in carbonate because of dissolution effects. (From Saito, 1976, fig. 2, courtesy of the Geological Society of America.)

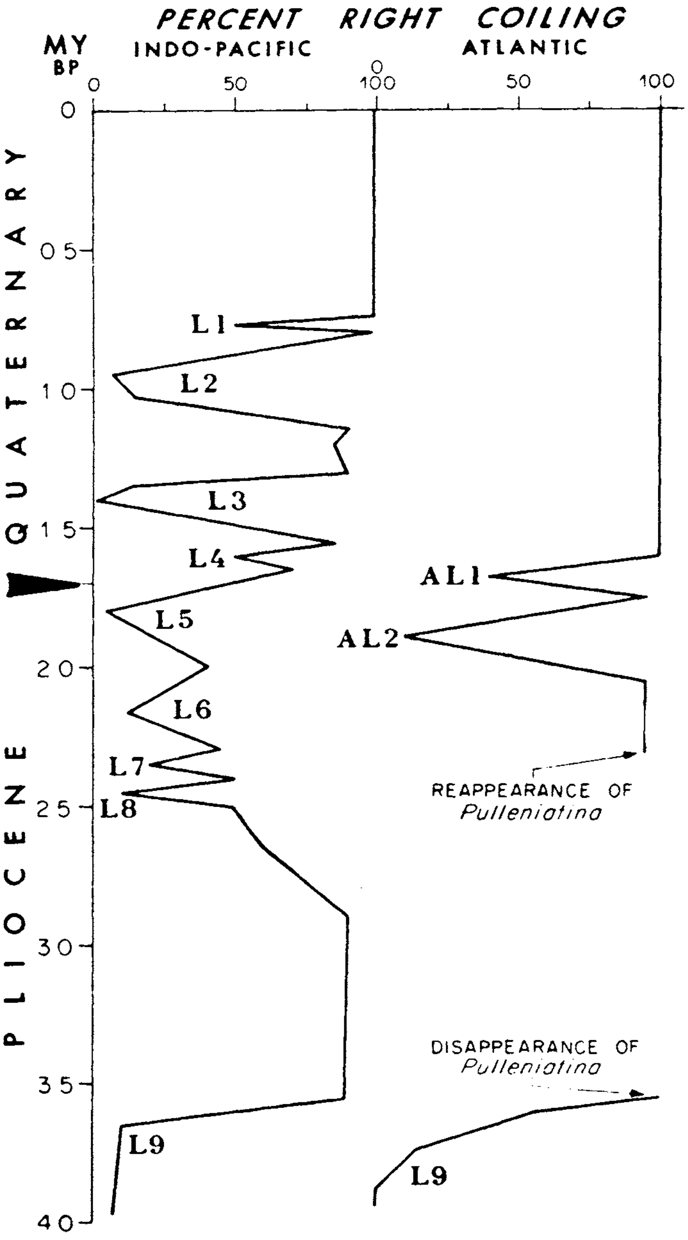


Figure 7.5. Generalized trends of coiling direction for *Pulleniatina* populations since 4 Ma in the Indo-Pacific and Atlantic faunal provinces. (From Saito, 1976, fig. 3, courtesy of the Geological Society of America.)

of what was to remain a dextral coiling pattern in the equatorial Atlantic (Figure 7.5).

Quantitative studies on the calcareous nannoplankton (Backman and Shackleton, 1983) have refined the application of the datum levels in this group to biostratigraphic problems. The extinction of the *Discoaster brouweri* group, just slightly below the Olduvai subchronozone, has essentially verified and given greater precision to previous studies (Haq et al., 1977) that had suggested an association of these two events. The LAD of *Calcidiscus macintyrei* is also seen to be stratigraphically above the Pliocene–Pleistocene boundary in the Vrica section as well as in deep-sea cores (Backman, Shackleton, and Tauxe, 1983, figure 1; Shackleton et al., 1995).

Quantitative data on the LADs for several late Pliocene calcareous nannoplankton have been delineated by Backman and Shackleton (1983) and are shown here (Figures 7.6 and 7.7) by way of reinforcing this approach as a powerful tool in magnetobiochronologic research.

In diatom-bearing sediments of the deep sea, the Olduvai subchronozone has been identified in the Pacific, equatorial Atlantic, and Southern oceans. In the equatorial Pacific, the Olduvai subchronozone is characterized by a number of first and last appearances of diatoms (Burckle, 1972, 1977). The LAD of *Thalassiosira convexa* occurs in the lower Matuyama, about one-third of the way from the top of the Gauss to the base of the Olduvai, whereas the first specimens of *Pseudoeunotia doliolus* in most deep-sea cores are usually found in the lower part of the Olduvai subchronozone. The actual FAD for this species appears to be just below the Olduvai (Shackleton et al., 1995), but it tends to be very rare in the lowest part of its range. The LAD of *Rhizosolenia praebergonii* var. *praebergonii* is just above the base of the Olduvai, while very slightly above its top is the LAD of *R. praebergonii* var. *robusta*; for a summary of these data, see Figures 7.2 and 7.8. In the equatorial Atlantic, Plio–Pleistocene diatom data from DSDP site 397 were reported by Burckle (1979), who found the LAD of *Thalassiosira convexa* just above the Gauss chronozone, but found no other datum levels nearer to the boundary (Figure 7.9).

Few quantitative data of that age are available for diatoms across this boundary in the equatorial Pacific. Burckle (1971) noted some evidence for a slight cooling trend throughout the Olduvai, a point that was also made by McIntyre, Bé, and Pretiskas (1967). Burckle and Trainer (1979) concluded that climatic fluctuations played an important role in the equatorial Pacific in dictating diatom first and last appearances during the late Pliocene–early Pleistocene.

Paleomagnetically dated deep-sea cores of that age are not available from high northern latitudes in the Pacific, but rich assemblages of boreal and cold-temperate marine diatoms occur in the Olkhov Formation of eastern Kamchatka, dated to Matuyama age. According to Gladenkov (1994), the Olkhov beds represent subarctic shelf environments, with winter ice cover, in which the level of the Vrica boundary in the upper Olduvai subchronozone is associated with the LADs for *Thalassiosira antiqua* and *Neodenticula koizumii*, above the FAD of *Pyxidicula californica*, and just below the FADs for *Simonseniella curvirostris* and *Thalassiosira pacifica*. This relationship correlates to the uppermost part of the *Neodenticula koizumii* zone, just below the boundary with the zone of *Actinocyclus oculatus*, as defined in deep-sea cores from the northwestern Pacific and the Sea of Japan (Koizumi, 1985, 1992). In some parts of the northwestern Pacific, the last occurrence of *Rhizosolenia praebergonii* var. *robusta* is available to help identify the immediate post-Olduvai levels, but this datum tends to be diachronous in the higher latitudes.

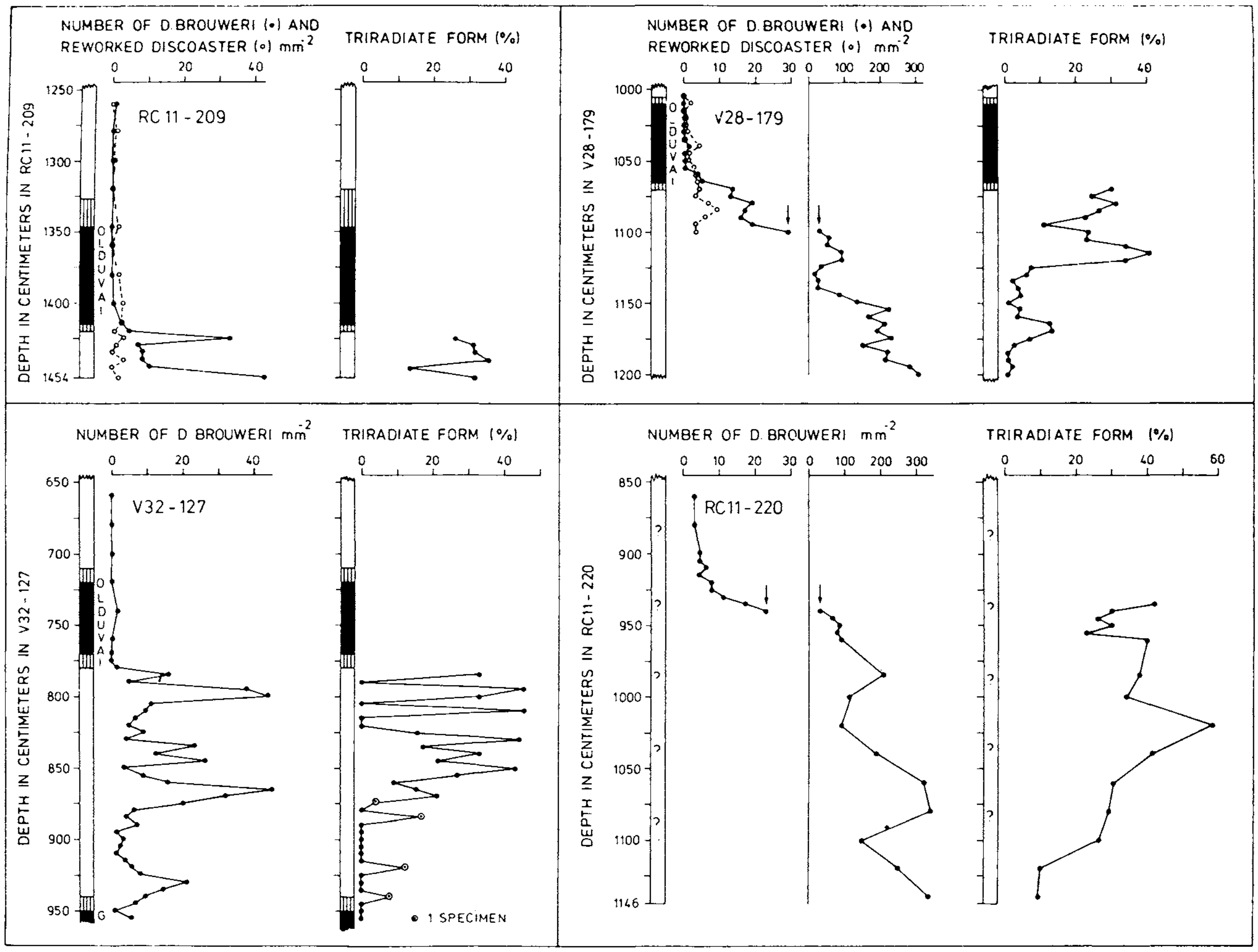


Figure 7.6. ***D. brouweri*** **abundances in four mid- and low-latitude Pacific cores. For each core, the abundance of *D. brouweri* is shown on the left, and the relative abundance of triradiate forms on the right. Open circles indicate the abundance of reworked Pliocene discoasters. (From Backman and Shackleton, 1983, fig. 4, with permission of Elsevier Publishing Co.)**

In the Southern Ocean, the Olduvai subchronozone has been identified in a number of piston cores and DSDP sites, in which it is characterized by the LAD of *Cosmiodiscus insignis* just above the top of the Gauss chronozone, the LAD of *Coscinodiscus vulnificus* midway between the top of the Gauss chronozone and the base of the Olduvai subchronozone, the LAD of *Coscinodiscus kolbei* just below the Olduvai, and the LAD of *Rhizosolenia barboi* just above the Olduvai (Donahue, 1970; McCollum, 1975; Cieselski, 1983). One of the major problems in detailing diatom datum events through the Plio–Pleistocene interval in the Southern Ocean is the uncertainty concerning the effects of the different water masses on these datum events. Most of the same biostratigraphic criteria apply to the sub-Antarctic region, but we know less about the region south of the polar front.

Our knowledge of radiolarian datum levels in the Pliocene–Pleistocene boundary interval of paleomagnetically dated cores is largely derived from the classical studies of Hays and coworkers (Hays, 1965, 1970; Opdyke et al., 1966; Hays and Opdyke, 1967; Hays et al., 1969; Saito, Burckle, and Hays, 1975). The most significant datum in the equatorial Pacific is the widely recorded extinction of *Pterocanium prismatium* near the Pliocene–Pleistocene boundary. Hays et al. (1969) and Saito et al. (1975) reported that it was absent from the upper part of the Olduvai subchron (Figure 7.2), while Shackleton et al. (1995) placed this datum just above its top. The latter authors also noted the FAD of *Anthocyrtidium angulare* coincident with the LAD of *Globigerinoides fistulosus* at the upper boundary of the Olduvai subchronozone, and the LAD of *Anthocyrtidium jenghisi* coincident with the FAD of *Globorotalia truncatulinoides* at about 0.5 m.y. earlier than its base. The other significant radiolarian biostratigraphic events in the vicinity of the boundary are the last appearance of *Stichcorys peregrina* in the upper part of the Gauss chron and the successive last appearances of *Lamprocyrtis heteroporos, Pterocorys minithorax*, and *Theocorythium vetulum* in the interval between the Olduvai and Jaramillo subchrons, the last through evolutionary

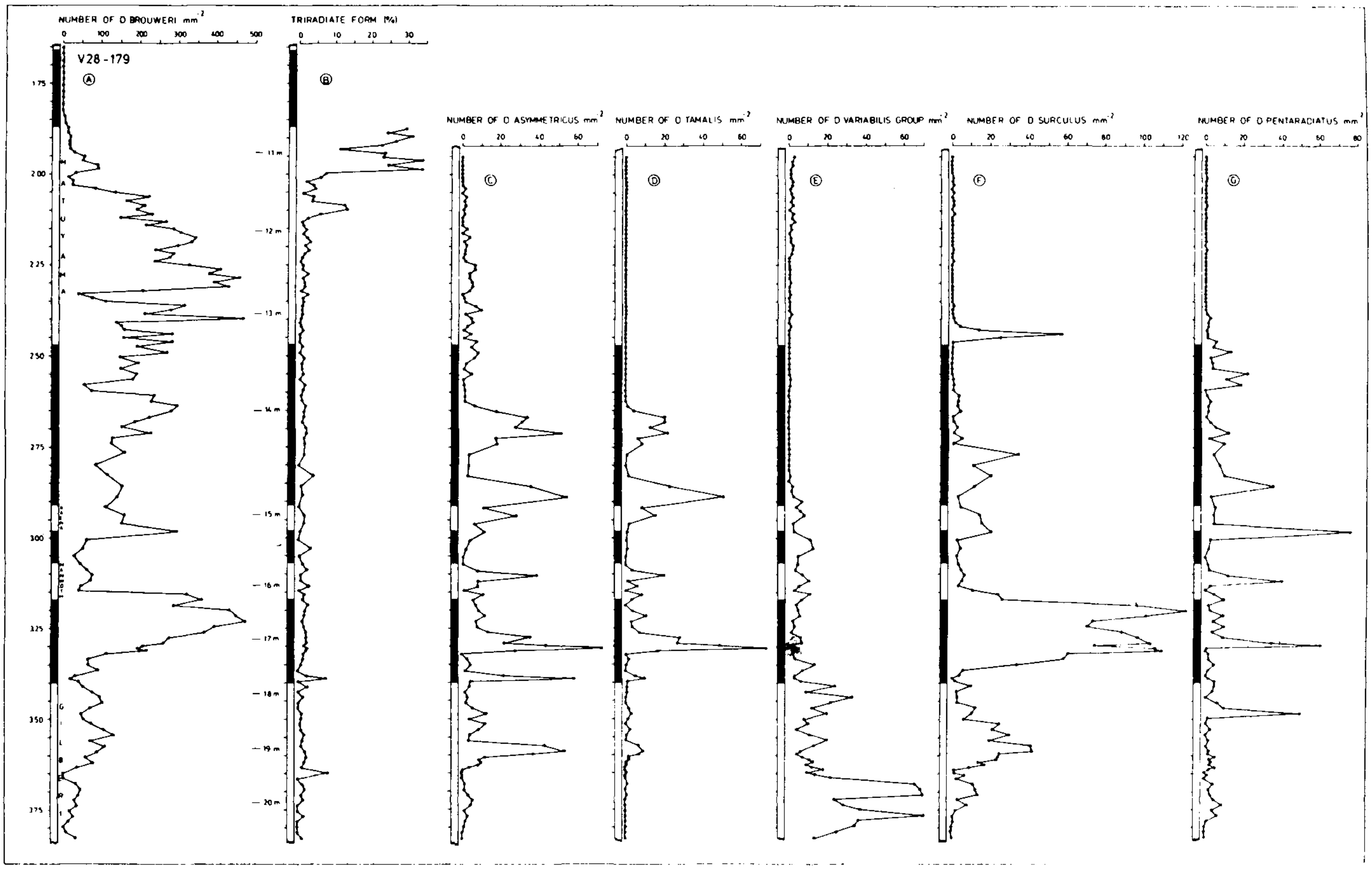


Figure 7.7. Abundances of *Discoaster* species in V28–179, central equatorial Pacific, in the interval from the top of the Olduvai to the base of the Gauss (see Preface, this volume, for modern time scale values): A, abundance of *D. brouweri*; B, abundance of *D. triradiatus*; C, abundance of *D. tamalis*; D, relative abundance of *D. tamalis*; E, abundance of the *D. variabilis* group (including *D. challengeri* and *D. decorus*); F, abundance of *D. surculus*; G, abundance of *D. pentaradiatus*. The sampling interval is 5 cm, but the data are presented on a time-scale based on paleomagnetic data. Note scale difference between *D. brouweri* and the other discoasters. (From Backman and Shackleton, 1983, fig. 6, with permission of Elsevier Publishing Co.)

replacement by *Theocorythium trachelium* (Nigrini, 1970; Johnson and Knoll, 1975; Shackleton et al., 1995).

In the northern Pacific, three radiolarian datum levels are recognized in the Pliocene–Pleistocene boundary interval (Hays, 1970) (Figure 7.2). The first, the LAD of *Eucyrtidium elongatum peregrinum*, occurs in the middle of the Gauss chronozone, as it does in the central Pacific. The second is the LAD of *Lamprocyrtis heteroporos* near the base of the Olduvai subchronozone, although, as discussed earlier, the southern population of this species in the equatorial Pacific became extinct at a later date, as was also the case in the high latitudes of the Southern Ocean (Hays, 1970). Third, *Eucyrtidium matuyamai* evolved from *E. calvertense* near the base of the Olduvai subchron. Hays (1970) notes that in the Southern Ocean, *E. calvertense* disappears in the lower part of the Olduvai subchron at approximately the same level where it gives rise to *E. matuyamai* in the northern Pacific.

In the Southern Ocean, Hays (1965), Opdyke et al. (1966), and Hays and Opdyke (1967) reported a number of radiolarian events near the Pliocene–Pleistocene boundary interval. *Desmospyris spongiosa* and *Helotholus vema* disappear just above the Gauss chronozone, and *Clathrocyclas bicornis* disappears near the top of the Olduvai subchronozone. In terms of assemblage zones, the Pliocene–Pleistocene boundary would fall within the X Zone of Hays (1965). Hays and Opdyke (1967) also reported a sharp decrease in the abundances of the more widely ranging, presumably less cold-tolerant radiolarians in the Southern Ocean just above the Olduvai.

Stable-isotope record

It is well known that in the Southern Hemisphere, Antarctica has had a continental-ice record extending as far back as the middle Miocene, probably into the late Oligocene, and, according to some interpretations of the stable-isotope record, as far back as the Eocene–Oligocene transition. In contrast, the biostratigraphic studies of North Atlantic deep-sea cores suggest that the initiation of ice rafting and thus Northern Hemisphere glaciation

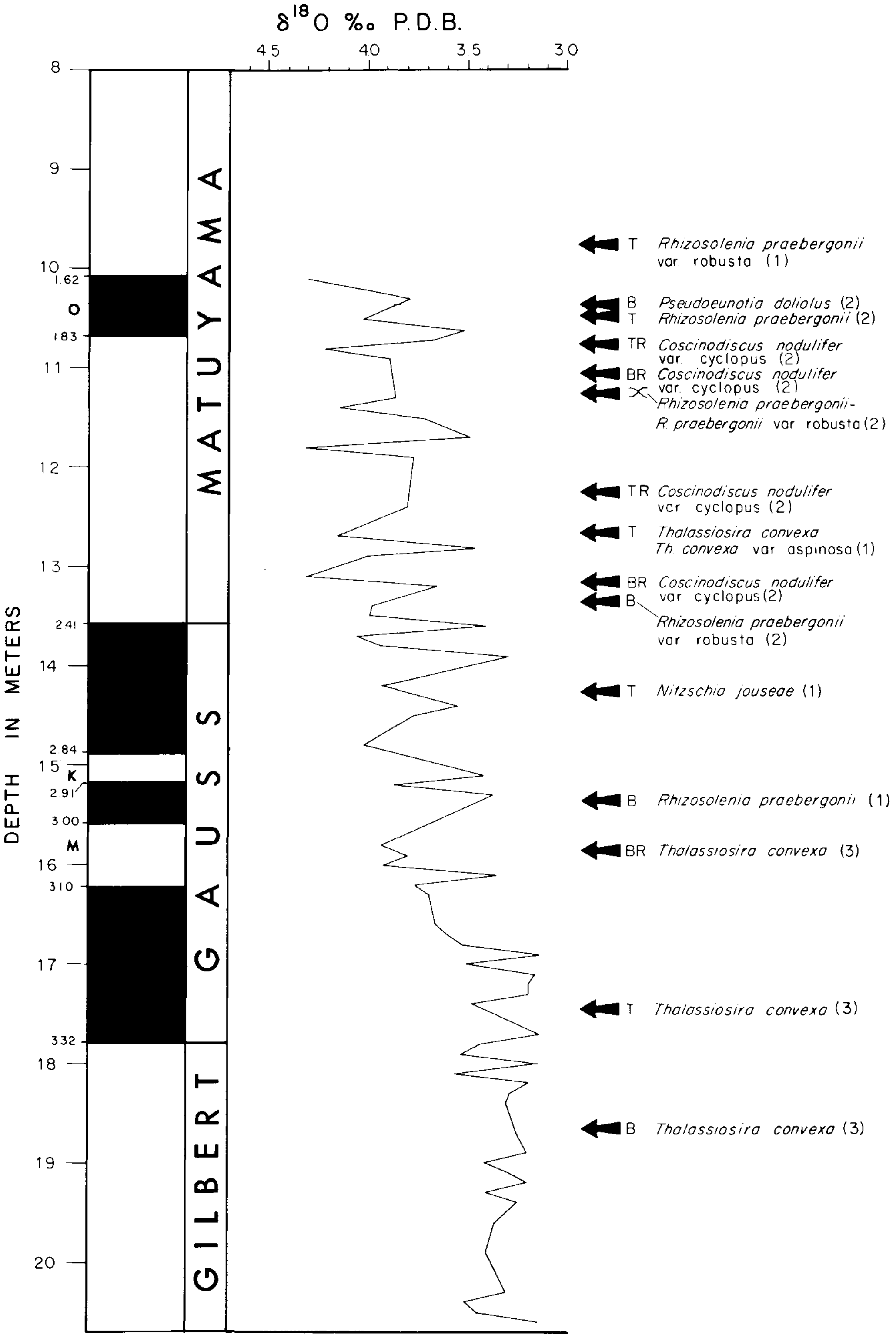


Figure 7.8. Magnetostratigraphy and diatom datum levels for Upper Pliocene–Lower Pleistocene sediments of the equatorial Pacific. (From Burckle and Trainer, 1979, courtesy of Micropaleontology Press.)

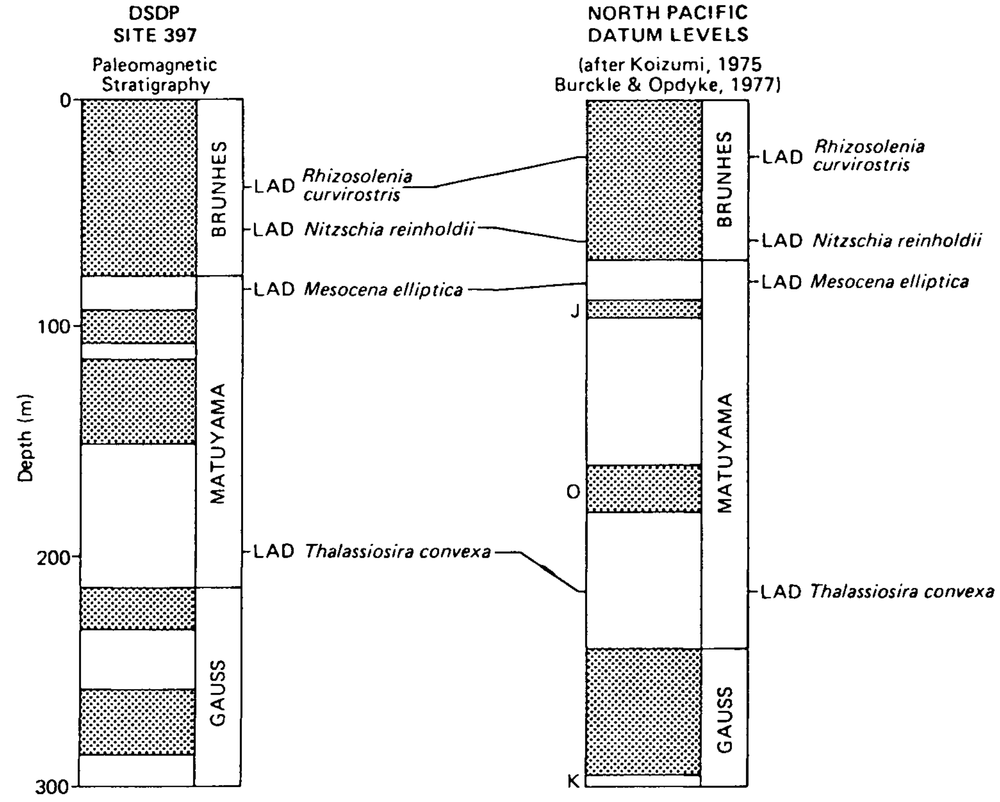


Figure 7.9. Magnetostratigraphy and diatom datum levels for Upper Pliocene–Lower Pleistocene sediments for DSDP site 397 in the equatorial Atlantic. (From Burckle, 1979, courtesy of the Offshore Drilling Program.)

occurred at about 3 Ma (Berggren, 1972). That remained the orthodox opinion for nearly a decade. A biostratigraphic restudy of North Atlantic core material (Backman, 1979) has more recently suggested that the initiation of glaciation may have occurred closer to 2.5 Ma than 3.0 Ma, corroborated by oxygen-isotope records in the equatorial Pacific (Shackleton and Opdyke, 1977; Shackleton et al., 1995) and the North Atlantic (Shackleton et al., 1984, 1990).

It is now apparent that although there was marked climatic variability on a global scale in the early–middle Pliocene, global climate cycles did not induce sea-level glaciation until about 2.5 Ma. That conspicuous climatic threshold corresponds temporally to the Pretiglian cool-climate interval in Europe, which saw significant vegetational changes just after the Gauss/Matuyama polarity reversal, and that change has been interpreted by some (e.g., Zagwijn, 1974) as indicative of the Pliocene–Pleistocene boundary. In fact, the Pretiglian cold-climate peak seems to have been a stepping-stone midway on the descent from preglacial conditions in the mid-Pliocene (3.0 Ma) to fully glacial conditions in the mid-Pleistocene (0.9 Ma). As discussed elsewhere (Hays and Berggren, 1971; Berggren and Van Couvering, 1979; Aguirre and Pasini, 1985; Shackleton et al., 1990), that climatic event was not of the same age as the Lyellian boundary recognized by the International Geological Congress in 1948 (King and Oakley, 1950) and formally adopted at Vrica (Selli, 1977; Cowie and Bassett, 1989).

The main features of the late Pliocene–Pleistocene oxygen-isotope record include:

1. minor fluctuations prior to 2.4 Ma
2. a relatively major (approximately 1%) increase in ^{18}O of ocean waters at about 2.5 Ma, corresponding to a major expansion of glacial (year-round) ice in the Northern Hemisphere and an expansion to the limits of drifting ice in the northern oceans
3. stronger cycles since that time, with fluctuations in oxygen-isotope ratios on the order of 1% and periodicities that, at least over the past million years, have been correlated to periodic variabilities in the rotation and orbital path of the earth (Milankovitch cycles) (Shackleton et al., 1990, 1995).

An inspection of late Pliocene–Pleistocene oxygen-isotope records (Figures 7.10 and 7.11) shows two climatic regimes characterized by average glacial–interglacial variations in oxygen-isotope ratios, the first (between 2.5 and 0.9 Ma) varying by about 1%, and the second (0.9 Ma to present day) by more than 1%. The change to more strongly contrasting values in the middle and late Pleistocene (i.e., after the Jaramillo subchron) indicates a change to more severe glacial conditions during climatic lows. In terms of the Pliocene–Pleistocene boundary, oxygen-isotope variations during the approximately 220-k.y.-

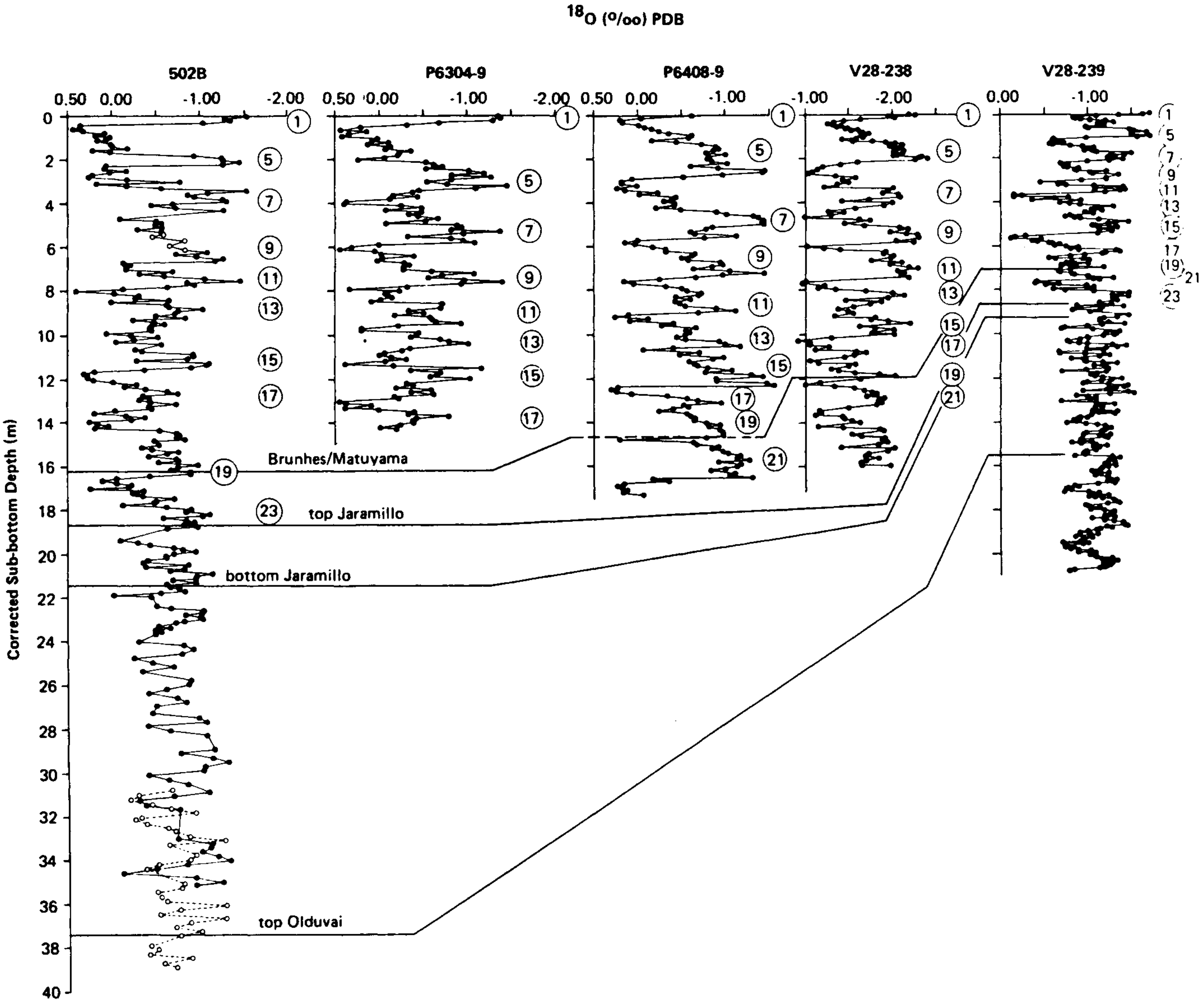


Figure 7.10. Correlation of oxygen-isotope stratigraphy between the Atlantic (DSDP site 502B, and piston cores P6304–9 and P6408–9) and the Pacific (cores V28–238 and V28–239). Interglacial isotope stage numbers are circled. Where present, magnetostratigraphic boundaries are shown. Cores P6304–9 and P6408–9 are from Emiliani (1966, 1978), and cores V28–238 and V28–239 are from Shackleton and Opdyke (1976, 1977). Site 502B is from Prell (1982, fig. 2).

long Olduvai subchron correspond to approximately three stadial–interstadial cycles. According to previous interpretations of the Vrica section (Aguirre and Pasini, 1985), the Pliocene–Pleistocene boundary placed just above the top of the Olduvai interval would fall within an interstadial period (Shackleton et al., 1984, 1995) (Figure 7.12), equivalent to oxygen-isotope stage 63 (Shackleton et al., 1990, figure 4). If the boundary is placed within the uppermost part of the Olduvai subchronozone, however, according to revised magnetostratigraphy at Vrica (Zijderveld et al., 1991), it will fall instead in the cold-climate maximum of oxygen-isotope stage 64 (Shackleton et al., 1990, figure 4).

Carbonate stratigraphy

Cyclic fluctuations in the carbonate contents of deep-sea sediment cores have long been known to correlate with, and reflect, climatic oscillations on a global scale. In general, high carbonate values correspond to glacial intervals, and low carbonate values correspond to relatively warm interglacial intervals. When correlated by means of biostratigraphy and calibrated to a geochronologic system by means of paleomagnetic stratigraphy, carbonate stratigraphy has become a powerful tool for late Neogene global correlations, as summarized by Vincent (1981).

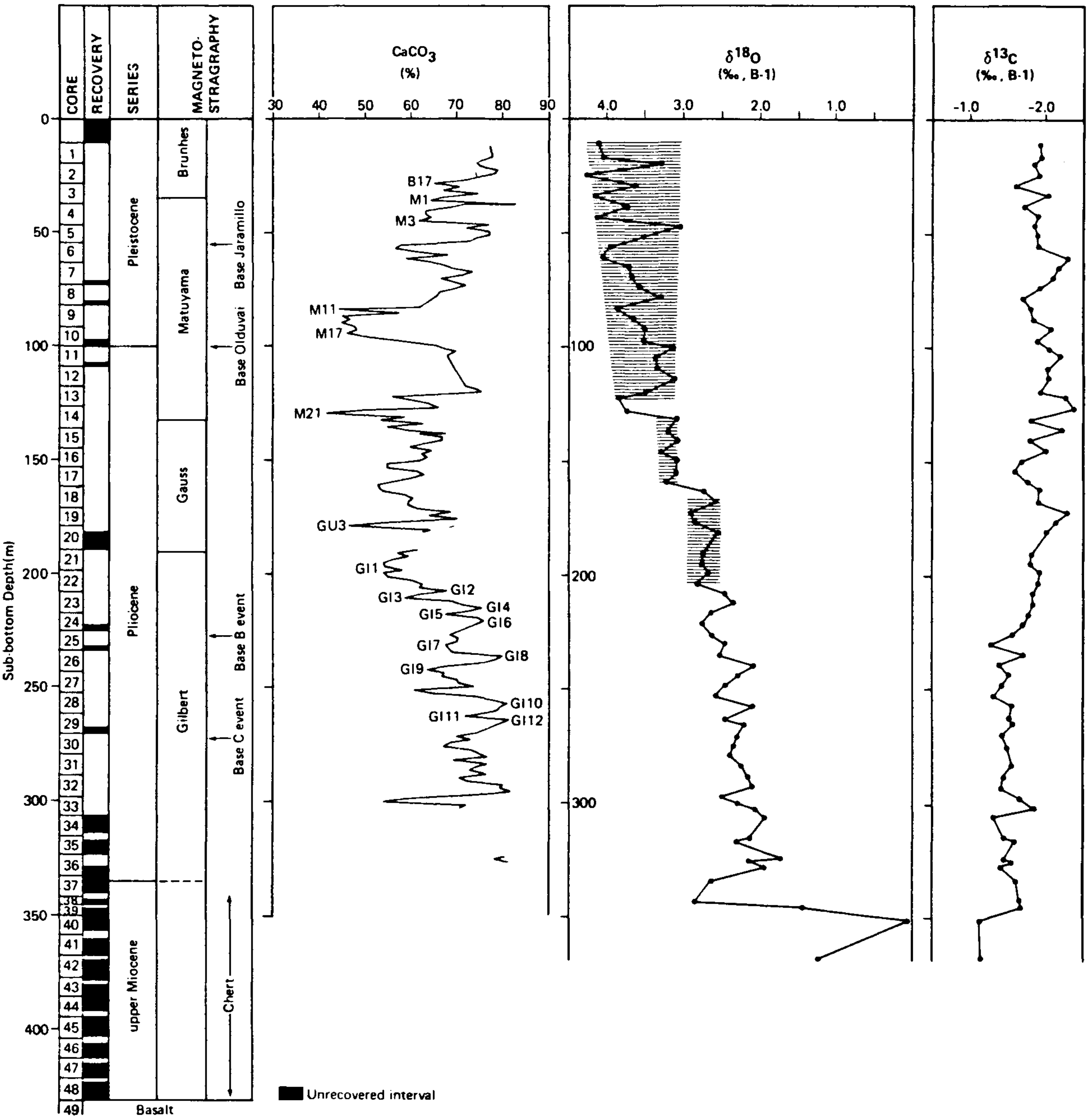
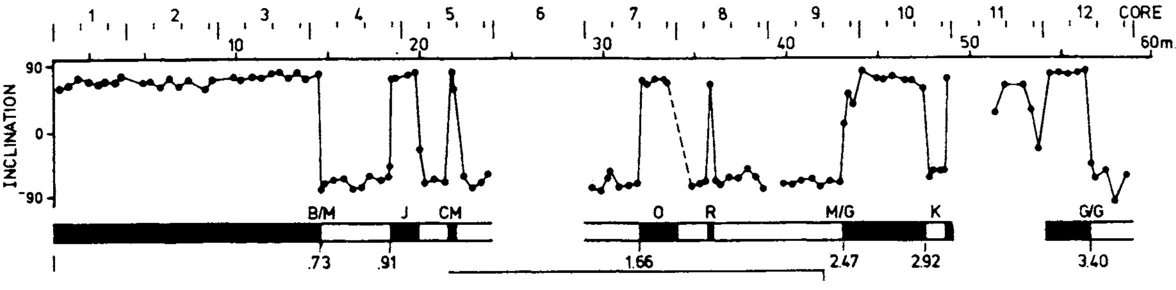


Figure 7.11. Magnetostratigraphy, carbonate stratigraphy, and isotope stratigraphy at DSDP site 157. Magnetostratigraphy derived from correlations with paleomagnetically dated cores (Kaneps, 1973). Carbonate data from Bode and Cronin (1973). Isotope data from Keigwin (1979) on *Uvigerina*. Note the coincidence, as at site 310, of low-carbonate event M21 with the enrichment in ^{18}O just above the Matuyama–Gauss boundary. (From Vincent, 1981, fig. 4, courtesy of the Offshore Drilling Program.)

A sequence of eight carbonate cycles (corresponding to glacial–interglacial episodes) was identified in the equatorial Pacific (Hays et al., 1969) throughout the Brunhes magnetic-polarity chronozone. Periodicities ranged from about 75,000 years in the late Brunhes to over 100,000 years in the early Brunhes. In the Matuyama chron (down to the top of the Olduvai subchron) seven more cycles were identified, with average, but more irregular, periodicities of 100,000 years. A numerical system was established, with odd and even numbers denoting low (interstadial warm climate) and high (stadial cold climate) carbonate intervals, respectively. Carbonate cycles were recognized down to the Gilbert C event in the early Pliocene. That sequence was subsequently extended down into the late Miocene by Saito et al. (1975). Kaneps (1973) extended the nomenclature system below the mid-Pliocene (GU3) to the early Gilbert and informally numbered 25 carbonate peaks and valleys



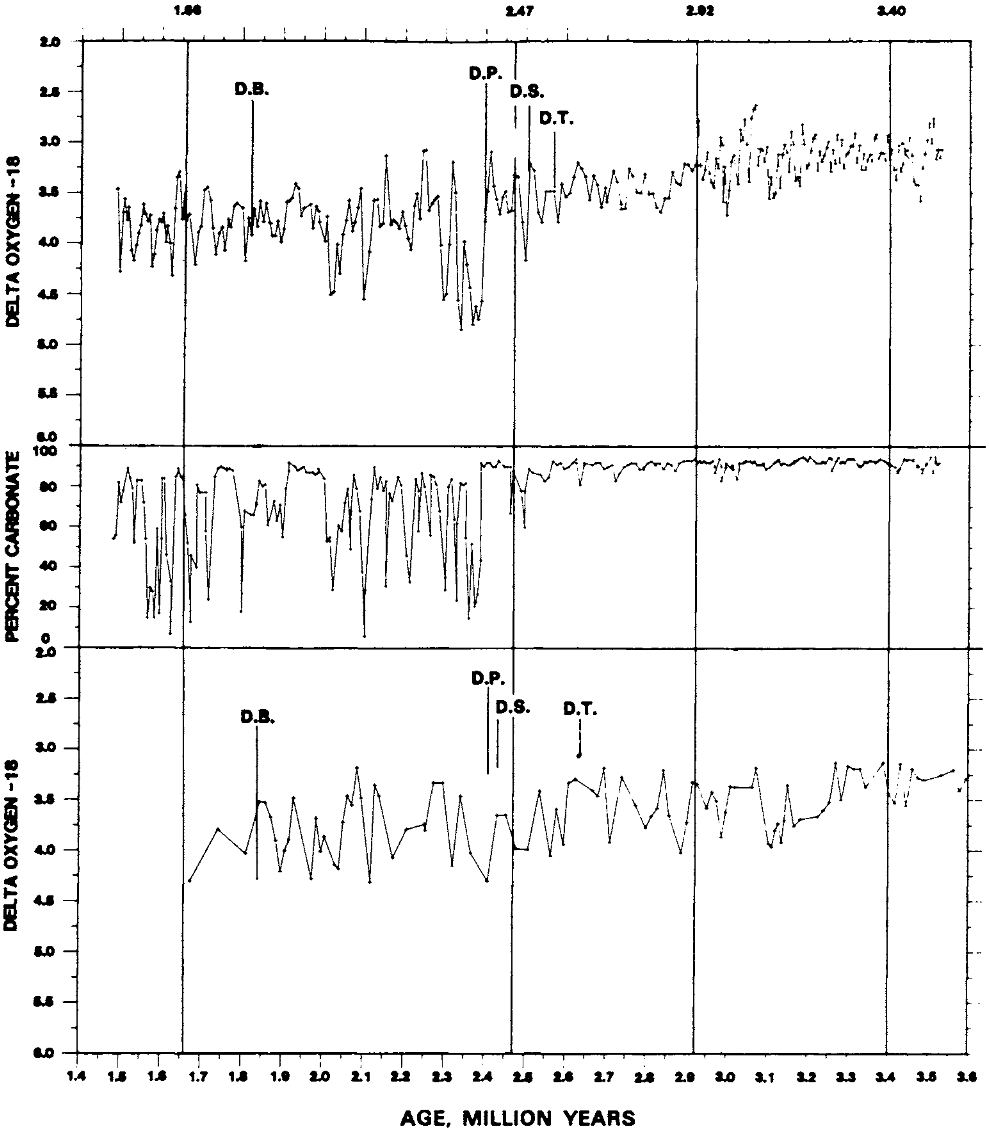


Figure 7.12. Combined magnetic and oxygen-isotope records for DSDP site 552A. Top: Magnetic record for site 552A. Demagnetized inclinations are shown only for apparently undisturbed parts of the cores (the data from core 11 suggest the presence of some unrecognized disturbance). Bottom: Upper panel shows the oxygen-isotope record of site 552A, cores 7–12. The plotting scale is linear between magnetic-reversal horizons. The lower panel shows the oxygen-isotope record in Pacific core V28–179 for comparison. Vertical lines show horizons used for time control as follows: top of the Olduvai subchron at 1.66 Ma, base of the Matuyama reversed chron at 2.4 Ma, top of the Kaena reversed subchron at 2.92 Ma, base of the Gauss normal chron at 3.40 Ma. Nannofossil extinction horizons determined on both cores are indicated as follows: DT, *Discoaster tamalis*; DS, *D. surculus*; DP, *D. pentaradiatus*; DB, *D. brouweri*. (From Shackleton et al., 1984, figs. 2 and 4, with permission of the editors of *Nature*, London.)

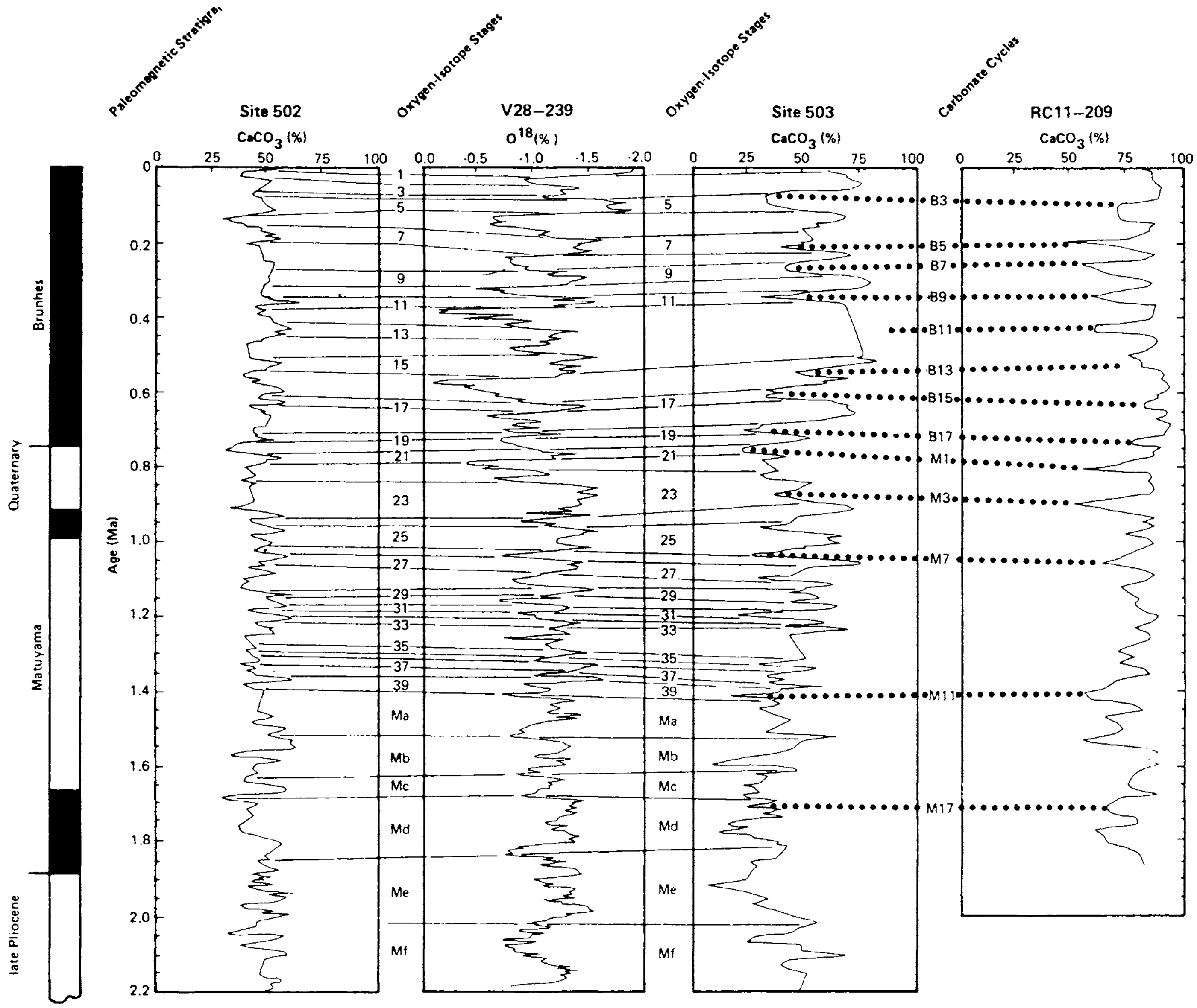
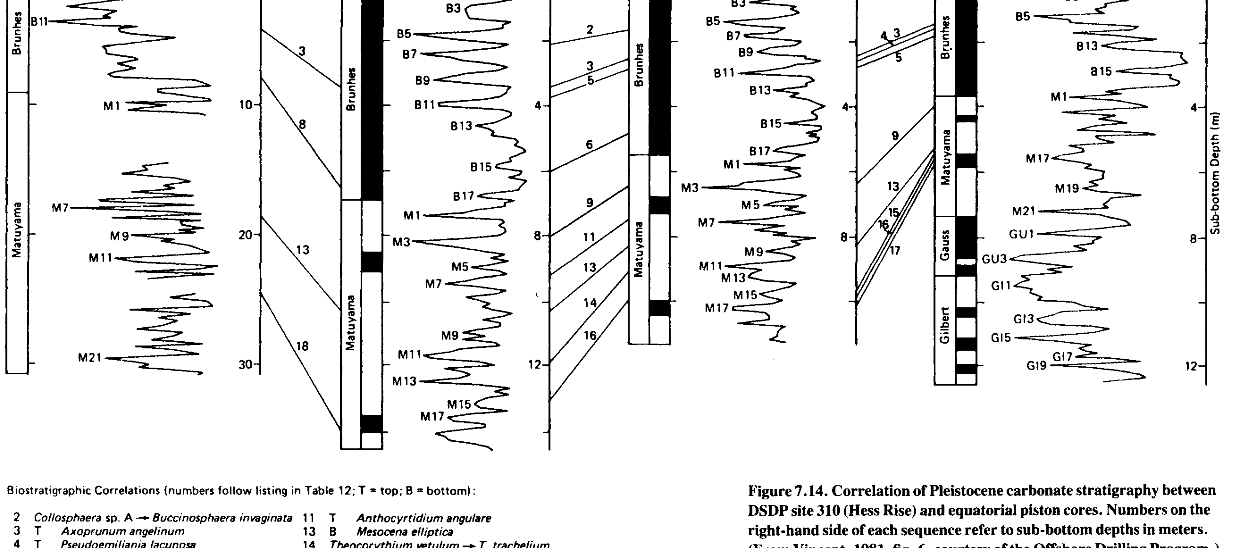


Figure 7.13. Correlations between the carbonate stratigraphy in site 502 and the oxygen-isotope stratigraphy in core V28–239, and between the carbonate stratigraphies in site 503 and core RC11–209 and the oxygen-isotope stratigraphy in core V28–239. The oxygen-isotope stages are modified from Shackleton and Opdyke (1976), and the carbonate cycles are from Hays et al. (1969). (From Gardner, 1982, fig. 10, courtesy of the Offshore Drilling Program.)

with lowercase-letter designations. That informal system was replaced by a formal nomenclature in which carbonate events were designated according to their association within a particular magnetic-polarity chron (Dunn and Moore, 1981). Dunn and Moore extended their carbonate stratigraphy down to magnetic chron 9 and gave alternating lowercase notations to carbonate minima (a, c, e, etc.) and maxima (b, d, f, etc.).

In the western Caribbean (DSDP site 502) and eastern equatorial Pacific (DSDP site 503), a high-resolution carbonate stratigraphy has been established for the late Pliocene and Pleistocene (Gardner, 1982) (Figure 7.13) in which carbonate variations are linked with the oxygen-isotope stages of Shackleton and Opdyke (1976), and their system of informal designation has been extended into the early Pleistocene. In this scheme, lowercase letters are used, each preceded by the uppercase abbreviation for the associated magnetic-polarity chron (i.e., Mb = Matuyama b). Similar high-resolution carbonate stratigraphies have been developed for the Pacific Ocean (Figure 7.14).

In terms of the Pliocene–Pleistocene boundary, it can be seen from the data summarized by Vincent (1981) and Gardner (1982) that the Olduvai subchronozone is bracketed by a carbonate low (M17) at the top and a carbonate peak (M18) at the base, in the terminology of Hays et al. (1969). The Pliocene–Pleistocene boundary, if located just above the top of the Olduvai subchron,



Biostratigraphic Correlations (numbers follow listing in Table 12; T = top; B = bottom):

2 *Collosphaera* sp. A → *Buccinosphaera invaginata*
3 T *Axoprunum angelinum*
4 T *Pseudoemiliania lacunosa*
5 B *Collosphaera tuberosa*
6 B *Collospaera* sp. A
8 T *Nitzschia reinholdii*
9 T *Mesocena eliptica*
11 T *Anthocyrtidium angulare*
13 B *Mesocena elliptica*
14 *Theocorythium vetulum* → *T. trachelium*
15 T *Rhizosolenia praebergonii*
16 T *Pterocanium prismatium*
18 B *Pseudoeunotia doliolus*

Figure 7.14. Correlation of Pleistocene carbonate stratigraphy between DSDP site 310 (Hess Rise) and equatorial piston cores. Numbers on the right-hand side of each sequence refer to sub-bottom depths in meters. (From Vincent, 1981, fig. 6, courtesy of the Offshore Drilling Program.)

would correspond to the carbonate minimum M15, but if located within the latest part of the Olduvai subchron it would correlate with the more pronounced minimum M17; both are within the interval of low carbonate values corresponding to oxygen stage Mb of Gardner (1982).

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8 Late Cenozoic changes of flora in extra-tropical Eurasia in the light of paleomagnetic stratigraphy

VLADIMIR P. GRICHUK

Introduction

Most studies of the problem of stratigraphic resolution in the transition zone between the Pliocene and the Pleistocene have been based on geology, mammals, malacofauna, or marine microfauna. Paleobotany has been used to a much lesser extent, and mainly for interpretation of the vegetational changes resulting from continental glaciations. Good examples of this approach are the studies of Suc and Zagwijn (1983) and Lona and Bertoldi (1973). Both in those studies and in the majority of others, evolution of flora has been considered on a quite limited scale and as a rule has concerned only evolutionary sequences in the development of individual species (mainly of water plants and bog plants). Such studies seldom have been concerned with overall changes in regional floras.

The paleobotanic data on the deposits of the late Cenozoic period show, quite reliably, that the flora of the Neogene system, which was of a subtropical character, was considerably richer than that of the Quaternary. The latter, in the whole extra-tropical space of the Northern Hemisphere, had the properties of a temperate-type flora. That gives us grounds to believe that studies of the processes by which floristic types change may yield quite important insights into stratigraphic problems at the border between the Neogene and the Quaternary.

Unfortunately, a detailed characterization of the successive paleofloras of the late Cenozoic period is possible, at present, in only a few well-studied regions. There is, however, one circumstance that can greatly help in obtaining the data necessary for such an investigation. During the late Cenozoic period, the vegetation cover over the greater part of extra-tropical Eurasia was clearly of forest character. In botanic studies, all of the main types of extra-tropical forests (boreal, nemoral, subtropical shrub–arboreous) usually are described according to tree species. According to data reported by Sokolov and Svyazeva (1965), however, the characteristic differences, in terms of both systematics and diversity, are already accounted for when dendroflora are classified at the genus level. Genera are also more certainly identifiable, and more widespread, than any of the species of which they are composed. The genus is therefore the most appropriate basis for a reliable and regionally applicable characterization of the late Cenozoic forest floras. Furthermore, studies of fossil dendroflora at the genus level can considerably broaden the usefulness of paleobotanic materials, in which palynology based on the determination of tree pollen at the genus level rather than the species level occupies the central place.

Paleofloral associations and climate change

In this review we shall consider data from a few regions within the extra-tropical part of Eurasia. These regions have been chosen according to the following two criteria: (1) adequate sample availability across the Pliocene–Pleistocene boundary interval and (2) the availability, in the given region, of paleomagnetic definitions, which make it possible to establish reliable correlations between the characterized horizons and the general paleomagnetic scale. The latter requirement, being absolutely necessary, has limited this study to the six regions described herein (Figure 8.1).

The genera of dendroflora in the Upper Cenozoic deposits of these regions have been compared with the present distributions of all the species in each genus. In this way it is possible to group the genera according to the modern areas in which they are most abundant and most diversified, which for the most part reflects their basic adaptive characteristics. The use of this system allows us to follow the process of change in the compositions of dendrofloras and to interpret the differentiation of floristic adaptations during the late Cenozoic period in terms of climatic change.

The geographic associations of genera (Grichuk, 1959) according to this method of organization are as follows:

1. Panholarctic
2. American–Eurasiatic
3. American–Mediterranean–Asiatic
4. American–Eastasiatic
5. Eastasiatic
6. North American (abbreviated as N.A. or A. in some tables)
7. Tropical, including subgroups of pluricontinental and South Asiatic–American (abbreviated as Tr. in some tables)

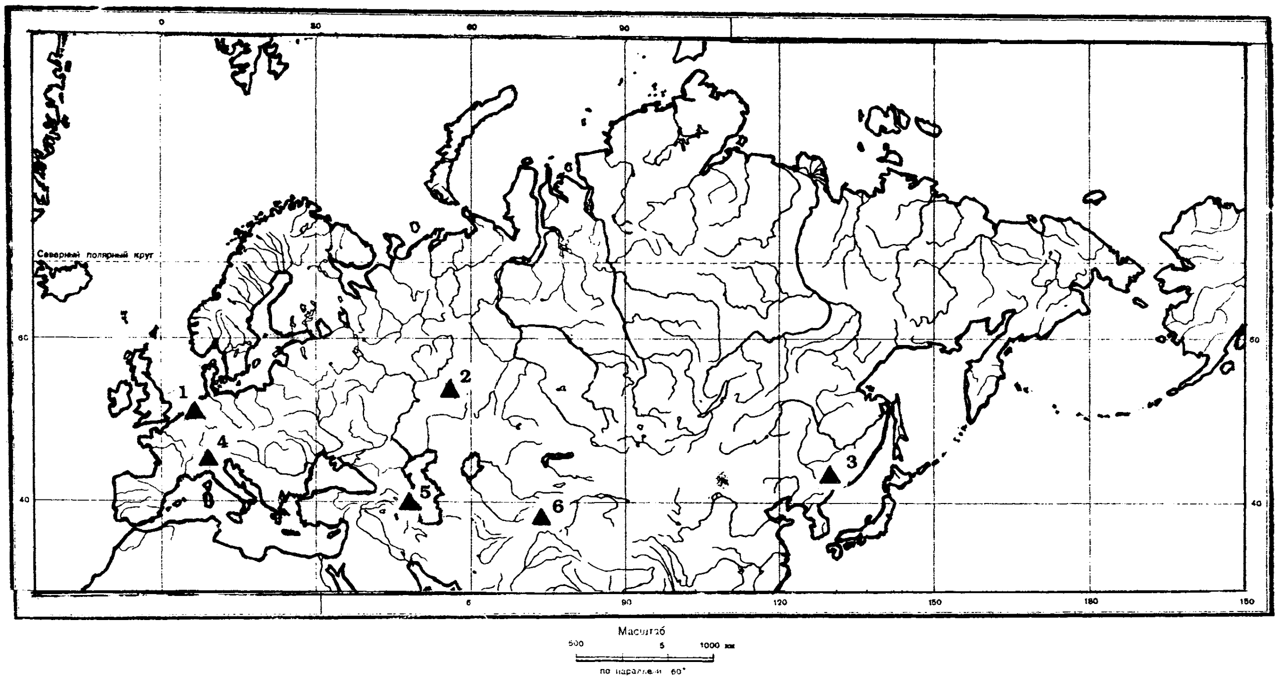

Figure 8.1. Map of regions for which the changes in the genus diversity of Upper Cenozoic dendroflora were analyzed (see Tables 8.1–8.6): 1, The Netherlands; 2, the Bashkirian foreland of the Urals; 3, Primoriye; 4, nortnern Italy; 5, the Kura lowland and the southeastern foothills of the High Caucasus; 6, the Pamirs.

The Netherlands

Table 8.1 shows data on changes in the regional dendroflora in the period from the middle Pliocene (Brunssumian) to the Holocene. In this table, as in the others in this chapter, only horizons with thermophilic flora are shown. These horizons correspond to the subdivisions of the Pliocene and interglacial epochs distinguished in the stratigraphic scheme of Holland (Zagwijn, 1963). Data on the flora of glacial horizons are not shown, because those floras are not important in showing the evolution of forest floras.

South Ural forelands and Bashkiria

Changes at the genus level in the late Cenozoic dendroflora in this region are shown in Table 8.2. In this region, all the Pliocene horizons, as well as those corresponding to the interglacial epochs distinguished in the stratigraphic scheme of Yakhimovich (1970; Yakhimovich and Suleimanov, 1981), are present. Detailed paleomagnetic studies have been undertaken within the Bashkirian foreland, making possible comparison of the stratigraphic horizons with the paleomagnetic scale (Yakhimovich and Suleimanov, 1981). Long-term paleocarpological and palynological studies have provided materials that fully highlight the changes in composition of the dendroflora through the late Cenozoic deposits.

Primoriye

The materials from the Primoriye territory are relatively limited. The most diverse and consistent are those for the late Cenozoic in the south of the region, as published in two generalized monographs (Korotkiy, Karaulova, and Troitskaya, 1980; Golubeva and Karaulova, 1983). Table 8.3 shows Upper Cenozoic horizons in the stratigraphic scheme of Korotkiy (Korotkiy et al., 1980). Unfortunately, there have been no systematic paleomagnetic studies within Primoriye, but the studies by Alekseev (1978) and Korotkiy et al. (1980) make it possible to establish a definite link between the stratigraphic scheme in this region and the paleomagnetic scale.

Northern Italy

The paleofloristic history in this region is presented in Table 8.4, based on the stratigraphic scale of Selli (1967). The paleomagnetic investigations undertaken in Italy have pertained mainly to its southern part, with only a few for northern Italy. Nevertheless, biostratigraphic analysis makes it possible to compare the stratigraphic horizons with the paleomagnetic scale (Ryan, 1973). In this region, the paleobotanical material consists mainly of fossil pollen, and we have detailed descriptions of macrofloral remains (leaves, seeds, etc.) only for the Astian Stage (i.e., continental Upper Pliocene) and certain parts of the Calabrian Stage.

Kura lowland and southeastern Greater Caucasus foothills

Data on this area are given in Table 8.5, based on the stratigraphic scale of Isaeva-Petrova (1972). Paleomagnetic research here has been extensive, and there are no doubts as to the comparison of the stratigraphic sequence with the paleomagnetic scale (Grishanov et al., 1983). The abundant paleobotanic material has allowed us to document the late Cenozoic dendrofloral history in detail.

We cannot dwell on the voluminous body of information concerning the western Caucasus region in this chapter. In particular, the situation in the Kolkhida, or Chalcedonia, with its enormously rich relict flora, makes it hardly comparable with other territories. However, similar general tendencies are quite clearly reflected here as well.

The Pamirs

Generalizations about the paleobotany of the Pamirs are difficult because of many controversies about Upper Cenozoic stratigraphy (Pakhomov, 1980). The compilation in Table 8.6, which reflects the changes of the fossil dendroflora in the Pamir deposits, is based on the stratigraphic scheme of Chediya (1971). The great volume of paleomagnetic data from this region makes it possible to link this succession quite definitively to the paleomagnetic scale (Dodonov, 1980). At the generic level, the late Cenozoic history of the continental dendroflora is fairly well known (although with a certain degree of fragmentation), mainly through the palynological studies of Pakhomov (1980, 1983).

Summary

The data given in Tables 8.1–8.6 show that one clear and sharp change was associated with or was quite close to the paleomagnetic reversal at the Gauss–Matuyama boundary, coeval with the transition from Reuverian to Tiglian paleofloras in Holland. Another sharp change between "Astian" (i.e., Piacenzian) and Calabrian paleofloras in Italy is close to the Olduvai subchronozone. In eastern Europe, the most important reorganization was associated with the middle horizons of the Akchagylian. Research on the history of the flora of the southern European part of the former USSR has shown that this phenomenon was associated with the establishment there of a climate with winter periods characterized by temperatures below freezing (Grichuk, 1959). In that period, representatives of tropical and North American genera completely disappeared except for certain species of North American origin that were preserved in the southern and western Caucasus.

In the Lower Pleistocene stratigraphic horizons that overlie beds with lower Matuyama paleomagnetic polarity there is only a gradual reduction in the other groups that are now extinct in the respective regions. The final disappearances of genera belonging to presently "alien" groups (now occurring in restricted or displaced ranges, as compared with their Pliocene distributions)

Table 8.1. *Dendroflora of Upper Cenozoic warm-climate horizons of The Netherlands*

Geographical groups	GENERA	GAUSS		MATUYAMA ↓Olduvai		BRUNHES			
		Brunssum	Reuver	Tegel	Waal	Cromer	Holstein-Treene	Eem	Holocene
				Praetiglian	*Eburon*	*Menap*	*Elster*	*Saale* *Warthe*	*Weichsel*
Panholarctic	*Pinus*	----------	----------	----------	----------	----------	----------	----------	----------
	Salix	----------	----------	----------	----------	----------	----------	----------	----------
	Myrica	----------	----------	----------	----------	----------	----------	----------	----------
	Alnus	----------	----------	----------	----------	----------	----------	----------	----------
	Rhamnus	----------	----------	----------	----------	----------	----------	----------	----------
	Cornus	----------	----------	----------	----------	----------	----------	----------	----------
	Sambucus	----------	----------	----------	----------	----------	----------	----------	----------
	Viburnum	----------	----------	----------	----------	----------	----------	----------	
	Picea	----------	----------	----------	----------	----------	----------	----------	
	Abies	----------	----------	----------	----------	----------	----------	----------	
American-eurasiatic	*Carpinus*	----------	----------	----------	----------	----------	----------	----------	----------
	Corylus	----------	----------	----------	----------	----------	----------	----------	----------
	Fagus	----------	----------	----------	----------	----------	----------	----------	----------
	Quercus	----------	----------	----------	----------	----------	----------	----------	----------
	Ulmus	----------	----------	----------	----------	----------	----------	----------	----------
	Ilex	----------	----------	----------	----------	----------	----------	----------	----------
	Acer	----------	----------	----------	----------	----------	----------	----------	----------
	Tilia	----------	----------	----------	----------	----------	----------	----------	----------
	Fraxinus	----------	----------	----------	----------	----------	----------	----------	----------
American-mediterranean-asiatic	*Pyrus*	----------	----------	----------	----------	----------	----------		
	Vitis	----------	----------	----------	----------	----------	----------		
	Staphylea	----------	----------	----------	----------	----------	----------		
	Celtis	----------	----------	----------	----------	----------			
	Juglans	----------	----------	----------	----------				
	Castanea	----------	----------	----------	----------				
	Ostrya	----------	----------	----------	----------				
	Pterocarya	----------	----------	----------	----------				
	Rhus	----------	----------						
	Zelcova	----------	----------						
	Liquidambar	----------	----------						
	Aesculus	----------	----------						
	Stirax	----------	----------						
	Diospyrus	----------							
	Elaeagnus	----------							
American-eastasiatic	*Parthenocissus*	----------	----------	----------	----------	----------			
	Tsuga	----------	----------	----------	----------				
	Carya	----------	----------	----------	----------				
	Magnolia	----------	----------	----------	----------				
	Liriodendron	----------	----------	----------					
	Pyrularia	----------	----------						
	Nyssa	----------	----------						
	Stewartia	----------	----------						
	Fothergillia	----------							
	Meliosma	----------							
	Barchemia	----------							
	Torreya	----------							
Eastasiatic	*Eucommia*	----------	----------	----------	----------				
	Sciadopitys	----------	----------	----------					
	Phyllodendron	----------	----------	----------					
	Actinidia	----------	----------	----------					
	Halesia	----------	----------						
	Pseudolarix	----------	----------						
	Corylopsis	----------							
	Cunninghamia	----------							
	Cyclocaria	----------							
	Glyptostrobus	----------							
	Schizandra	----------							
N.A.	*Taxodium*	----------	----------						
	Sequoia	----------	----------						
Tr.	*Simplocos*	----------	----------						
	Alangium	----------							
Number of genera		**61**	**51**	**37**	**33**	**25**	**23**	**20**	**18**
Floral groups				**I.**			**II.**	**III.**	

Note: Cold-climate intervals are indicated in italics. Correlation of the paleofloral units of The Netherlands with the paleomagnetic scale is according to Zagwijn (Chapter 16, this volume).

Table 8.2. *Dendroflora of Upper Cenozoic warm-climate horizons of the Bashkirian piedmont of the Urals*

Geographical groups	GENERA	GAUSS	MATUYAMA ↓Olduvai		BRUNHES				
		Kumurlin N_2ak_1 Kartaman N_2ak_1	Voevode N_2ak_3 Akkulaeva N_2ak_2 Zilim-Vasil'y N_2ak_1	Upp. Apsheron N_2ap_3 Davlikanov N_2ap_2 Demsk N_2ap_1	Low. Pleistocene $Q_{3/I}$ (Chui-Atasiev)	Low. Pleistocene $Q_{1/II}$ (Likhvin)	Mid Pleistocene $Q_{3/II}$ (Odintsov)	Upp. Pleistocene $Q_{I/III}$ (Mikulino)	Holocene and modern Q_{IV}
		1, 2	*3, 4*	*5, 6*	*7, 8*	*9, 10*	*11, 12, 13*	*11, 14, 15*	*16*
Panholarctic	*Pinus*	----------	----------	----------	----------	----------	----------	----------	----------
	Abies	----------	----------	----------	----------	----------	----------	----------	----------
	Picea	----------	----------	----------	----------	----------	----------	----------	----------
	Salix	----------	----------	----------	----------	----------	----------	----------	----------
	Populus	----------	----------	----------	----------	----------	----------	----------	----------
	Betula	----------	----------	----------	----------	----------	----------	----------	----------
	Alnus	----------	----------	----------	----------	----------	----------	----------	----------
	Prunus	----------	----------	----------	----------	----------	----------	----------	----------
	Sambucus	----------	----------	----------	----------	----------	----------	----------	----------
	Viburnum	----------	----------	----------	----------	----------	----------	----------	----------
	Larix	----------	----------	----------	----------				
	Swida	----------							
	Myrica	----------							
American-eastasiatic-euroasiatic	*Corylus*	----------	----------	----------	----------	----------	----------	----------	----------
	Quercus	----------	----------	----------	----------	----------	----------	----------	----------
	Ulmus	----------	----------	----------	----------	----------	----------	----------	----------
	Acer	----------	----------	----------	----------	----------	----------	----------	----------
	Tilia	----------	----------	----------	----------	----------	----------	----------	----------
	Fraxinus	----------	----------	----------	----------	----------	----------		
	Carpinus	----------	----------	----------	----------	----------	----------		
	Ilex	----------	----------	----------	----------	----------			
	Fagus	----------	----------						
	Taxus	----------							
American-mediterr-anean-asiatic	*Celtis*	----------	----------	----------	----------	----------	----------		
	Pterocarya	----------	----------	----------	----------				
	Juglans	----------	----------	----------	----------				
	Zelcova	----------	----------						
	Elaeagnus	----------							
	Cerasus	----------							
	Vitis	----------							
	Paliurus	----------							
American-eastasiatic	*Tsuga*	----------	----------	----------	----------				
	Chamaecypris	----------							
	Carya	----------							
	Abelia	----------							
	Aralia	----------							
	Liriodendron	----------							
East-asiatic	*Actinidia*	----------							
	Ealauterococcus	----------							
	Phyllodendron	----------							
	Weigelia	----------							
Number of genera		**41**	**25**	**23**	**23**	**19**	**18**	**15**	**15**
Floral groups			**I.**			**II.**		**III.**	

Note: Correlation of the stratigraphic sequence (Yakhimovich, 1970) to the paleomagnetic scale is according to the data of Yakhimovich and Suleimanov (1981). The main fossiliferous sections are as follows: 1, Simbugino; 2, Belekes, Kumurly, Khabarovka; 3, Voevodskoye (lower flora); 4, Nagayevo, Tukayevo; 5, Akkulayevo; 6, Chiki-Anachevo; 7, Chui-Atasevo; 8, Tirlan'; 9, Baisakal; 10, Afonasovo; 11, Voevodskoye (upper flora); 12, alluvium of Terrace IV, Belaya River basin; 13, Gremyachy Creek; 14, Minueshty Creek; 15, alluvium of Terrace II, Belaya River basin; 16, floodplain alluvium of Belaya River basin.

Table 8.3. *Dendroflora of Upper Cenozoic warm-climate horizons of the Primoriye region, eastern Siberia*

Geographical groups	GENERA	GAUSS		MAT. ↓Olduv.	BRUNHES					
		Lower Sulfun Suite $N_{1/2}$	Upper Sulfun Suite $N_{1/2}$	Upper Krasnozvetna $N_{2/2}$	Ussurlya $Q_{1/I}$	Khankay $Q_{1/II}$	Sungach $Q_{3/II}$	Nakhodkin $Q_{1/III}$	Chernoruch'y $Q_{3/III}$	Holocene Q_{IV}
		1	1, 2	3, 4	5, 6	7, 8	5, 9	5, 10, 11	5, 12	13, 14
Panholarctic	*Pinus*	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Abies	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Picea	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Larix	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Betula	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Alnus	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Myrica	----------	----------	----------	----------					
American-eurasiatic	*Carpinus*	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Corylus	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Quercus	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Ulmus	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Tilia	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Fraxinus	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Acer	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Fagus	----------	----------	----------	----------	----------	----------	----------	----------	
	Castanea	----------	----------	----------	----------	----------				
	Ilex	----------	----------	----------	----------					
American-mediterr-anean-asiatic	*Juglans*	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Syringa	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Pterocarya	----------	----------	----------	----------	----------				
	Zelkova	----------	----------	----------	----------	----------				
	Rhus	----------	----------	----------	----------					
	Ostrya	----------	----------							
	Liquidambar	----------	----------							
	Celtis	----------								
	Morus	----------								
American-eastasiatic	*Aralia*	----------	----------	----------	----------	----------	----------			
	Tsuga	----------	----------	----------	----------	----------				
	Carya	----------	----------	----------	----------					
	Torreya	----------	----------	----------						
	Nyssa	----------								
Eastasiatic	*Phyllodendron*	----------	----------	----------	----------	----------	----------	----------	----------	----------
	Kalopanax	----------	----------	----------	----------	----------	----------			
	Weigelia	----------	----------	----------	----------					
	Cryptomeria	----------	----------	----------						
	Sciadopitys	----------	----------							
	Glyptostrobus	----------	----------							
	Engelhardtia	----------								
	Ginkgo	----------								
North-amer.	*Taxodium*	----------	----------							
	Sequoia	----------	----------							
	Planera	----------								
Number of genera		**42**	**36**	**30**	**28**	**23**	**19**	**17**	**17**	**16**
Floral groups				**I.**		**II.**		**III.**		

Note: Correlation of the stratigraphic sequence of Korotkiy et al. (1980) to the paleomagnetic scale is according to the data of Alekseev (1978) and Korotkiy et al. (1980). The main fossiliferous sections are as follows: 1, Perevoznaya Bay; 2, Povorotny Cape; 3, Krasnozvetna Series of Tokhtin depression; 4, Spassk-Dalny; 5, Ussuri-Khankai depression; 6, Bolshaya Ussurska River; 7, Melgunovka River mouth; 8, interfluve of Sungach and Ussuri rivers; 9, Terney village; 10, Vostok Bay; 11, Tumangan River; 12, Belaya Scala Bay; 13, Tal'ma Lake; 14, Amur Bay.

Table 8.4. *Dendroflora of Upper Cenozoic warm-climate horizons of northern Italy*

Geographical groups	GENERA	GAUSS	MATUYAMA ↓Olduvai		BRUNHES			
		Astian	Calabr. (Lower)	Donau-Günz	Günz-Mindel	Mindel-Riss	Riss-Würm	Holocene
		1, 2	3, 4	5	5	6, 7	8, 9	10
Panholarctic	*Pinus*	----------	----------	----------	----------	----------	----------	----------
	Abies	----------	----------	----------	----------	----------	----------	----------
	Populus	----------	----------	----------	----------	----------	----------	----------
	Salix	----------	----------	----------	----------	----------	----------	----------
	Betula	----------	----------	----------	----------	----------	----------	----------
	Alnus	----------	----------	----------	----------	----------	----------	----------
	Rhamnus	----------	----------	----------	----------	----------	----------	----------
	Viburnum	----------	----------	----------	----------	----------	----------	----------
	Picea	----------	----------	----------	----------	----------	----------	
	Larix	----------	----------					
American-eurasiatic	*Taxus*	----------	----------	----------	----------	----------	----------	----------
	Carpinus	----------	----------	----------	----------	----------	----------	----------
	Corylus	----------	----------	----------	----------	----------	----------	----------
	Fagus	----------	----------	----------	----------	----------	----------	----------
	Quercus	----------	----------	----------	----------	----------	----------	----------
	Ulmus	----------	----------	----------	----------	----------	----------	----------
	Ilex	----------	----------	----------	----------	----------	----------	----------
	Acer	----------	----------	----------	----------	----------	----------	----------
	Tilia	----------	----------	----------	----------	----------	----------	----------
	Fraxinus	----------	----------	----------	----------	----------	----------	----------
American-mediterranean-asiatic	*Ostrya*	----------	----------	----------	----------	----------	----------	----------
	Castanea	----------	----------	----------	----------	----------	----------	----------
	Vitis	----------	----------	----------	----------	----------	----------	----------
	Diospyrus	----------	----------	----------	----------	----------	----------	----------
	Cedrus	----------	----------	----------	----------	----------	----------	
	Pterocarya	----------	----------	----------	----------	----------		
	Zelcova	----------	----------	----------	----------	----------		
	Juglans	----------	----------	----------	----------			
	Aesculus	----------	----------	----------				
	Laurus	----------	----------					
	Liquidambar	----------	----------					
	Platanus	----------						
	Amygdalus	----------						
American-eastasiatic	*Tsuga*	----------	----------	----------	----------	----------	----------	
	Carya	----------	----------	----------	----------	----------		
	Pseudotsuga	----------	----------	----------				
	Magnolia	----------	----------					
	Nyssa	----------	----------					
	Benzoin	----------						
	Sophora	----------						
	Sapindus	----------						
	Borchemia	----------						
East-asiatic	*Eucommia*	----------	----------	----------	----------			
	Keteleeria	----------	----------					
	Cephalotaxus	----------	----------					
	Ginkgo	----------						
North-amer.	*Planera*	----------	----------	----------				
	Taxodium	----------	----------					
	Asimina	----------						
	Ptelea	----------						
Tropical (pluricontinental, Southasiatic, Afroasiatic)	*Geonoma*	----------						
	Ficus	----------						
	Pheobe	----------						
	Litsea	----------						
	Cassia	----------						
	Machaerium	----------						
	Celastrus	----------						
	Sterculia	----------						
	Terminalia	----------						
	Leuconthoë	----------						
	Porana	----------						
	Persea	----------						
	Eugena	----------						
	Appolonias	----------						
	Boscia	----------						
	Pittosporum	----------						
	Combretum	----------						
Number of genera		**65**	**41**	**33**	**30**	**28**	**24**	**23**
Floral suites			**I.**			**II.**	**III.**	

Note: Correlation of the stratigraphic sequence of Selli (1967) to the paleomagnetic scale is according to the data of Ryan (1973). The main sections are as follows: 1, Asti; 2, Mongardino; 3, Principe; 4, Stirone; 5, Leffe; 6, Pianura Padana; 7, Ca Marcozzi; 8, Padova; 9, Pianico-Selleri; 10, Polgaria.

Table 8.5. *Dendroflora of Upper Cenozoic warm-climate horizons of the Kura depression and the southesat piedmont of the Urals*

Geographical groups	GENERA	GAUSS	MATUYAMA ↓Olduvai		BRUNHES			
		Product-Ivna	**Akchagyl**	**Apsheron**	**Baku**	**Khazar**	**Khvalyn**	**Holocene**
		1	2, 3	4, 5	6, 7	6, 8	9, 10, 11	
Panholarctic	*Pinus*	------------	------------	------------	------------	------------	------------	------------
	Salix	------------	------------	------------	------------	------------	------------	------------
	Populus	------------	------------	------------	------------	------------	------------	------------
	Betula	------------	------------	------------	------------	------------	------------	------------
	Alnus	------------	------------	------------	------------	------------	------------	------------
	Rhamnus	------------	------------	------------	------------	------------	------------	------------
	Lonicera	------------	------------	------------	------------	------------	------------	------------
	Picea	------------	------------	------------				
	Abies	------------	------------	------------				
	Myrica	------------	------------					
American-eurasiatic	*Carpinus*	------------	------------	------------	------------	------------	------------	------------
	Corylus	------------	------------	------------	------------	------------	------------	------------
	Fagus	------------	------------	------------	------------	------------	------------	------------
	Quercus	------------	------------	------------	------------	------------	------------	------------
	Ulmus	------------	------------	------------	------------	------------	------------	------------
	Acer	------------	------------	------------	------------	------------	------------	------------
	Tilia	------------	------------	------------	------------	------------	------------	------------
	Fraxinus	------------	------------	------------	------------	------------	------------	------------
	Ilex	------------	------------	------------	------------			
American-mediterranean-asiatic	*Celtis*	------------	------------	------------	------------	------------	------------	------------
	Pyrus	------------	------------	------------	------------	------------	------------	------------
	Rhus	------------	------------	------------	------------	------------	------------	------------
	Vitis	------------	------------	------------	------------	------------	------------	------------
	Punica	------------	------------	------------	------------	------------	------------	------------
	Elaeagnus	------------	------------	------------	------------	------------	------------	
	Pterocarya	------------	------------	------------	------------	------------		
	Ostrya	------------	------------	------------	------------	------------		
	Parrotia	------------	------------	------------	------------	------------		
	Castanea	------------	------------	------------	------------	------------		
	Juglans	------------	------------	------------	------------			
	Buxus	------------	------------	------------	------------			
	Vitis	------------	------------	------------				
	Morus	------------	------------	------------				
	Zelcova	------------	------------	------------				
	Cedrus	------------	------------	------------				
American-eastasiatic	*Carya*	------------	------------	------------	------------	------------		
	Tsuga	------------	------------	------------	------------			
	Nyssa	------------						
	Libocedrus	------------						
	Aralia	------------						
East-asiatic	*Paulovnia*	------------	------------					
	Thuja	------------						
	Platicarya	------------						
N.A.	*Taxodium*	------------	------------	------------				
	Sequoia	------------	------------					
Tr.	*Cinnamomum*	------------						
	Persea	------------						
Number of genera		**46**	**38**	**35**	**28**	**25**	**21**	**20**
Floral suites			**I.**		**II.**		**III.**	

Note: Correlation of the 1963 MSC (Modern Stratigraphic Code) standard sequence of the USSR to the paleomagnetic scale is according to data of Grishanov et al. (1983). The main sections are as follows: 1, Baku or Bakinsky Archipelago, western Caspian; 2, Shirak steppe; 3, Kvabebi; 4, Lengibiz Ridge; 5, Oblivnoi Island; 6, Tagirkent; 7, Divichi; 8, Kysyl-Burun; 9, Kudialchai; 10, Shura-Ozen; 11, Binagady.

Table 8.6 *Dendroflora of Upper Cenozoic warm-climate or interglacial sequences of the Pamir Range, Tadjikistan*

Geographical groups	GENERA	GAUSS	MATUYAMA Olduvai↓			BRUNHES				
		Polizak	Kuruksay		Kayrubak	Vakhsh	Ilyak	Dushanbe		Amudar
		"Gan" subtillite	Nizhne-kilimbin	Kokbai intergl.			Akdzhar intergl.	Early Altyndar	Late Altyndar	Holocene
		1	2	3, 4, 5			6, 7	8	9, 10	11
Panholarctic	*Juniperus*	----------	----------	----------			----------	----------	----------	----------
	Ephedra	----------	----------	----------			----------	----------	----------	----------
	Salix	----------	----------	----------			----------	----------	----------	----------
	Betula	----------	----------	----------			----------	----------	----------	----------
	Pinus	----------	----------	----------			----------	----------	----------	
	Picea	----------	----------	----------			----------	----------	----------	
	Alnus	----------	----------	----------			----------	----------	----------	
	Rosa	----------	----------	----------			----------	----------	----------	
	Cornus	----------	----------	----------			----------	----------	----------	
	Abies	----------	----------	----------			----------			
	Rhododendron	----------	----------	----------			----------			
	Populus	----------	----------							
American-eurasiatic	*Hippophaë*	----------	----------	----------			----------	----------	----------	----------
	Corylus	----------	----------	----------			----------	----------	----------	
	Quercus	----------	----------	----------			----------	----------	----------	
	Ulmus	----------	----------	----------			----------	----------	----------	
	Acer	----------	----------	----------			----------	----------	----------	
	Tilia	----------	----------	----------			----------	----------	----------	
	Fraxinus	----------	----------	----------			----------	----------	----------	
	Carpinus	----------	----------	----------			----------			
	Fagus	----------	----------	----------						
	Ilex	----------								
American-mediterranean-asiatic	*Pistacia*	----------	----------	----------			----------	----------	----------	----------
	Rhus	----------	----------	----------			----------	----------	----------	----------
	Elaeagnus	----------	----------	----------			----------	----------	----------	----------
	Cedrus	----------	----------	----------			----------	----------	----------	
	Juglans	----------	----------	----------			----------	----------	----------	
	Celtis	----------	----------	----------			----------	----------	----------	
	Berberis	----------	----------	----------			----------	----------	----------	
	Platanus	----------	----------	----------			----------	----------	----------	
	Tamarix	----------	----------	----------			----------	----------	----------	
	Pterocarya	----------	----------	----------			----------	----------		
	Zelcova	----------	----------	----------						
	Ostrya	----------	----------	----------						
	Morus	----------	----------							
	Vitis	----------	----------							
	Zygophyllum	----------								
	Liquidambar	----------								
Amer.-eastas.	*Tsuga*	----------	----------	----------			----------			
	Carya	----------	----------	----------						
	Menispermum	----------								
	Fothergillia	----------								
Eastasiatic	*Platycaria*	----------	----------	----------						
	Engelhardtia	----------	----------	----------						
	Cerdidiphyllum	----------	----------							
	Glyptostrobus	----------								
	Corylopsis	----------								
A	*Taxodium*	----------								
Tr.	*Altingia*	----------								
	Sabal	----------								
No. of genera		**51**	**40**	**36**			**30**	**26**	**25**	**8**
Floral groups			**I.**				**II.**	**III.**		

Note: Correlation of the stratigraphic sequence of Chediya (1971) to the paleomagnetic scale is according to data of Dodonov (1980). The main sections are as follows: 1, Orta-Uchkul' (lower levels); 2, Khyrga-Dara; 3, Orta-Uchkul' (upper levels); 4, Kokdzhar-Uchkul'; 5, Khiriak-Dara; 6, Akdzhar; 7, Karatau-1; 8, Lakhuti; 9, Ogzikichik; 10, Khudzhi; 11, Shugnou.

are observed somewhat above levels containing the Brunhes–Matuyama boundary.

At the stratigraphic level corresponding to the Olduvai event, sharp changes in dendroflora at the specific level have not been recorded. On the other hand, at the genus level all Eastasiatic genera, with the exception of *Eucommia*, disappeared from northern Italy and Holland at that time, including such cosmopolitan American–Eastasiatic genera as *Taxodium*, *Torrea*, *Magnolia*, and *Nyssa*. On the whole, those changes were similar in scale to the later major floristic changes seen during the Pleistocene.

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9 Plio–Pleistocene mammal faunas: an overview

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Introduction

The evidence from mammalian biostratigraphy with regard to the boundary between the Neogene and the Quaternary, or, more precisely, the base of the Pleistocene, can be analyzed in various ways. Long-established regional biostratigraphic scales are still widely used, although for the most part they are based on uneven, inadequate evidence and thus are subject to differing and unreliable interpretations. In seeking for greater reliability, vertebrate paleontologists in the past few decades have attempted to subdivide the geological time scale into probabilistic mammal ages and mammal zones, such as the MN (Mammal Neogene) zones of Mein. That approach is equally subject to imprecision and subjective bias (De Bruijn et al., 1992; Fahlbusch, 1991), and scholars may, consciously or unconsciously, constrain the paleofaunal changes to coincide with a particular magnetic reversal, a climate event, or any other significant, even preselected, level.

The objective approach is to begin with the ages of local faunal horizons that are directly based on radiometric dates or paleomagnetic analysis and build a biochronologic framework on this ground (e.g., Lindsay et al., 1987). Interpolations are legitimate, as are correlations based on faunal similarities, when the limits of probable error are realistically included. In cases where a local fauna presents many first appearances or last appearances, it is reasonable to suspect a lack of information in the preceding or following time interval, respectively. When the sampling factor is corrected, if unusual numbers of earliest and latest occurrences are still seen at one level, a genuine faunal overturn can be inferred. Normally, such overturns are more likely to have been the result of shifting paleobiogeographic boundaries (due either to shifts in regional environmental conditions or to relaxation of previous physical barriers) than to have been coincidental speciation or extinction events in the endemic regional fauna.

Plio–Pleistocene mammalian faunal succession in Europe

Gauss chronozone. The pre-Pleistocene local faunas of fossil mammals from the older part of the Gauss chron in western Europe include those at Layna, with evidence of a change from warm to cold climate, Las Higueruelas, dated to 3.2 Ma or earlier (Aguirre and Morales, 1990), and Triversa (De Giuli et al., 1983). These are succeeded by the l.f. (local fauna) of Vialette, dated to about 3 Ma, and its correlative at Villarroya. In central Europe, the Hajnácka l.f. is also referred to the Gauss (Fejfar, Heinrich, and Heintz, 1990), as are the Ripa Skortselskaya l.f. (Konstantinova, 1967) and the Tulucesti-Cernatesti group of localities (Ghenea, 1981) in the southeast region of Europe (Figure 9.1).

Changes in the rodent paleofaunas in the middle or, perhaps, the earlier part of the Gauss chronozone are evident in the local faunas of Arondelli and Hajnácka in Europe and at Betekia in western Asia. In eastern Europe, *Dolomys* and *Borsodia* are characteristic of this interval, with the latter genus also known in Siberia and Transbaikalia. Species of *Mimomys* (*M. hintoni, M. polonicus, M. stehlini*) were widely distributed at the beginning of that time throughout Eurasia. Among large mammals, several last occurrences characterize the record of that time interval, while the FAD (first-appearance datum, or lowest stratigraphic occurrence) of *Mammuthus* occurs with *Equus* in Ripa Skortselskaya (Konstantinova, 1967). The genus *Equus* has been reported from Malusteni in Romania (Ghenea, 1981; Chapter 20, this volume) and Etulia in Moldava, both localities being assigned to younger Gilbert ages (i.e., 4 Ma, at least). No other localities with *Equus* older than 3 Ma are known in Europe or in Asia, nor even in North America, where the genus appears to have had its origin (Lindsay, Chapter 30, this volume). This evidence suggests either that *Equus* evolved in the lower Danube basin more than 1 m.y. before it appeared anywhere else or that the attribution of the Romanian and Moldavan sites to the Gilbert is wrong.

The earliest occurrences of *Equus* in western Europe are documented in the Rincón 1 local fauna (Alberdi et al., 1982) and in the Roccaneyra and Étouaires local faunas in France, dated to 2.5 Ma. A later occurrence was reported from the Montopoli l.f. in Italy, which was placed in the lowest level of the Matuyama chron (Lindsay, Opdyke, and Johnson, 1980), and the Huélago-Carretera l.f. of Spain (Alberdi et al., 1989) is probably of similar age. In these two latter faunas, *Mammuthus* is also present. The FAD of *Mammuthus* in western Europe is in the Valdeganga l.f., in a horizon geographically and chronologically close to Rincón 1 (Alberdi et al., 1982). The dispersal of the true elephants and horses into this part of Europe is thus certain to have occurred before 2.6 Ma.

Among a number of faunal events in these horizons we should mention the LAD (last-appearance datum, or highest stratigraphic occurrence) of *Ursus minimus, Cervus perrieri*, and *Arvernoceros ardei* and the FAD of *Procamptoceras* and *Gallogoral*. The changes in ruminant populations can be dated to just before the Gauss–Matuyama paleomagnetic reversal. The local faunas of Kaltensundheim and Stranzendorf C illustrate the same time slice for central Europe, the former with normal magnetization, and the latter virtually coincident with the Gauss–Matuyama reversal, according to Rabeder (1981; von der Brelie et al., Chapter 17, this volume).

Lower Matuyama chronozone. The fossil record for mammals in Europe between the base of the Matuyama chronozone and the Réunion excursions (2.6–2.2 Ma) is poor. During much of that time span, marine regression, tectonism, and erosion reduced the sedimentary record over wide areas of the continental regions. On the other hand, a number of reliably dated post-Réunion sites are known. These include the following: Senèze, with reverse polarity, above a normal (Réunion?) excursion, which would indicate an age of 2 Ma or less (Boeuf, 1990); Le Coupet, dated to 1.92 Ma (Bonifay, 1991); Chilhac, which is not much younger than 1.85 Ma (Boeuf, 1990). Saint-Vallier and Puebla de Valverde have yielded fossil assemblages no younger than those from the dated sites and thus should be dated to approximately 1.7 or 1.8 Ma. The preservation of the Saint-Vallier fossils in a loessic formation is indicative of cold (Eburonian?) climate. In eastern Europe, the Liventsovka l.f. (Virina, Dobrodeev, and Faustov, 1971) and the lower part of the Kryzhanovka section (Tretyak and Volok, 1974) have a reversed remanent paleomagnetic polarity that corresponds to a part of the Matuyama between Réunion and Olduvai. The Gräunceanu l.f. and Tetoiu l.f. in Romania (Bolomay, 1965; Radulesco and Samson, 1990) are pre-Olduvai as well.

Several biochronologically important events in the evolution and dispersal of Arvicolidae occurred during that time, which we can consider as latest Pliocene, but the dating is not completely established. Repenning (1987) estimated an age of 2.6 Ma for the migration of *Clethrionomys, Lagurodon*, and *Pliomys* into Europe; further documentation of the actual FAD of *Pliomys* comes from the lower horizons of Liventsovka (Aleksandrova, 1976), which are significantly older than 2.1 Ma and perhaps near the base of the Matuyama, and from Wėze, which is considered to be older than 3.0 Ma. Evolution in *Mimomys* produced *M. pliocaenicus* near the time of the Gauss–Matuyama reversal (Vangengeim, 1977; Nikiforova and Aleksandrova, 1987), according to the conventional correlation of the base of the Khaprovian "complex" in which the FAD of this species in European Russia is recorded. Chaline (1986; Chapter 14, this volume), however, has assigned that event to a level within the lower Matuyama. The material referred to *Mimomys* cf. *M. pliocaenicus* at Rincón 1, Spain, is assigned to the later Gauss (as mentioned earlier); this species is well documented in all horizons at Liventsovka and Kisláng, together with *M. reidi*, and also at Gundersheim.

In the interval between the Réunion and the Olduvai subchronozones, the faunae are distinguished by several LADs and FADs and by a number of taxa that occur exclusively in this particular group of sites. The last occurrences of *Mastodon, Anancus, Hipparion, Nyctereutes, Gazella, Gazellospira*, and *Gallogoral* are noted here, and the new cervid taxa *"Cervus" rhenanus, Pseudodama pardinensis*, and *Pseudodama philisi* did not survive into the Olduvai subchronozone. In addition, the LAD of *Croizetoceros ramosus* in western Eurasia appears to be at Chilhac. Most of the Tegelen faunal succession can be correlated with this assemblage (Masini and Torre, 1990), although uppermost Tegel levels probably date to the lower part of the Olduvai subchronozone.

Upper Matuyama chronozone. Evidence of rapid replacement in arvicolid species starts with the FAD of *Allophaiomys* in the upper Tegel deposits of The Netherlands (Van Kolfschoten, 1990), in the lowermost Olduvai subchronozone. Such replacements appear to have taken place almost simultaneously in North America (Lundelius et al., 1987; Repenning, Fejfar, and Heinrich, 1990; Repenning and Brouwers, 1992). The species *A. deucalion* occurs slightly later in strata dating to the upper Olduvai at Villány 5 and Kadzielna (Vangengeim, 1977), and possibly at Orce (as discussed later). The FAD of *Mimomys tornensis*, whether at Kadzielna or at the age-equivalent site of Almenara-Casablanca (Esteban Aenlle and López Martínez, 1987), the FAD of *Mimomys pusillus* in the upper horizon at Liventsovka, the FAD of *Villanyia exilis* in Villány 5, and the FAD of *Mimomys ostramosensis* in the lower horizons of Kolinany also occur within the age limits of the Olduvai subchronozone (cf. Rabeder, 1981).

Among large mammals, the earliest occurrences of *Panthera gombaszoegensis, Canis etruscus, Pachycrocuta brevirostris, Pseudodama nestii*, and *Eucladoceros dicranios* have been reported from Olivola (Azzaroli et al., 1986), jointly with the last occurrences of *Chasmaporthetes* and *Procamptoceras*. The precise age of Olivola has been difficult to establish; De Giuli and Masini (1986) proposed a post-Olduvai age, on faunal evidence that it postdates the Tegelen fauna, in rough correlation with climatic cooling and the Aullan diastrophism. This is, however, probably equivalent to the Eburonian cold-climate phase that actually begins in the uppermost Olduvai subchron (Pasini and Colalongo, Chapter 2, this volume). In the Almenara-Casablanca cave filling, where *Mimomys tornensis* is found together with *M. medasensis* and *Mimomys* aff. *M. rex*, we find the latest occurrence of *Gazellospira torticornis*, together with early occurrences of *Pachycrocuta brevirostris, Canis etruscus*, and *Pseudodama nestii* (Soto and Morales, 1985). At this site, a species of Ovibovini, more progressive than the *Megalovis* from Senèze and resembling the *Praeovibos* from Sinzelles (Moyà-

Ma	Magnetic scale	Western Europe	C-Europe & Balkans	Eastern Europe	South Africa	C-East Africa	N. Africa Mid.-East	North+West Asia
0.5								
0.6								
0.7	B M 0.73	*ATAPUERCA 3*						
0.8			Betfia 5	*KARAI-DUBINA*		OLDUVAI IV Δ		*LAKHUTI 2* Oler
0.9	0.88 0.94	*SOLILHAC* *VALLONET*	*UNTER-MASS FIELD*	*NOGAISK-2*	Swartkrans 3			*VIATKINO*
1		Monte Peglia Pirro		*SENNAYA BALKA*	Swartkrans 2			
1.1		Imola Venta Micena					Ubeidiya	
1.2			D. Altenburg 4 Betfia 2			OLDUVAI III Δ		*LAKHUTI 1*
1.3		Cueva Victoria	D. Altenburg 2	Psekups Upper		OLDUVAI II u Δ	Aïn-Hanech	
1.4		SINZELLES Casa Frata	Vcelare 3B1	*KRYZHANOVKA*		CHARI Δ SHUNGURA. L Δ		
1.5		Brielle	Osztramos 3	Dmanisi	Sterkfontein 5 Kromdraai A			Kizikha
1.6			Kolinany 3		Swartkrans 1	OLDUVAI II m Δ		
1.7	1.72	Olivola *TEGELEN 6*			Kromdraai B3	SHUNGURA. J Δ OKOTE Δ		
1.8		*ORCE 2* Almenara	*KADZIELNIA* Villany 5			OLDUVAI II low Δ		
1.9	1.88	CHILHAC COUPET	Kislang *SLATINA 2*	*LIVENTSOVKA*		SHUNGURA. H Δ KBS_ Δ		*KURUKSAI*
2	2.04	SENEZE		Khapry		OLDUVAI I Δ		*PODPUSK* *LEBYAZHIE*
2.1		Saint Vallier	*RIPPERSRODA*					
2.2								
2.3						SHUNGURA. G Δ		
2.4		*MONTOPOLI*		GOMARETI		KALOCHORO Δ SHUNGURA. E Δ		
2.5	M G 2.47	ROCCANEYRA *RINCON 1*	*STRANZENDORF-C*	Kotlovina?	Sterkfontein 4	SHUNGURA. D Δ		*BETEKE*
2.6		ETOUAIRES	*KALTENSUNDH.*			BURGI Tuff-u Δ		
2.7						u. SHUNGURA. C	Aïn-Brimba	
2.8			*TULUCESTI*	*RIPA SKOR-TSELSKAYA*		KADA HADAR Δ	L. Ichkeul	
2.9	2.88	Villarroya				LOMEKWI Δ SHUNGURA. C Δ DENEN DORA		
3	2.96	VIALETTE HIGUERUELAS	*HAJNACKA*		*MAKAPAN 3*	TULU BOR Δ SHUNGURA. B Δ		
3.1	3.07	Triversa						
3.2	3.16							

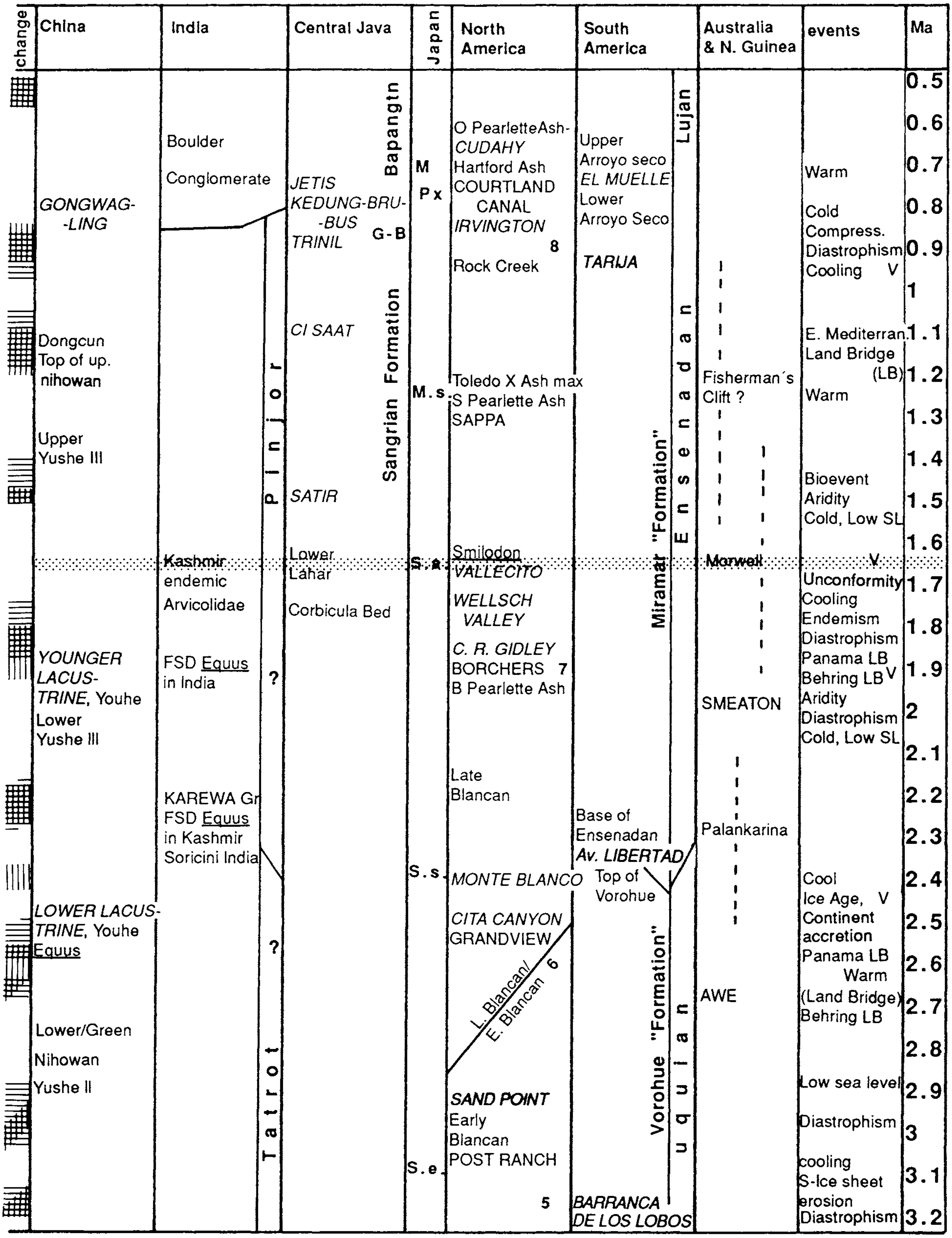

Figure 9.1. Pliocene–Pleistocene mammal correlations.

Solà and Menéndez, 1986), is also noteworthy. The Almenara-Casablanca l.f. is therefore most probably close to Olivola and Kadzielna in age, and possibly precedes Kolinany. All these localities probably should be dated to the upper part of the Olduvai subchron, if not to the top.

Normal magnetic polarity, possibly representing the Olduvai subchronozone, has also been reported from Orce 2, a horizon in the Baza Basin in southern Spain that yields a faunule that includes *Mimomys ostramosensis*, together with *Mimomys pusillus, Allophaiomys* cf. *A. deucalion, Apodemus mystacinus, Castillomys crusafonti, Gazellospira torticornis,* and *Leptobos etruscus* (Agustí et al., 1987; Aguirre, 1989b). The closest correlation of this fauna is to Kolinany, close to the top of the Olduvai subchron. A similar situation has been suggested for the Kotsakuri l.f. of Georgia (Gabunia and Vekua, 1981).

All of these faunal horizons together represent a faunal sequence that gives a fairly good picture of the diversity of large and small mammals in Europe at the end of the Olduvai subchron, and thus the time of transition from the Pliocene to the basal Pleistocene. The accelerated pace of evolution in mammalian faunas seen in this interval may represent a true faunal overturn, or an accelerated *tempo* of replacement, during this transition.

As a continuation of the same rapid pace of adaptation, some of the arvicolids that appeared in the lower Olduvai subchron became extinct near its top, such as *Villanyia exilis*, or slightly above this level, such as *Allophaiomys deucalion*, more or less simultaneously with the first appearance of *Allophaiomys pliocaenicus*. The occurrences of this taxon at Vcelare 3 B1 and Kamyk are probably its earliest record (Nadachowski, 1990), and it is widespread and abundant throughout Europe in other sites that date to just above the Olduvai subchron (Chaline, Chapter 14, this volume), at the beginning of an important evolutionary lineage for the Quaternary (Van der Meulen, 1973).

Almost at the same time, a number of changes occurred among the European large-mammal assemblages. The evidence comes from such sites as Sinzelles, with a date of 1.4 Ma (Bonifay, 1991); Cueva Victoria in southern Spain, which is similar to Sinzelles, or perhaps younger (Ferrández et al., 1989); and Casa Frata, an Italian locality stratigraphically equivalent to the ancient Tasso faunal assemblage, which has been dated to be older than Sinzelles and younger than Olivola (De Giuli and Masini, 1986). The immigrations of *Hippopotamus* and *Cervalces* are recorded at Tasso and Sinzelles, the latter also including the earliest record of *Praemegaceros* in western Europe. The Cueva Victoria l.f. is important for yielding the earliest record of *Dolichodoriceros* (= *Praedama*) and of *Cervus* ex gr. *C. elaphus* (Azanza and Sánchez, 1990). Fossil canids questionably attributed to *Canis falconeri* or *Xenocyon* are found at Casa Frata and Cueva Victoria, and the last mention of *Leptobos etruscus* also comes from these localities. The local faunas of Tetoiu 2 (Radulesco and Samson, 1990) and Lybakos (Steensma, 1988) in the Balkan basins correspond to this part of the Matuyama; Tetoiu 2 records the last occurrence of *Megalovis*. The new site of Dmanisi, in Georgia, with occurrences of *Bos, Capra, Ovis,* and a fossil *Homo* (Dzaparidze et al., 1991), has a local fauna in which most elements (other than the bovids) are held in common with the localities mentioned earler for the upper Matuyama time span, above the Olduvai subchron.

The local small-mammal faunas of Brielle (Van der Meulen and Zagwijn, 1974), Neuleiningen-15 (Fejfar and Heinrich, 1981, given as Neuleiningen-11), and Bagur (López, Michaux, and Villalta, 1976), as well as the local faunas of Psekups, Akkulaevo, and the "Odessan Complex" of eastern Europe (Nikiforova, Chapter 21, this volume), may correspond to the younger, relatively more stable part of this interval during the transition from cool climate to the warmer Waalian phase.

A succeeding interval of more rapid change began at about 1.2 Ma, the approximate age of Betfia 2, Deutsch-Altenburg, and Mas Rambault. A number of local faunas have been dated by interpolation in paleomagnetic profiles, such as the following: the central European site of Untermassfeld, with reversed polarity that is either just older or just younger than the Jaramillo subchron; Sennaya-Balka, Nogaisk, and Vallonet, all of which are contemporary with the Jaramillo (deLumley et al., 1988; Markova, 1990); and Solilhac, which also has been referred to the Jaramillo (Bonifay, 1991), but which more probably is related to a post-Jaramillo normal-polarity excursion. The youngest part of the Matuyama, from levels just prior to the Jaramillo up to the Brunhes reversal, also includes the undated but correlative local faunas of Betfia 2, Les Valerots, Monte Peglia, Cava Pirro, Imola, and Venta Micena (Agustí, 1986; Agustí et al., 1987; De Giuli, Masini, and Torre, 1990).

In the short interval from 1.2 Ma to the beginning of the Matuyama at 0.78 Ma, important changes occurred in mammalian microfaunas. The last *Mimomys* with closed-root incisors (*M. savini*) appears in Betfia 2 and in the upper Kryzhanovka level, dated to about 1.2 Ma, as well as Deutsch-Altenburg 4 and Cueva Victoria. In the same faunas, we find evidence of cladogenesis with the earliest appearance of *Microtus* (subgenus *Pitymys*) and related genera such as *Stenocranius* and *Iberomys*. The earliest occurrences of *Microtus sensu stricto* and *Pitymys* in central and western Asia are penecontemporaneous, in Itantsa and Razdolie, respectively (Vangengeim, 1977; Zazhigin, 1980). Beginning with the pre-Jaramillo levels at sites such as Betfia 2, Venta Micena, Monte Peglia, and Les Valerots we also find new species of *Allophaiomys*: *A. laguroides, A. nutiensis*, and *A. burgondiae* (Van der Meulen, 1973; Chaline, 1986). Finally, at approximately the same time, the dispersal of *Ellobius* and the immigration of *Eolagurus* are recorded in the Nogaisk 1 l.f. (Vangengeim, 1977; Markova, 1990).

In large-mammal faunas dating to the upper Matuyama, the most notable change is the regional diversification of *Mammuthus meridionalis* into progressive varieties (Aguirre, 1972): *M. m. tamanensis* from Sennaya Balka and many other localities in Ukraine, southern Russia, Georgia, and Romania, *M. m. vestinus* from Imola and Pirro in Italy, *M. m. wuesti* from central Europe, and perhaps other varieties. Remains of these progressive varieties have frequently been confused with *M. trogontherii*. As an immigrant from Asia, *Elasmotherium caucasicum* occurs in the upper Kryzhanovka l.f., shortly before the Jaramillo (Tretyak and Volok, 1974); *Dicerorhinus etruscus* must be read for *D. hemitoechus* at Solilhac. *Equus suessenbornensis* and *Equus altidens* replace the last representatives of the *E. stenonis* group, *Dama* ex. gr. *D. dama* replaces *Dama (Pseudodama) nestii*, and *Cervalces latifrons* replaces *C. gallicus*; at the same time, Megacerini and modern *Capreolus* become abundant. *Bison schoetensacki* and *Soergelia* also appear before the Matuyama–Brunhes reversal in some of these sites. The first species of *Crocuta*, new canids such as *Canis arnensis, Canis mosbachensis*, and *Canis tamanensis*, and ursids such as *Ursus deningeri* appear. The last occurrence of European *Acinonyx* is noted at Le Vallonet, and the last *Megantereon* is found in Untermassfeld.

Warm-climate episodes were followed by the Menapian cooling trend near the time of the Jaramillo and by an increase in compressive tectonics in various regions, together with uplifting and erosion, as, for instance, the Cassian unconformity in Italy. The succeeding faunas were but slightly modified from those of the Olduvai–Jaramillo interval, as, for instance, in Akhalkalaki (Vekua, 1987), Karai Dubina (Markova, 1990), Petropavlovka, and Huéscar (Alberdi et al., 1989), which are just slightly older than the Matuyama–Brunhes transition, and in Atapuerca TD3 (Gil, 1987; Aguirre, 1989a), a faunal horizon in which that paleomagnetic reversal is recorded.

Pliocene–Pleistocene mammalian successions in Africa

The African mammal assemblages seem to have been more stable than those of other continents. In the somewhat discontinuous record of southern Africa, accelerated faunal change has been reported between 3 and 2.5 Ma approximately, again between 2 and 1.5 Ma, and finally at about 1 Ma (Turner, 1990b). In East Africa, the more nearly continuous, radiometrically calibrated sequences, such as the Shungura Formation north of Lake Turkana, also display evidence of faunal change. Cooke (1978) recognized such changes within Member C (prior to 2.6–2.7 Ma) and within Member G (approximately 2 Ma). Those suggested changes do not coincide precisely with the interpretation of Maglio (1972) regarding the Koobi Fora Formation, northeast of Lake Turkana, where he proposed three zones: Zone 3, bracketed between the Tulu Bor Tuff (3.35 Ma) and the base of the KBS Tuff (1.95 Ma); Zone 2, from that level up to the base of the Okote Tuff (1.64 Ma); and Zone 1, from that level to the top of the preserved section. No major changes in bovid faunal composition were recognized by Cooke (1978) in the transition between Maglio's zones 2 and 1, around 1.6 Ma. Near to that level, however, Maglio placed the evolutionary change between the *Elephas recki* varieties 2 and 3, which coincides with the subdivision between lower Olduvai Bed II and its middle–upper part.

Overall, the major compositional changes of the southern and east-central fossil faunas of Africa appear to have been related to continent-wide turnovers, and the doubts expressed by Turner (1990b) about the possibility of correlation would seem to have been unfounded. Replacements at the species and variety levels in particular groups within diverse areas more probably were influenced by minor climatic cycles and vegetational changes.

Several species were consistently present in African faunas between 3 Ma and 0.74 Ma, among them the two living rhinoceroses, the extinct *Hipparion (Stylohipparion) lybicum*, the southern zebra *Equus capensis*, several bovids such as *Antidorcas recki* and *Oreotragus major*, and extant carnivores, including leopard, cheetah, serval, caracal, striped hyena, and spotted hyena. In faunas collected from levels of middle Gauss age, about 3–2.7 Ma, such as Makapansgat Mb 3 (McFadden, Brock, and Partridge, 1979), Shungura B, Denen Dora in the Hadar Formation, and the Tulu Bor level in Koobi Fora, we find the last occurrences of many Lower Pliocene taxa that characterize older associations at Langebaanweg, Kanapoi, Laetolil, and Usno: *Theropithecus darti, Australopithecus afarensis, Ancylotherium hennigi, Gazella vanhoepeni*, and *Simatherium kohllarseni* (Vrba, 1982). At slightly higher levels, for instance in the lower Burgi Member of Koobi Fora, in Shungura C and D of the Omo Basin, and in Sterkfontein 4, the most notable events are the last *Nyanzachoerus* and the earliest *Metridiochoerus*.

The next interval showing faunal changes in eastern Africa is bracketed between 2.0 and 1.75 Ma and includes local faunas from Shungura G to H, Olduvai Bed I–lower Bed II, and the upper Burgi and the lower KBS members at Koobi Fora. First occurrences during this interval are the African elephant *Loxodonta africana* in Shungura H and the earliest warthog, *Phacochoerus antiquus*, in lower Bed II at Olduvai (Cooke and Maglio, 1972), and the last *Notochoerus* are found at Shungura and Koobi Fora.

There are no known fossil-bearing beds within the 2.0–1.75-Ma interval in southern Africa, but several first and last occurrences have been noted in the fauna from Swartkrans 1 (the "hanging breccia" included) and in the Kromdraai B–East 3, roughly dated between 1.6 and 1.7 Ma (Partridge, 1982; Brain et al., 1988). Among the earliest appearances there are those of *Oreotragus major, Redunca* cf. *R. arundinum, Antidorcas australis, Rabaticeras arambourgi, Equus quagga*, and *Paranthropus robustus*, and the last occurrences of *Chasmaporthetes silberbergi* and *Hipparion steytleri* (Vrba, 1982; Klein, 1984). The upper part of Olduvai Bed II, above Tuff IIA, is of similar age, 1.67–1.65 Ma, with the first occurrences of *Elephas recki* stage III, *Damaliscus niro, Pelorovis olduwayensis*, and several suid taxa, and the last records of *Homo habilis, Parmularius braini, Parmularius angusticornis*, and *Mammuthus africanavus* in East Africa (Cooke, 1978).

Further first and last stratigraphic data have been noted in Kromdraai A, involving a number of carnivores (Turner, 1990a), and in Sterkfontein Mb 5, with an indirectly estimated age of about 1.5 Ma. At that time, a number of new suid taxa appeared in South African assemblages, including the Phacochoerini *Afrochoerus, Tapinochoerus*, and *Orthostonyx* (Cooke and Maglio, 1972). Members 2 and 3 of Swartkrans, nearing 1 Ma, show minor changes (Vrba, 1982). In fact, the fossil record in Africa between the middle of the early Pleistocene (1.25 Ma) and the middle Pleistocene (0.8 Ma) is relatively poor and discontinuous. In North Africa, late Pliocene and early Pleistocene faunas are becoming better known, but still seem to be quite discontinuous (Geraads, 1987); the Pliocene–Pleistocene formations, which include the sites of Aïn Boucherit and Aïn Hanech, deserve further study.

Pliocene–Pleistocene mammals of Asia and Australasia

Low sea levels associated with glacial maxima at the end of the Olduvai and again above the Jaramillo would have favored migrations between southwestern Asia and southern Europe, between eastern Africa and Indo-Pakistan via southern Arabia, and between the mainland of eastern Asia and the half-submerged peninsulas of Japan, the Philippines, Taiwan, and Indonesia, to say nothing of the Beringia lowlands.

In extreme southwestern Asia, the faunal association at Tell 'Ubeidiya in Israel includes a few taxa that also occur in the East African Lower Pleistocene levels, such as *Kolpochoerus olduvaiensis, Hippopotamus gorgops, Pelorovis olduwayensis*, and *Crocuta crocuta*, and one, *Equus* cf. *E. tabeti*, in common with the North African site of Aïn Hanech. The majority, however, have been identified with European species (Tchernov, 1988). Most of these species disappeared from Europe at about 0.8 Ma, but the 'Ubeidiya fauna includes a few that have been reported in Europe only at sites older than 1.6 Ma, plus several others that did not appear in Europe until after 0.8 Ma. Tchernov (1988), following the advice of F. C. Howell, referred to species of *Lagurodon* in estimating the correlation of 'Ubeidiya to an age between 1.4 Ma and 0.8 Ma. Another criterion of this age is the presence of *Mammuthus* (= *Archidiskodon*) *meridionalis tamanensis*, which is part of the diversified group of meridional mammoths found throughout Eurasia, from Britain to Japan, in the time span between the Jaramillo (1.1 Ma) and levels that date to slightly younger than 0.6 Ma. The combination of both criteria, taking into account the fact that the number of taxa shared in common with European sites younger than 1.1 Ma greatly exceeds the number in common with sites dated to between 1.4 and 1.1 Ma, suggests that 'Ubeidiya has an age close to 1.0 Ma. The species in common with East African assemblages at 'Ubeidiya agree with the more restricted time span.

It is easy to correlate the mammal assemblages from the western regions of Asia with the calibrated European sequences because of the high number of taxa common to both sides of the Urals (Vangengeim, 1977). Considerable effort has been devoted during the past two decades to fine-tuning the correlation of the local faunas of eastern Europe and northern Asia to the paleomagnetic scale. The Betekia l.f. in western Siberia corresponds to the Gauss chronozone (Gnibidenko and Pospelova, 1981), with faunal correlations to the Chikoi l.f. in Transbaikalia (Vangengeim, 1977). The Kuruksai l.f. in Tadjikistan has been referred to the lower Matuyama, with two horizons straddling the Réunion (Dodonov, 1986). Podpusk and Lebiazhie in western Siberia correspond to horizons with reverse polarity immediately below the Olduvai normal subzone (Gnibidenko, Erbaeva, and Pospelova, 1976). Local faunas of the Dodogol horizon in Transbaikalia and Kizhikia have been correlated with the Odessan faunal complex of eastern Europe by the presence of *Allactaga, Lagurodon,* and *Allophaiomys*; Dodogol also contains *Eolagurus*. The Razdolie l.f. of Siberia is correlated with the Tamanian complex (Vangengeim, 1977; Erbaeva, 1986). In Tadjikistan, the Lakhuti 1 small mammal fauna precedes the Jaramillo, and the Lakhuti 2 l.f., with large mammals, shortly precedes the Matuyama–Brunhes reversal (Dodonov, 1986; Sotnikova and Vislobokova, 1990).

Faunas in the Plio–Pleistocene Kopali Formation in Kazakhstan can be correlated to European and other western Asian sites by virtue of cosmopolitan rodent taxa that occur jointly with endemic species (Tjutkova, 1991). The Kazakhstan assemblage from Iliyskaya correlates with Kisláng and Villány 3 and 5 because of the presence of *Mimomys newtoni* and diverse species of *Villanyia*. The Tsharynskaya unit is probably younger than 1.7 Ma, considering the presence of *Allophaiomys pliocaenicus* together with diverse species of *Allactaga, Ellobius,* and *Allocricetulus*, a combination that in Europe dates to about 1.6 Ma. The units Jalanashkaya and Kopalinskaya are younger, between 1.4 Ma and 0.8 Ma.

Marked ecological barriers between Europe and Asia, as well as between diverse Asian regions, are reflected in many taxonomic differences among the ungulates at the genus level. Dispersal of the carnivore guild (*sensu* Turner, 1990b) was not impeded by ecosystem fragmentation to the same extent, and so fossil carnivores provide an important basis for correlation throughout the whole Eurasian continent (Sotnikova, 1991). The local faunas of Dongcun in the Chifeng Beds (Lu, 1991) and Nalaika in Mongolia (Sotnikova, 1991) include taxa in common with western Asian and (to a lesser extent) European localities of younger Matuyama age. The local faunas of Dodogol and Zasuhino 2, in Transbaikalia, have a number of taxa in common with European localities considered to date to the post-Olduvai interval, as well as several large mammals also known from the later part of the Nihewan faunal assemblage in China. A closer resemblance to the upper Nihewan sequence is found in the local fauna from the upper Zasuhino 3 horizon (Erbaeva, 1986; Vangengeim, Erbaeva, and Sotnikova, 1990).

The upper Nihewan Xiaodukou beds, with a post-*Hipparion* fauna and a considerable Paleolithic tool industry, is positioned just below the Jaramillo (Li and Wang, 1982), almost coincident with the transition from Wucheng loess to Lishi loess, and is dated to 1.11 Ma (Liu and Ding, 1983; Heller and Wang, 1991; Zhang, Chapter 26, this volume). The lower member of the Yuanmou Formation has a fauna generally characteristic of middle Villafranchian age; paleomagnetic analyses indicate that this unit extends up from the Gauss–Matuyama boundary to normal-polarity sediments, the correlation of which is uncertain because of compressed and possibly incomplete stratigraphy. The upper Yuanmou member, or Shangnabang unit, is above the normal-polarity strata and has a small fauna with *Stegodon orientalis* and *Homo erectus* (Zhang, Chapter 26, this volume). According to Liu and Ding (1983) and Jiang, Sun, and Liang

(1988), it belongs totally to the Brunhes, but it could reasonably be placed in the latest Matuyama above the Jaramillo, between 1.0 and 0.8 Ma. Either interpretation of the paleomagnetic record will be consistent with the faunal record of the upper Yuanmou, which is definitely younger than the pre-Jaramillo Xiaodukou faunas at Nihewan (Li, 1983b). Consequently, the age of the Transbaikalian Zasuhino 2 l.f. may be approximately 1.8–1.6 Ma, whereas the Zasuhino 3 l.f. and the similar Dongcun l.f., equivalent to the upper Nihewan Xiaodukou levels, are closer to 1.2 Ma.

High rates of faunal change, including rapid spreading of several taxa through Eurasia and their subsequent isolation and diversification, were the rule during the youngest part of the Matuyama, beginning just before 1.2 Ma and ending just after 0.8 Ma. Although Paleolithic tools indicate the presence of humans during that interval, the oldest known remains of *Homo* in China are specimens attributed to *H. erectus* from Gongwangling, Shaanxi province, almost at the top of the Matuyama, and perhaps (depending on interpretation) also at Yuanmou. Although the Gongwangling l.f. shares many elements with that of Zhokoudian and other mid-Pleistocene sites (Li, 1983b), its most probable correlation, from the broad paleofaunal viewpoint, is to the earliest mid-Pleistocene, prior to the Matuyama–Brunhes reversal. The close similarities with the Jetis paleofauna, with its reversed magnetic polarity (Zhou, Yanxian, and Wang, 1982), support this interpretation.

Japan. Kamei and Otsuka (1981) established a succession of mammal units for Japan, based on subspecies of proboscideans and controlled by abundant paleomagnetic data from the continental strata, that affords correlations to other Asian time scales. Kamei and Otsuka (1981) identified a major turnover in their oldest zone, called the *S. s.* (*Stegodon sugiyamai*) zone, which dates from about 3 Ma to the base of the Olduvai subchronozone. The turnover corresponds to the boundary between the Daodi and Nihewan (*sensu stricto*) local faunas in northern China and to the end of the Tatrot in India. In each instance these changes have been dated to the early Matuyama. The strongly marked turnover between the early Nihewan faunas and the typical levels, such as Xiashagou and Danangou (middle Yushe III in earlier notation), characterized by the expansion of Indo-Malaysian elements and the appearance of western taxa such as *Rusa, Eucladoceros*, and *Bison* appears to be of immediate pre-Olduvai age in the mainland faunas (Zheng and Cai, 1991); in Japan, that turnover has been correlated with the median part of the *S. a.* (*Stegodon akashiensis*) zone, which contains the Olduvai subchron and a part of the later Matuyama with reverse polarity. J. A. Van Couvering (personal communication) noted that the delay of this large-mammal turnover to the end of the Olduvai in Japan may reflect the fact that migration to those islands from mainland faunas is dependent on glacially lowered sea levels, such as at the beginning of the Pleistocene. The *S. a.* unit is succeeded by the *M. s.* (*Mammuthus meridionalis shigensis*) mammal unit, which extends to the base of the Jaramillo subchron. This fauna is distinguished by the presence of the defining subspecies, as well as several endemic cervids, suggesting a period of insularity and high sea level in Japan. Finally, the unit *M. px.* (*Mammuthus meridionalis proximus*) is characterized by a progressive form of *M. meridionalis* that appears in the Jaramillo and is replaced by *M. trogontherii* in the early part of the Brunhes, in Japan as well as in Europe.

Indo-Pakistan. There are diverse opinions as to the correct dating for the transition from Tatrot to Pinjor mammal ages in the Upper Siwalik sequence of northwestern India. The Tatrot and Pinjor paleofaunas are broadly distinguished by the presence of *Hipparion* and the absence of *Equus* in the former, and the presence of *Equus* and the absence of *Hipparion* in the latter (West, 1981). Several suid species that are found in Tatrot assemblages do not occur in the Pinjor, which is marked by the first regional appearances of *Bubalus, Hemibos*, and *Bos* (Badam, 1979; Yokoyama, 1981). The earliest true elephants, assigned to *Elephas planifrons*, are recognized already in the Tatrot and consequently precede *Equus* in this area. Yokoyama (1981) placed the base of the Pinjor within a normal-polarity interval; he identified it as the Olduvai, but that is somewhat questionable. Ranga Rao et al. (1981) noted that the evidence supported two possible correlations, the other being to a level within the Gauss closer to 3 Ma. In earlier works, a pre-Olduvai Pinjor faunal horizon was noted by Dodonov, Pevzner, and Penkova (1979), and the change from Tatrot to Pinjor assemblages was dated to the uppermost Gauss, at about 2.5 Ma (now 2.7), by Opdyke et al. (1979) in Pakistan. The definition of the basal Pinjor levels on the basis of the FAD of *Equus* is cited in all of those contradictory opinions.

Kotlia (1991) pointed to the arrival of *Equus* in Kashmir at 2.2 Ma, significantly earlier than the FAD of this taxon in the Siwaliks, according to Yokoyama (1981). Kotlia noted that the first appearances of arvicolids in Kashmir, at 2.4 Ma, were followed by a period of local evolution in that group, until about 1.6 Ma. The arvicolid immigration in Kashmir coincided with the appearances of Soriculini and Beremendini in the Indian foothills.

Java. In a thorough review of the mammal paleofaunas of Java and their stratigraphic position, Sondaar (1984) proposed the faunal sequence Satir, Ci Saat, Trinil, and Kedung Brubus, in ascending order, the last of which includes the Jetis l.f. (or Djetis). Paleomagnetic studies in the area by Sémah (1986) resulted in somewhat different age estimations than Sondaar's. A more consistent framework for the dating of formations and fossil assemblages in central Java was established as a result of multidisciplinary work on the Kendeng Group in the Sangiran area by Watanabe and Kadar (1985), with further refinements by Sudijono (1987) and Sémah (Chapter 28, this volume).

The earliest known Javanese fossil mammals appear to have lived under insular conditions; this is the Satir or upper Kaliglagah fauna, from lacustrine beds in the lower part of the Sangiran Formation, just above the normal Olduvai subchronozone. The earliest Satir fauna is therefore assigned an age of about 1.5 Ma. In the upper part of the "Black Clays" of the upper Sangiran, the Ci Saat l.f. represents a faunal change, with the presence of new immigrants. That change has been correlated with the Jaramillo, above a horizon dated at 1.16 Ma. The base of the Kabuh Formation is formed by the "Grenzbank" conglomerates containing the Trinil l.f., similar to that of Ci Saat. Farther east, on the Solo River, the Bapang Formation (equivalent to the Kabuh of Sangiran Dome) starts above the Jaramillo and extends up to levels dated to 0.5 Ma within the Brunhes. In the Bapang section, the Kedung Brubus and Jetis local faunas come from the middle of the unit, in a sequence of beds dated by a tektite fission-track age of 0.78 Ma and reverse polarity to the very top of the Matuyama chron (Aimi and Aziz, in Watanabe and Kadar, 1985, pp. 155–168).

Australia. Late Cenozoic fossil land vertebrates are known from sites in all parts of Australia, and in New Guinea as well. One of the earliest, the Awe l.f. of New Guinea, was assigned the early Pliocene by Plane (1967). In a comprehensive review of Australo-Papuan vertebrate paleontology, Rich et al. (1988) dated the Awe to an age between 3 and 2.5 Ma. Other Pliocene sites are the Chinchilla l.f. of Queensland, indirectly dated to an age younger than 4 Ma, and the Smeaton l.f., from beds overlying a basalt dated to 2.1 Ma. The Palankarinna l.f. of South Australia is recognized as younger than the Awe, but older than definitively Pleistocene assemblages, and the Morwell, Kanunka, and Bone Gulch are other local faunas given a transitional late Pliocene–early Pleistocene status. The local fauna of Fisherman's Cliff is more certainly attributed to early Pleistocene.

Changes in the late Pliocene and early Pleistocene mammal faunas of the Americas

North America. The Plio–Pleistocene interval in North America is documented in the fossil vertebrate faunas of the classic Blancan and Irvingtonian mammal ages, knowledge of which has been greatly advanced in the past few decades (Lindsay, Chapter 30, this volume). Thirteen distinct fossiliferous horizons have been synchronized with the paleomagnetic scale in San Pedro Valley, Arizona, between the Gilbert–Gauss boundary and the late Matuyama below the Jaramillo (Johnson, Opdyke, and Lindsay, 1975), resulting in an average interval between local faunas of 150 k.y. About 20 different horizons have yielded fossils of latest Blancan and Irvingtonian mammal ages in the paleomagnetically calibrated Vallecito Creek section of Anza-Borrego, in southern California (Downs and White, 1960; Lindsay et al., 1987), with an interval of less than 100 k.y. as an average. In addition, numerous fossiliferous sections in the High Plains of Kansas and Texas have now been correlated according to new evidence from paleomagnetism and tephrochronology (Lindsay, Johnson, and Opdyke, 1976; Lindsay, Chapter 30, this volume). Churcher (1983) has published a similar review of Canadian sites and their correlations with those to the south. From those studies it is clear that the chronologic ranges of individual taxa in different regions of the North American continent are diachronous, due to paleoecological and paleoclimatic factors. Scholars therefore prefer speaking of lowest and highest occurrences of taxa within local or regional stratigraphic formations, expressed here by the terms FAD (first-appearance datum) and LAD (last-appearance datum). Note that the use of LSD (lowest stratigraphic datum) favored by Lindsay (Lindsay et al., 1987; Lindsay, Chapter 30, this volume) is confusing, because "L" is also used for "last" in other stratigraphic abbreviations.

Dates for subdivision of the Blancan and for the Blancan–Irvingtonian transition have been controversial. Webb (1984) and other scholars have argued that the late Blancan should begin just after the Sand Point l.f., equivalent to the Kaena subchron at about 3.3 Ma. Others, however, would locate the type Monte Blanco level, with reversed magnetic polarity of the earliest Matuyama, in the early Blancan. A compromise has been suggested by those who favor the base of the late Blancan being traced just below the Monte Blanco bed, thus placing the division closer to the Gauss–Matuyama reversal (Lundelius et al., 1987), at 2.6 Ma. Criteria that have been proposed for the beginning of Irvingtonian time include the immigration of *Mammuthus*, the FAD of *Ondatra* and *Smilodon*, and the replacement of archaeolagine rabbits (*Hypolagus*) by the modern leporines such as *Sylvilagus* and *Lepus*. The latter event is evidenced almost simultaneously in the Curtis Ranch l.f. of San Pedro Valley, just below the Olduvai at about 2.0 Ma, and in the Borchers l.f. of Kansas immediately above the Pearlette B Ash, dated to 1.9–2.0 Ma. The first *Smilodon* in Vallecito Creek falls between 1.7 and 1.6 Ma, slightly above the Olduvai, and the first *Mammuthus* appears to date to about the same time. In this context, a strictly defined boundary would be difficult to establish, but the more extensive small-mammal record suggests a Blancan–Irvingtonian transition that would be older than, but close to coincident with, the Pliocene–Pleistocene boundary at the top of the Olduvai (Lindsay, Chapter 30, this volume).

Lundelius et al. (1987) revised the division of the Irvingtonian, as earlier proposed by Schultz et al. (1978), into three parts: in ascending order, the Sappan, Cudahayan, and Sheridanian. The base of the Sheridanian is defined by the Pearlette O Ash, dated to 0.61 Ma, and this unit therefore belongs totally to the middle Pleistocene. The Cudahayan is considered to include the Rock Creek l.f. and the Irvington l.f., the type of the Irvingtonian mammal age, with reverse magnetic polarity, interpreted as a part of the Matuyama above the Jaramillo. Also included in the Cudahayan is the Courtland Canal l.f., which underlies the Hartford Ash, dated to 0.74 Ma, and is thus close to the Matuyama–Brunhes boundary. The Sappan starts with the Borchers and Kentuck local faunas immediately overlying the 2.0-Ma Huckleberry Ridge or Pearlette B Ash in Kansas, while the type Sappa l.f. of Nebraska directly underlies the Mesa Falls or Pearlette S Ash, dated to 1.27 Ma. Thus, the Sappan-to-

Cudahayan transition would be placed between 1.2 and 1.0 Ma, close to the warm-climate interval preceding the Jaramillo subchron, and the base of the Sappan may be older than the Pliocene–Pleistocene boundary.

A number of significant faunal replacements that occurred during the late Gauss are recorded in strata younger than 3 Ma: for instance, the FADs of *Glossotherium* at several sites and of *Glyptotherium* in Great Plains faunas, marking South American connections, and the FADs of *Tremarctos, Paramylodon,* and *Tetrameryx* in southwestern sites such as Anza-Borrego. Datum events in faunas of early Matuyama age in the western USA are the LADs of *Hypolagus* and *Nannippus* and the FADs of *Ondatra* and *Nothriotheriops* between 2.4 and 2.2 Ma. The FAD of *Sylvilagus* at 1.9 Ma is followed by the appearance of *Mammuthus meridionalis* and *Allophaiomys* at the end of the Olduvai, in the Wellsch Valley l.f. of Canada, and the appearance of *Smilodon* in Anza-Borrego. In this same time span, the LADs of *Hypolagus, Borophagus,* and *Pliophenacomys* are seen in the Wellsch Valley fauna, and the last *Equus* and *Paradipodomys* occur in Anza-Borrego. The *Stegomastodon* LAD in the Great Plains and in the Gilliland local fauna of Texas occurs near the end of the Sappan, while *Cervus* and *Rangifer* first appear in the Cudahayan (Lundelius et al., 1987).

Repenning (1987) proposed somewhat different subdivisions of the Blancan and the Irvingtonian, based on faunal changes (immigrations, replacements) in microtine rodents, which he correlated with parallel events in Europe (Lindsay, Chapter 30, this volume, Figure 30.3). The Blancan IV unit, between events 5 (3.2 Ma) and 6 (2.6 Ma), displays stability in the endemic populations of the western USA. The Blancan V, between events 6 and 7 (1.9 Ma), is marked by the immigration of *Synaptomys (s.s.)* and *Synaptomys (Mictomys)*, synchronously with the appearances of *Clethrionomys, Pliomys,* and *Lagurodon* in Europe. The Irvingtonian I unit, between events 7 and 8 (0.85 Ma), is characterized by the immigration of *Microtus, Allophaiomys, Phenacomys,* and *Proneofiber* to North America and the contemporaneous appearances of *Microtus, Allophaiomys, Pitymys,* and *Dicrostonyx* in Europe. Event 8 itself is marked by the immigration of *Clethrionomys* and *Pitymys* into western USA faunas, which Repenning correlated with the dispersal into eastern Europe of the genera *Eolagurus* and *Lagurus* and with the maximum southern extension of the range of *Microtus* species.

South America. The general lithologic uniformities of the Upper Cenozoic continental formations of South America, together with the high degree of endemism in the successive vertebrate paleofaunas and the latitudinal and geographic environmental diversity, make it difficult to correlate the local biostratigraphic divisions to the global scale. Even though the abundance and diversity of fossil mammals in South America offer an outstanding opportunity for studies on biodiversity and evolution, their distribution through time has been subject to diverse interpretations, so that many geologists lack confidence in their utility as stratigraphic tools and prefer to group them in parastratigraphic units. Tuning the South American mammal successions to the magnetostratigraphic scale, as recommended by Pascual and Fidalgo (1972), has met with partial success for the Ensenadan sequence of Buenos Aires. Paleomagnetic studies of beds attributed to the lower part of the pre-Ensenadan Uquia Formation, from which sparse remains representing not more than nine mammalian genera have been obtained, indicate an age between 3.4 and 2.5 Ma (Orgeira, 1987); the upper part is not fossiliferous. Considering the Uquian to be poorly documented and its typical fauna to be of Chapadamalalan age, Cione and Tonni (1995) have proposed Marplatan as a new southern South American continental stage for the interval prior to the Ensenadan. The unit is based on abundant fossils representing more than 120 genera collected from strata exposed in sea-cliffs near Mar del Plata, some 1,800 km south of Buenos Aires, and directly overlying the beds with the type fauna of the Chapadamalalan. A major faunal turnover, involving a great extinction of autochthonous South American taxa, attributed to climate change rather than to invading taxa from North America, is clearly evident at the top of the Marplatan. The small-mammal evidence suggests a correlation of the Marplatan with the upper part of the North American Blancan, and reversed paleomagnetic polarities in the upper part of the typical Marplatan, at Barranca de los Lobos, most probably reflect the lower Matuyama chronozone (Orgeira, 1990; Cione and Tonni, 1995). The upper beds of the Uquia Formation may also correlate to this level (Marshall et al., 1982; Orgeira, 1987). On that basis, it can be suggested that the beginning of the Ensenadan mammal age, with its well-recognized faunal turnover, should be dated to the upper part of the lower Matuyama.

Determining the duration of the Ensenadan is complicated by uncertainty as to definition of the Ensenadan–Lujanian boundary. This is rooted in the placement of the Arroyo Seco unit in the Barranca de los Lobos sequence near Mar del Plata, in which the Matuyama–Brunhes reversal has been observed. If the whole Arroyo Seco is included into the Lujanian South American Land Mammal Age, as suggested by Tonni et al. (1992), then the conspicuous faunal overturn at the base of the Lujanian must be located in the upper part of the Matuyama. On the other hand, MacFadden et al. (1983), in a study of the paleomagnetic profile of a composite section near Tarija, Bolivia, and other related sequences with several fossiliferous horizons, placed the base of the Ensenadan not in the early Matuyama but instead just below the Jaramillo, close to 1.2 Ma, and the Ensenadan–Lujanian transition within the Brunhes. The latter correlation would be acceptable only if the Arroyo Seco l.f. were retained in the upper Ensenadan.

Subdivision of the Ensenadan was long ago proposed by Ameghino (*fide* Pascual and Fidalgo, 1972) on the basis of an interbedded marine tongue near the middle of the type sequence. A more detailed analysis of differences between the faunas collected in the upper and lower sections is desirable; in modern studies, the Ensenadan usually is treated as a uniform faunal unit (Tonni et al., 1992), without consideration of even minor changes.

The South American Ensenadan of the type area and the North American Irvingtonian are similar in that both begin close to (but below) the Olduvai subchron and in the fact that both record high rates of exchange between North America and South America. As Cione and Tonni have shown, most of the North American invaders attributed to "Uquian" in earlier reports were actually part of the great Ensenadan exchange that followed some time after the mass extinction of autochthonous forms that opened the age. With regard to that extinction event, it should be noted that a pronounced austral cooling occurred just above the Gauss–Matuyama reversal, at about 2.5 Ma (Tonni et al., 1992), but that glacial episode appears to have been too early to have been synchronized with the early Ensenadan faunal crisis. No further step in cooling is recorded prior to a paleoclimatic deterioration near the Jaramillo, at about 1 Ma, which seems to have been too late. It is possible, therefore, to imagine that the great basal-Ensenadan extinction event might be close in age to the Pliocene–Pleistocene boundary, but the evidence for that is far from conclusive.

Conclusion

The peaks of faunal overturn and the intervals of stability were approximately synchronous in the mammal communities of different continents during the Pliocene and early Pleistocene. It can be inferred, not surprisingly, that the tempo of dispersal and/or evolutionary change in mammalian communities was influenced by major changes in the global environment.

In relating faunal overturns to geological and climatic events, one has to consider that the interdependence and development of those phenomena were not uniform. Occasional global events that lead to migration and dispersal may affect faunal stability and produce major departures from patterns of "normal" ecological dynamics. A "dispersal event" actually involves several concepts. In the first case, it can mean the immigration of one or more taxa into areas where they are pre-adapted for success, as in the exchanges via the filter of Central America or Beringia. In the second case, it may mean the expansion of endemic lineages that have evolved adaptations to exploit the resources of a larger region, such as the spread of *Allophaiomys, Microtus, Bison, Equus*, and the progressive forms of *Mammuthus meridionalis*. The third case may arise from the effects of a geographic expansion of favorable conditions. In the second case, the effect is enhanced if it coincides with the development of corridors or land bridges, but in each of the noted examples the dispersal would have come about through the same favorable adaptation: increased hypsodonty, with continued enamel production in the teeth-forming tissues. The success of such adaptations depends on vegetation, and that is favored by expansion of suitable plant associations, in this case the grasses of the family Poaceae; and that, in turn, is also an effect of climate change. Thus, there is an overlap with the third type of dispersal. Other instances of the third case are the dispersal of the carnivore "guild" (Turner, 1990b; Masini and Torre, 1990), that of the new cervids, *Cervalces* and the Megacerini, and that of *Hippopotamus*. All of those would have been predictable, taking into account the lowered sea levels and dominantly mild and moist climate up to the end of the Eburonian.

With respect to the stratigraphic questions, it is necessary to recognize that all of the earlier described examples of faunal overturns represent accelerated or powered processes that were not instantaneous, but time-transgressive events, the same as other geodynamic and paleogeographic changes. The type of events that are closest to a geological "instant" are the magnetic reversals. The ages and successions of land-mammal faunas are now better calibrated and more precisely understood, so that faunal changes, whether major or minor, global or regional, can be tuned to the astronomically forced climate cycles, paleomagnetic reversals, paleogeographical and tectonic events, as well as the global stratotype sections and points (GSSPs).

The basal boundary of the Pleistocene, as proposed by Aguirre and Pasini (1985) and adopted in the Global Stratigraphic Scale (Cowie and Bassett, 1989), is very close to the faunal overturns represented in *Allophaiomys* event in Eurasia, to the dispersal of the carnivore guild mentioned earlier, and to the changes in African fauna seen in Olduvai Bed II. The faunal horizons closest in time to these events are as follows: Olivola in northern Italy; Barranca de los Conejos, close to Orce 2 in Spain; Villány 5 in Hungary; Kadzielna in Ukraine; Shungura J7, the lowest part of the Okote beds, and the earliest faunal levels in Olduvai Bed II in East Africa; the *Smilodon* FAD in Vallecito Creek, southern California; the top of the Uquia Formation, Argentina.

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10 Human evolution in the Plio–Pleistocene interval

EMILIANO AGUIRRE

Introduction

For many years, the main problem in relating the Pliocene–Pleistocene boundary to human evolution has been the controversy over "Tertiary man." For some, this has had serious philosophical implications arising from the definition of the Quaternary period in anthropological terms. The development of modern stratigraphical standards, however, makes this a merely nominal question, because such boundaries are now based entirely on geological principles. At the same time, the terms "Tertiary" and "Quaternary," having been drawn from primitive concepts of stratigraphy that once also distinguished a "Primary" and a "Secondary," have been called into question. Nevertheless, the chronology of early humans is pertinent to the goal of this project, for other than merely historical reasons.

The search for fossil Hominidae in the past three decades has produced an immense amount of geological and paleontological evidence. In Africa, particularly, well-exposed and relatively thick Plio–Pleistocene sections yielding abundant remains of Hominidae and a wide diversity of other fossils have been very well characterized in terms of biochronology, and also with reference to magnetostratigraphy and isotopic dates. Similar data, if not so well developed, have been recovered from other areas as a direct result of paleoanthropological studies.

It is now clear that key events in human evolution occurred within the time interval that also encompassed the Pliocene–Pleistocene boundary (i.e., within the limits of the Matuyama paleomagnetic chron, from 2.6 to 0.8 Ma) (Figure 10.1). It was during that time that the genus *Homo* first appeared in Africa, giving rise to the key species *Homo habilis* Leakey, Tobias, and Napier, at about 1.8 Ma, almost exactly coincident with the beginning of the Pleistocene. At some later time, perhaps very soon after its evolution in Africa at about 1.6 Ma, the clade of *Homo ergaster/erectus* began to disperse into southern Eurasia.

Africa

Intensive research over the past few decades has uncovered a nearly continuous record of hominid evolution in Africa during the period in question that conclusively demonstrates the origin of our genus in the mammalian fauna of that continent. No less than four separate areas have yielded numerous fossil hominids in stratified context, each with abundant associations of tools and animals, and three of which are also well suited for geochronologic analysis.

Turkana Basin

The sediments deposited in the margins and tributary valleys of the Lake Turkana depression in northernmost Kenya and southern Ethiopia contain an incredibly rich fossil record covering middle Pliocene to middle Pleistocene time, as well as earlier (i.e., Lothagam, Kalodirr) and later (Kibish, Galana Boi, Guomde) sites of interest to human evolution. Three major collecting areas in this basin are geographically separate but stratigraphically correlative, as demonstrated by identification of numerous tuff horizons that extend throughout the basin (Feibel, Brown, and McDougall, 1989). White (1988) has published a comprehensive inventory of the hominid remains from the Turkana Basin, identified to generic level, within the time span between 2.6 Ma and 1.4 Ma.

Shungura Formation. In the lower Omo valley, north of Lake Turkana in Ethiopia, a nearly continuous and outstandingly fossiliferous section, with many radiometric, paleomagnetic, and geochemical calibration controls, has been mapped by teams led by F. Clark Howell (Coppens et al., 1976). The Shungura section is divided into members, lettered A to L, each defined by a volcanic tuff at its base and including many other tuffs within its thickness. The age determinations for the Shungura tuffs are analytically reliable and stratigraphically consistent. The most pronounced faunal change is observed in the interval between tuffs C and G (Aguirre et al., Chapter 9, this volume), dated to approximately 2.85 Ma and 2.33 ±0.03 Ma, respectively. Supporting dates have been obtained on tuffs D and F within the interval, dated 2.52 ±0.11 and 2.36 ±0.05 Ma (Brown et al., 1985a,b; Feibel et al., 1989).

Shungura Members C, D, E, and F are highly fossiliferous, each with more than 10 localities that have yielded fossil hominids. Of hominids, the fossil material includes isolated teeth

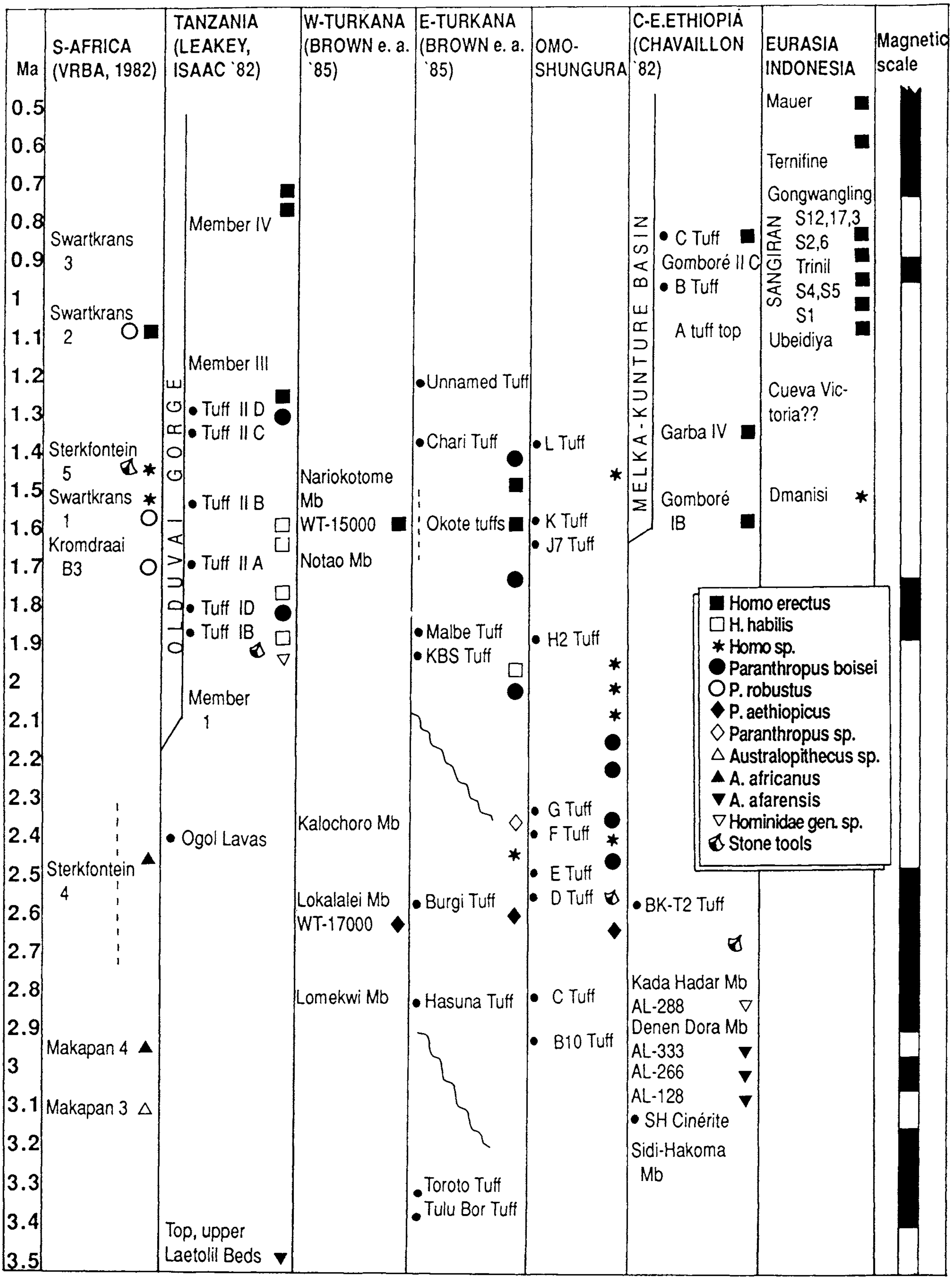

Figure 10.1. Principal occurrences of human fossils and paleolithic artifacts in the Old World during the late Pliocene and early Pleistocene. The paleomagnetic scale on the right shows the Matuyama chron between 0.7 and 2.5 Ma, with the Olduvai subchron normal-polarity interval at about 1.8 Ma.

(more than 130 in all), four mandibles, two fragmentary skulls, and several postcranial bones. Studies of these fossils are still largely unpublished. The earliest indications of tool-making, on stone and on bony material, also occur near the base of Member E, at 2.6 Ma. An assemblage of pebble and flake tools was found near the top of the same member, and artifacts become abundant at the base of Member F. It is noteworthy that the hominid remains from the Shungura below Member E are presently assigned to *Paranthropus aethiopicus* (Coppens) by Kimbell et al. (1988), whereas the earliest known remains of humans, classified as *Homo rudolfensis* Alexeev, are known from Koobi Fora, Chemeron, and Malawi only from beds dated to 2.6 Ma or younger, as discussed later. *Paranthropus boisei* (Leakey and Leakey) is recognized within Member E, and the first fossil remains attributable to *Homo* occur at the top.

The upper part of the Shungura Formation starts with Tuff G (2.33 Ma) and ends with Member L (ca. 1.4 Ma); members H, J, and K are much poorer in fossil hominid remains than the lower members. Within Member H, Tuff H-2 is dated 1.88 ±0.02 Ma, and Tuff L has a date of 1.39 Ma (Brown et al., 1985b). The top of the Olduvai subchron is found in Member J, and Cerling and Brown (1982) correlated Tuff J-7 with a tuff in the Okote complex at Koobi Fora, with an age measured at the time close to 1.6 Ma. Hominid fossil remains from Member G include two fragmentary skulls, a maxilla, 4 mandibles, 45 isolated teeth, and several postcranial bones.

East Turkana. Radiometrically dated tuffs in the Koobi Fora and Guomde formations east of Lake Turkana also cover the time interval we are concerned with. Several depositional hiatuses are recognized in the Koobi Fora sequence: an unconformity in the strata between the Tcroto Tuff (3.32 ±0.02 Ma) and the Burgi Tuff (2.6 Ma), and a larger hiatus representing most of the time between the Burgi Tuff and the KBS Tuff (1.88 Ma) (Brown et al., 1985b; Feibel et al., 1989). The upper part of the Koobi Fora sequence, above the KBS Tuff, is calibrated on the basis of the Okote complex of tuffs (1.6–1.5 Ma) and the Chari Tuff (1.39 ±0.01 Ma), another tuff dated at 1.25 ±0.02 Ma, and at the top the Silbo Tuff (0.74 ±0.01 Ma) (Brown et al., 1985b; Feibel et al., 1989).

Several localities in the lower part of the Koobi Fora, at levels straddling the Burgi Tuff, have yielded fossil hominids, but most of the extremely abundant and generally well-preserved fossil hominids in East Turkana occur in the upper part of the Koobi Fora Formation, from just below the KBS Tuff up to the Chari complex. In the lower Koobi Fora, both *Paranthropus* and *Homo* are recognized in beds dating to about 2.5 Ma (White, 1988, table 1), though according to my 1977 notes on the KNM collection, the specimen numbers of KNM-ER 1482 on a mandible assigned to *Homo* and KNM-ER 1483 on a mandibular body without tooth crowns assigned to *Paranthropus* were transposed in that publication. Wood (1992) considers mandibles KNM-ER 1482 and 1483, and also the well-preserved crania KNM-ER 1470 and 3732 just below the KBS Tuff and dating to about 2.0 Ma, to be among a number of specimens in this age range that should be considered as an independent species, *Homo rudolfensis* Alexeev, 1986. Other crania and mandibles from just below the KBS Tuff (i.e., KNM-ER 1813, 3735), as well as many others from the KBS Member above the tuff, were retained by Wood in *H. habilis*. The alternative view can also be supported, that is, that the differences between the nearly coeval 1470 *rudolfensis* and 1813 *habilis* crania are sufficiently explained by sexual dimorphism, whereas differences seen in the other specimens from near the KBS Tuff can be ascribed, quite reasonably, to intraspecific variation.

In this regard, a new mandible from the Chiwondo Beds of Malawi, UR-501, is said to be closely comparable to the specimen KNM-ER "1483" (*vel* 1482) of the lower Koobi Fora (T. Bromage, personal communication, 1992). The Malawi specimen is also of roughly the same age as "1483," to judge by faunal context, but whether these older specimens are correctly *Homo*, and not *Paranthropus* or even a directly ancestral *Australopithecus*, will be difficult to determine without a better definition of the cladogenetic event. In any event, many characteristics in these specimens are at the extremes of their ranges of variation for *H. habilis*.

In the upper Koobi Fora, *Paranthropus boisei* is recognized in many fossils from the KBS and Okote members up to but not above the Chari Tuff, at 1.4 Ma. As to human remains, a fossil skull, KNM-ER 3733, from the upper KBS Member just below the Okote Tuff complex, is the earliest specimen to be identified to *Homo erectus*; its stratigraphic position is dated slightly older than 1.6 Ma. Within the Okote Member, coeval with the youngest *P. boisei*, we find a cranium (KNM-ER 3883) and mandibles (KNM-ER 992, 730) of other *H. erectus*.

The East Turkana sequence has been correlated on the basis of physical stratigraphy of airfall tuffs, as well as by radiometric dates, paleomagnetic reversals, and paleontology, with the Shungura Formation north of Lake Turkana, with West Turkana, with the Awash region in the Ethiopian Rift Valley, and with Olduvai–Laetoli in Tanzania (Cooke, Chapter 27, this volume).

West Turkana. In the Nariokotome III site west of Lake Turkana, a nearly complete skeleton of an adolescent *Homo erectus* was discovered in 1985 (Brown et al., 1985a), with additional elements found in later field seasons (Leakey and Walker, 1989). The cranium, mandible, and most of the postcranium are exceptionally well preserved, allowing accurate estimation of the ratio of brain size to body weight and other pertinent comparisons to modern humans. Tuff correlations (Brown et al., 1985b) suggest an age close to 1.6 Ma for this specimen. At a lower level, within the Lomekwi Member, remains of *Paranthropus aethiopicus* are dated by tuff correlations to an age of 2.6 Ma, closely contemporaneous with the other occurrences of this species in the Shungura and Koobi Fora formations (Walker et al., 1986).

Lake Baringo. A hominid temporal bone from the Lake Baringo basin, in the upper or type Chemeron Beds, has been assigned to the genus *Homo* (*H. rudolfensis* Alexeev, according to Wood, 1992). This specimen is dated to about 2.45 Ma (Hill et al., 1992), approximately coeval with earliest remains and artifacts from the Turkana Basin. Even earlier artifacts, dated to about 2.85 Ma, are known from the Hadar Formation in Ethiopia, as discussed later.

Awash Valley

In central Ethiopia, the time interval of interest here is represented by two areas in the Awash Valley. The first area is that of the exposures of the Sagantole and Hadar formations, with a formational contact dated by the presence of the Lokochot (Shungura A) Tuff, at 3.47 Ma, and the Tulu Bor (Shungura B) Tuff, at 3.35 Ma, in the Sidi Hakoma Member of the lower Hadar Formation (Haileab and Brown, 1992). Australopithecine fossils are known from the Sagantole Formation at Belohdelie, dated to about 3.8 Ma, but otherwise all the human remains come from the middle part of the Hadar Formation in the Denen Dora and lower Kada Hadar members, in the interval between the base of the Denen Dora at 3.1 Ma and the BKT Tuff dated to 2.85 Ma (Haileab and Brown, 1992). Site AL 333 in the Denen Dora has yielded more than 300 specimens, attributed to a minimum number of 13 individuals, showing differences that can be considered as a close approximation of a speciation event in human evolution. From AL 288, just above an unconformity near the base of the Kada Hadar Member, the skeleton nicknamed "Lucy" has been assigned to *Australopithecus afarensis*, but it could reasonably be assigned to *Homo* instead. The earliest dated assemblage of stone artifacts in East Africa derives from the lower part of the Kada Hadar Member, just below the 2.85-Ma BKT Tuff. This tuff was previously K/Ar-dated to 2.63 ±0.05 Ma by Taieb and Tiercelin (in Chavaillon, 1982), and to 2.58 ±0.23 Ma by fission-track dating (Chavaillon, 1982; Harris, 1986).

The record in this region is continued at Melka-Kunturé, farther to the west, which is also well calibrated with paleomagnetic and radiometric dates (Chavaillon, 1982). The lower levels, with reversed polarity, are correlated to the upper Matuyama above the Olduvai normal. The oldest sites contain tools of mixed Acheulean–Olduwan aspect, and at Gomboré I-B part of a human humerus is dated to approximately 1.6 Ma. Higher sites, such as Gomboré II and Garba XII, have yielded extremely abundant tools of Acheulean traditions, and other human fossils have been found at Garba IV (1.35 Ma) and again at Gomboré II-C between Tuff B and Tuff C, dated 0.84 and 0.80 Ma, respectively. The fossil hominids of Gomboré I, Garba IV, and Gomboré II are attributed to *Homo erectus*. Finally, Tuff D, dated 0.74 Ma, includes an artifact assemblage of evolved Acheulean tradition.

Olduvai Gorge

In central northern Tanzania, the Rift Valley encloses a broad sedimentary basin adjacent to an array of young volcanoes. The only place in which the Plio–Pleistocene sediments of this basin are deeply exposed is the unique canyon of Olduvai Gorge. Older Pliocene beds exposed in shallow gullies to the south at Laetoli, which have yielded australopithecine remains and footprints, as well as abundant mammalian fossils, appear to be a continuation of the Olduvai sequence (Cooke, Chapter 27, this volume).

The lower units (called "beds" by convention) of the Olduvai Gorge sequence, including most of the section and most of the fossil-bearing outcrops, have long been calibrated by radiometric dates, most recently by Walter et al. (1991). An age of 0.62 Ma has been estimated for the top of Bed IV (Leakey and Hay, 1982), while the earliest Bed I sediments at Olduvai, below the Lower Basalt in the upper gorge, date to approximately 2.0 Ma (Walter et al., 1991). The boundaries of the paleomagnetically normal strata representing the magnetostratotype of the Olduvai subchronozone overlap the interval from Tuff I-A to Tuff I-F, with fossiliferous deposits beginning at the level of Tuff I-B. The radiometric dating of these tuffs by Walter et al. (1991) suggested a duration for the Olduvai subchron between 2.0 and 1.75 Ma, which compares well to the astronomically tuned calibration of 1.95 to 1.79 Ma (Preface, this volume). Bed II, above Tuff I-F, extends up to about 1.5 Ma. Olduvai Bed IV (0.83–0.62 Ma) straddles the Brunhes–Matuyama magnetic reversal, which is recorded in Tuff IV-B.

The type specimen and associated material of *Homo habilis*, as well as specimens assigned to *Australopithecus boisei*, come from Bed I and lower Bed II. In particular, well-preserved remains of *H. habilis*, including KNM-OH7, OH8, OH13, OH24, OH35, and OH48, come from various levels, which range from just beneath Tuff I-B (1.95 Ma) to just below Tuff II-B (1.7 Ma). All of these fossils are younger than the hominids of Shungura E–G (before 2.0 Ma) and probably those found below the KBS Tuff of East Turkana (before 1.9 Ma), whereas on present dating they are slightly older than the earliest *Homo* at Melka Kunturé. The skull of OH9, attributed to *Homo erectus*, was discovered out of stratigraphic context but is inferentially related to a level in upper Bed II, above Tuff II-D. It is younger, in any case, than the *erectus* skull, KNM-ER 3733, in the upper KBS Member at Koobi Fora if the assumed dates are correct. The variability in dating does not preclude the possibility that the last *H. habilis* and the earliest *H. erectus* (i.e., *H. ergaster* of some authors) specimens are coincident. It is likely that representatives of the *H. erectus* clade were in fact present in the vicinity of the Rift Valley before 1.5 Ma, and perhaps before 1.7 Ma, according to the evidence from dated specimens at Olduvai and East Turkana. The robust australopithecines are not found above the top of Bed II, about the same age as the Chesowanja specimen (1.5 Ma); their last appearance is thus above the level of the robust specimen found at Peninj, west of Lake Natron (about 1.6 Ma), and below the level of the *Homo erectus* specimens associated with the Chari Tuff (1.4 Ma) at Koobi Fora and Shungura in the Turkana Basin.

There is a general lacuna in the East African hominid record from the level of upper Bed II until the uppermost Upper Pleistocene strata of Kanam (Lake Victoria), dated to about 1.1

Ma, Olduvai Bed IV, Olorgesailie (0.95 Ma), and Guomde (East Turkana), dated to about 1.0 Ma.

South Africa

The chronology of Plio–Pleistocene fossil Hominidae from the South African caves is less precise than that in East Africa, with no radiometric dates and only a debatable paleomagnetic record from Makapansgat. Two alternative age sequences have been proposed for the fossils from the younger "Australopithecine caves." The first is correlated in a morpho climatic approach, set out by Partridge (1982), while the second follows faunal correlations with East African sequences by Cooke (1978), Vrba (1982), and others. The two approaches almost coincide in their estimations for the most likely age of the main level of Sterkfontein (St. 4, with *Australopithecus africanus*), suggesting an age range between 2.8 Ma and 2.4 Ma. The upper level of Sterkfontein (St. 5, with *Homo* cf. *H. erectus*) is estimated to have an age between 1.6 and 1.4 Ma, if not slightly younger, which would correlate to a level in upper Olduvai Bed II.

The dates that have been suggested for the other sites are less consistent, regardless of the approach. Makapansgat Member 3, the earliest australopithecine level, has been assigned an age of just over 3 Ma by Partridge (1982), based on the interpretation that middle Gauss (Mammoth and Kaena) paleomagnetic polarities are recorded in the overlying Member 4, also with hominid remains. The paleomagnetic profile, however, could be interpreted as the upper part of the Gauss, with an age between 2.8 and 2.4 Ma, like the St. 4 main level at Sterkfontein. All of the fossil hominines from Makapansgat are commonly attributed to *A. africanus*, but Aguirre (1970) suggested that the lower jaw of an immature individual resembles *Paranthropus* in some features. This question should now be reassessed in view of the discovery of a primitive species of that genus, *P. aethiopicus*, from beds of that age in the Turkana Basin.

The age of the deposits of Swartkrans 1, with the robust australopithecine *Paranthropus* and indications of early *Homo*, should be 1.9–1.6 Ma, according to the faunal correlations, just as at Olduvai, whereas Partridge (1982) estimates it to be between 1.55 and 1.2 Ma, and thus significantly younger than the youngest known *Australopithecus* from the well-dated East African hominid record. The age of Swartkrans 2, with *H. erectus*, but no *Paranthropus*, should be 1.0 Ma, according to Vrba (1982), but Partridge (1982) estimates it as much younger, corresponding to early middle Pleistocene. Finally, two age estimates for Member B of Kromdraai, with robust australopithecines, are 2.1–1.8 Ma (Vrba, 1982) or 1.25–1.0 Ma (Partridge, 1982).

Northwest Africa

Among the fossil Hominidae known from Middle Pleistocene sites in Northwest Africa, the skull and associated material of the "advanced *Homo erectus*" of Tighenif (also known as Palikao and Ternifine), Algeria, must be mentioned, because the most recently estimated age there is 0.7 Ma or even older (Hublin, 1985). Other specimens of younger age, between 0.5 and 0.4 Ma, have been found in coastal-plain terraces on the Moroccan coast at Salé, Rabat, Sidi Abderrhaman, and Thomas Quarry (Casablanca), having features attributed to the transition from *erectus* to *sapiens*.

Dispersal event in Eurasia

The earliest human fossils outside of Africa, in order of increasing distance from mainland Africa, are the cranial fragments of Tell 'Ubeidiya in Israel, which in fact may be considered paleogeographically as an outlying part of the Afro-Arabian biotic province, the mandible recently discovered at Dmanisi, Georgia, the skull from Gongwangling (and possibly the teeth from Yuanmou) in China, and a partial cranium and two mandibles from the lowest levels at Sangiran, Java. All of these are assigned ages older than the Matuyama–Brunhes reversal, although questionably so in the case of Gongwangling, and much more questionably in the case of Sangiran. Most modern authors, however, believe that all the fossil hominids from Sangiran and Trinil date from the Matuyama, and a recent redating of the lower Sangiran (Swisher et al., 1994) has suggested that the Java record of *erectus* may be as old as in Africa. Overall, the Eurasian finds indicate that the earliest members of this species expanded rapidly into a much wider geographic range than had their African precursors, although perhaps in tropical areas before more temperate ones. Two localities in Spain, Orce and Cueva Victoria, have also been suggested to yield fossil hominids older than 1 Ma, but the evidence is not constraining in either of these two cases (Aguirre, Chapter 13, this volume).

Because their fossil remains are so rare, the dispersal of humans in Eurasia has been amplified appropriately by studies of archeological material, including tools and living sites. The most acceptable conclusion from this kind of evidence points to the early Pleistocene, just prior to 1 Ma, as the time when humans first became established outside of Afro-Arabia, in the warm-winter zones of Eurasia from the Near East to India, Java, and northern China. In agreement with older dates in Java (Swisher et al., 1994), fossil remains from the pioneer population, at present known only from the Caucasus (Dmanisi) and central Java, show plesiomorphic features in common with the earliest populations of *Homo erectus* (i.e., the earliest-evolving stem distinct from *H. habilis*) known in Africa. A later dispersal introduced humans into western Eurasia close to the beginning of the Brunhes chron (0.7 Ma), again documented primarily from archeological records. That dispersal involved a population of more advanced character, whose oldest fossil remains (e.g., Atapuerca, Arago, Petralona) have close similarities to East African material of the same age: the later part of the early Pleistocene to the early part of the middle Pleistocene (Aguirre et al., 1980; Arsuaga et al., 1993).

Middle East

In Tell 'Ubeidiya, Israel, human fossil remains assigned to *H. erectus* have been found in association with early Acheulean tools. The site has been dated, with some difficulty, according to uncertain correlation of the mammalian fauna. Most likely the age of the 'Ubeidiya humans could be placed in the later part of the early Pleistocene (Tchernov, 1987) at about 1.1 Ma.

What may be the oldest human fossil in mainland Eurasia was discovered at Dmanisi, Georgia, in 1991 (Dzaparidze et al., 1991). The specimen is a nearly complete, undeformed adult mandible that presents a mixture of progressive features, such as a molar series decreasing in size anteriorly, along with other features that are more primitive. It resembles *Homo habilis* and australopithecines in possessing such plesiomorphic features as a narrow mandibular body with slightly divergent horizontal branches, molar rows converging posteriorly, robust M1 with conspicuous cingulum, and an elongated P3 with a poorly developed lingual cuspule. A sinuosity in the vertical anterior profile of the symphyseal region creates a depression resulting in a chin-like prominence which resembles a feature seen in some australopithecines, in contrast to a true chin.

The fossil-bearing horizon at Dmanisi overlies a basalt dated to 1.8 ±0.1 Ma. The fauna associated with the mandible has been assigned to the Apsheronian Stage, or Lower Pleistocene in the Soviet schema (Nikiforova, Chapter 20, this volume). The fauna is of Upper Villafranchian character, with *Mammuthus meridionalis meridionalis, Canis etruscus, Bos*, and pre-caballine horses. Although it is not as old as the lower Apsheronian Kocachuri local fauna in which *Leptobos* rather than *Bos* is still present, the Dmanisi assemblage is also not younger than the levels at about 1.1 Ma in which advanced subspecies of *M. meridionalis* begin to appear. In other words, on preliminary assessment of faunal evidence, the age of the mandible appears to lie between 1.5 and 1.2 Ma, and if that should be substantiated, it will be further evidence of the dispersal of *Homo* out of Africa very shortly after the evolution of the *erectus* clade.

China

Paleomagnetic data suggest that the fossil skull of *H. erectus* from Gongwangling, Lantian area, has a pre-Brunhes age older than 0.8 Ma (Wu, 1985; Aguirre et al., Chapter 9, this volume), although Kahlke (1986) placed it younger than that, according to faunal correlation. Zhou (1990) proposed an age of 1.15 Ma, through correlation with the L15 loess, which agrees well with the dating of archeological material in the lower Nihewan sequence (as discussed later). Fossil teeth ascribed confidently to *erectus* from Yuanmou, Yunnan, were initially assigned an age of more than 1.6 Ma (*vel* 1.8) on the basis of paleomagnetic analysis. On faunal grounds, the underlying normal-polarity sediments have been reassigned to the Brunhes, so that the Yunnan fossils may be younger than 0.8 Ma (Liu and Ding, 1983), but the paleomagnetic interpretation and fauna are also consistent with a post-Jaramillo age, near 1.0 Ma. Finally, with regard to archeological evidence, in the Nihewan Basin of Hebei province, northern China, the occurrence of abundant paleolithic implements associated with late Villafranchian mammal remains at Donggutuo and several other pre-Jaramillo sites (Schick et al., 1991) appears to record human activity in that region at least as early as 1.2 Ma.

It is noteworthy that the ground-dwelling ape *Gigantopithecus* continued to live for a short time in southern Asia (and maybe also in Java) sympatrically with the earliest populations of *Homo erectus*. There may have been little competition between them in terms of ecological specialization and use of resources, as in the limited interaction between *Homo* and the other great apes, but perhaps the analogy should be drawn with *Paranthropus*, another large bipedal hominoid that became extinct shortly after the first Acheulean tool-users appeared in Africa.

Java

There are two opinions concerning the age of the "pithecanthropines," the original *Homo erectus* from Indonesia, even though the stratigraphy is fairly well established. The precise horizons from which the older collections came have been carefully identified, and an overall stratigraphic framework is generally accepted for the relative ages of the fossils recovered in different localities. The key unit is the Grenzbank, a conglomerate facies of variable thickness that rests on the Sangiran Black Clays and that underlies the Kabuh Formation at Sangiran on the upper Solo River (Sémah, Chapter 28, this volume). The Black Clays correlate to the Pucangan Formation and to the Ci Saat faunal assemblage, according to Leinders et al. (1985). Those authors further correlated the Trinil fauna found along the middle Solo River to the level of the Grenzbank and just below, and they assigned the Kedung Brubus fauna to the Kabuh Formation.

Opinions differ as to the relative ages of the local stratigraphic units, although recent opinion has held that the dating of a few decades ago was erroneously old. Magnetostratigraphic research (Shimizu et al., 1985; Sémah, 1986; Sémah, Chapter 28, this volume) strongly supports an upward revision for the earliest faunal levels, although the most recent dating (Swisher et al., 1994) is contradictory.

According to one alternative interpretation of paleomagnetic data, as discussed by Sémah (1986), there is enough uncertainty about the presence of reversed-polarity strata in the lower Kabuh that it is reasonable to place the entire Kabuh Formation in the Brunhes magnetic epoch, with the Grenzbank unit almost coincident with the Matuyama–Brunhes reversal. The age of the earliest hominid specimens of Java, those from the underlying Sangiran Black Clays, would thus be no older than the latest Matuyama, above the Jaramillo, and most of the classic "pithecanthropines," which come from the uppermost Sangiran–Grenzbank–lower Kabuh interval and the age-equivalent Trinil strata, would be dated between 0.6 and 1.0 Ma.

On the other hand, Shimizu et al. (1985) found that the top of the Jaramillo is clearly evidenced just below the basal Grenzbank, indicating that the controversial interpretations of reversed polarity in the Grenzbank and lower Kabuh are probably valid. In this view, the range of typical "pithecanthropines" in Java extended from the Matuyama–Brunhes reversal down to a level well below the Jaramillo, as discussed later. The dating of Swisher et al. (1994) on a tuff just above the oldest specimens at Sangiran (S27 and S31) indicates an age of 1.66 Ma, well down in the upper Matuyama. They also dated hornblende from normally magnetized volcaniclastic deposits at the site of the Modjokerto skull at Perning, chemically comparable to material from within the skull itself, at 1.81 Ma, consistent with an age within the Olduvai subchron and thus slightly older than the oldest dated *erectus*-grade hominids in Africa.

The biostratigraphic and radiometric data appear to be in better agreement with the older interpretation, rather than with the younger interpretation. In the biostratigraphic analysis of Leinders et al. (1985) the Kedung Brubus local fauna in the lower Kabuh Formation is correlated to an age older than 0.7 Ma. The fossils from Trinil and those of the Grenzbank at Sangiran are placed at the level of the Jaramillo subchron (1.1 Ma), while the material from the top of the Black Clays and the Ci Saat fauna are dated to about 1.3 Ma, or (*pace* Swisher et al., 1994) even older. Itihara, Kadar, and Watanabe (1985) noted that in regard to the three pumice tuffs identified within the Kabuh Formation (their Bapang), fission-track ages by Suzuki et al. (1985) dated the Upper Tuff to the lowermost Brunhes, and the Middle Tuff at 0.78 Ma to the latest Matuyama. That dating agrees with the determination that the 0.71-Ma tektites found at different levels in the middle Kabuh are in situ just above the Middle Tuff and that the Brunhes–Matuyama boundary is therefore correctly placed between these horizons.

Southern Europe

The attribution of a skull fragment and a fragment of bone from the locality of Venta Micena, near Orce (Gibert-Clols et al., 1989), to the genus *Homo* is questionable, although the stratigraphic position is well established in the Baza endorheic sequence (Anadon et al., 1987; Aguirre, Chapter 13, this volume). This level corresponds to the median third of the early Pleistocene and has a rich fauna younger than that of Sinzelles (1.3–1.4 Ma) and older than that of Vallonet (0.95 Ma) (Moyà-Solà et al., 1981). By interpolation, the specimen can thus be said to date from approximately 1.1 Ma, with an error margin of as much as 10%, which is about the same age and uncertainty as for the Tell 'Ubeidiya remains. The fossil in question is a small fragment of a cranial vault, including the lambda region, and what arguably may be a section of the coronoidal suture. The bone is very thin, with a very poorly developed pneumatized layer. That would indicate a very immature human, but it is inconsistent with a partial fusion of the sutures. Moreover, the shape of the suture is highly uncommon in humans, but is more often found in several other orders of Mammalia. A crestal ridge on the inner surface is quite similar in shape to one characteristically found in the Equidae (S. Moyà-Solà and J. Agustí, in Gibert-Clols et al., 1989), and very uncommon in humans. Nevertheless, immunological assay of the specimen has shown a closer affinity to humans than to horses (Gibert-Clols et al., 1989, pp. 225–228).

The Cueva Victoria fossil, found in a cave setting, is a second phalanx of the fifth finger (Gibert-Clols et al., 1989, pp. 395–406) that looks definitely human and cannot be attributed to any other known primate. Although it was not collected in situ, all the evidence leads to the conclusion that the Cueva Victoria phalanx came from a bone breccia that completely filled the cave in the early Pleistocene. The very abundant faunal remains, which are ascribed to hyena accumulation, are well correlated to the Sinzelles faunal assemblage (1.3–1.4 Ma) by Moyà-Solà and Menéndez (1986; Agustí, 1986; Aguirre et al., Chapter 9, this volume). Fossils of an undescribed giant baboon have been recovered from the Cueva Victoria breccia (J. Moyà-Solà, personal communication), but the possibility that this is the proper attribution for the phalanx can be positively ruled out (Pérez-Pérez, 1989).

Conclusion

Evolution of the genus *Homo* can be closely tied to the Pliocene–Pleistocene boundary, located in the uppermost part of the Olduvai subchron. The earliest known species, *H. rudolfensis*, was replaced by *H. habilis* within the time interval of the Olduvai, and shortly after it ended, the *H. erectus* clade evolved in Africa and rapidly spread into southeastern Asia. The dispersal of humans into southwestern Eurasia was not much later. The present state of knowledge therefore suggests that the first major dispersal of humans outside of Africa took place close to the Pliocene–Pleistocene boundary.

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Part IV

The Pleistocene boundary in regional sequences

11 The Pliocene–Pleistocene boundary in Italy

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Introduction

Italy is one of the primary sources of data for the Pliocene and Pleistocene. The Italian coastal margins were affected by tectonism during the Pleistocene more than any other part of western Europe, and as a result the Upper Cenozoic marine sequences that crop out along the coast in Sicily and in the mainland Apennine foothills (Figure 11.1) have been the subjects of classic studies on the chronostratigraphy of that age. Additionally, Plio–Pleistocene strata with stratigraphic thicknesses exceeding 8,000 m have been sampled in boreholes in the subsurface of the central Po Valley. Finally, Pliocene and Lower Pleistocene nonmarine strata are common in the montane basins of the Italian peninsula.

Since the very beginning of stratigraphy as a science, the abundance of Cenozoic fossils, particularly in the marine record, has attracted attention to the Italian sections. Lyell's original designations of the "older" and "younger" Pliocene (Lyell, 1833), the latter of which became the Pleistocene, relied heavily on the malacofaunas of the northern Apennines. For the same reason, all standard Pliocene and Pleistocene marine chronostratigraphic terms used in global correlation are based on the biostratigraphy of the Italian sequences. In recognition of this historical importance, the XVIII International Geological Congress in London in 1948 confirmed a general consensus that the Pliocene–Pleistocene boundary should be defined in the exposed Italian marine strata.

In this chapter we shall review the stratigraphic data available from the marine and continental records of Italy, as background for defining the Pliocene–Pleistocene boundary. The known Italian Plio–Pleistocene sections are so numerous that it would be impossible to treat them all in this brief review, and therefore we shall consider only those sections that are well documented and stratigraphically significant and that have been historically important in subdividing the Pliocene and the Pleistocene.

Biostratigraphic and chronostratigraphic framework

Several fossil groups (mollusks, ostracodes, foraminifers, calcareous nannofossils, pollen, etc.) are biostratigraphically useful in the Italian marine record. The most detailed and most reliable correlations are obtained through planktic foraminifera (Pasini and Colalongo, Chapter 2, this volume) and calcareous nannofossils (Rio, Raffi, and Backman, Chapter 5, this volume). The other groups are less reliable for long-distance correlations, although they can provide additional stratigraphic controls.

The biostratigraphic schemes based on planktic foraminifera and calcareous nannofossils, as discussed by the aforementioned authors, are shown in Figure 11.2. Figure 11.2 also shows the informal biostratigraphic subdivisions of the Lower Pleistocene of Ruggieri et al. (1976). Zones C, D, and E of Ruggieri and co-workers are defined, in ascending order, by the successive entrances into the Mediterranean basin of *Arctica islandica* (or other time-equivalent "northern guests"), *Hyalinea baltica*, and finally *Globorotalia truncatulinoides excelsa*. The chronostratigraphic subdivisions adopted for the Lower Pleistocene (Figure 11.2) are discussed in Ruggieri et al. (1984) and Rio et al. (1991). In particular, for the recognition of the Pliocene–Pleistocene boundary, we have used the definition proposed by INQUA Subcommission 1-d, "Pliocene/Pleistocene Boundary," and adopted by action of the IUGS, namely, the base of the claystone conformably overlying sapropelic layer *e* of the Vrica section in Calabria (Pasini and Colalongo, Chapter 2, this volume).

In reviewing the individual sections, we shall also discuss the available magnetostratigraphy in light of the biostratigraphic framework presented earlier. Cross-checks of magnetostratigraphy and biostratigraphy are crucial in the Italian sections considered here, because all of them are erosionally truncated and lack radiometric control. Therefore, only calibrated biostratigraphic events can allow correlation of the local magnetic-polarity reversals to the standard MPTS (magnetic polarity time scale). The chronometry followed here was originally developed according to the MPTS of Mankinen and Dalrymple (1979), but their values have been translated throughout this text according to the newly developed orbital time scale (Preface, this volume). The biostratigraphic events reported in Figures 11.2 and 11.9 have been tied to geomagnetic polarity, following Rio et al. (Chapter 5, this volume).

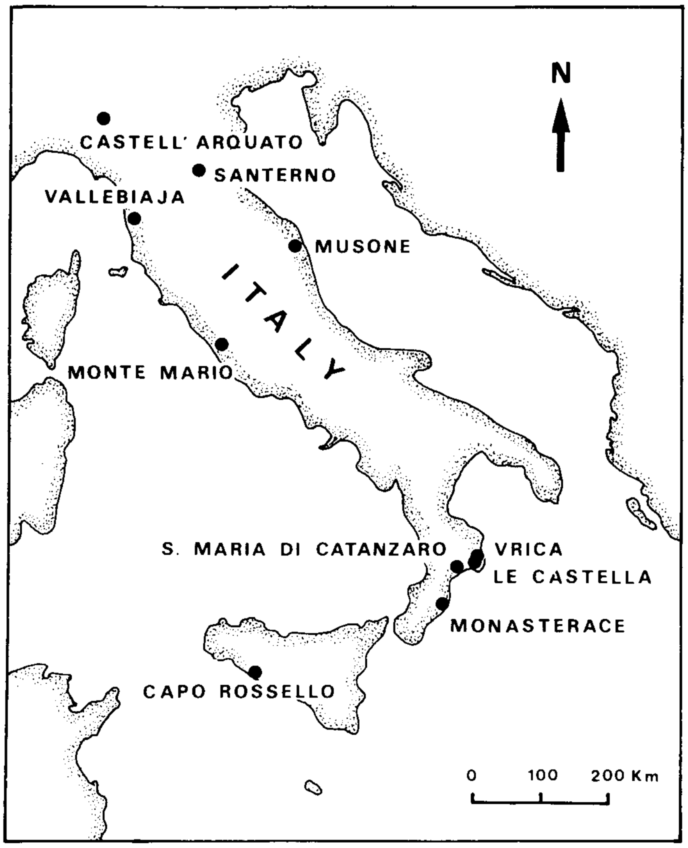

Figure 11.1. Locations of the Italian sections discussed in this chapter.

Marine record

We note, in reference to the marine criteria for subdividing the Upper Cenozoic of Italy, that *Globorotalia truncatulinoides excelsa* Sprovieri, Ruggieri, and Unti, 1981, is a subspecies representing the most advanced morphotypes of the species and is restricted to the later Pleistocene of the Mediterranean. The first occurrence of the typical *G. truncatulinoides truncatulinoides* in the Mediterranean is recorded at the stratigraphic level where *Globorotalia inflata* also first appears (Rio et al., 1984b). This stratigraphic level is considered to be equivalent to the base of the Upper Pliocene by scholars who use a threefold subdivision of the Pliocene (Colalongo, Elmi, and Sartoni, 1974). In the earlier literature, of course, the *excelsa* forms were not distinguished from the earlier, more typical subspecies.

Northern Italy

Santerno Valley. In the Santerno Valley (Bologna province), between the villages of Tossignano and Imola, a thick sequence of marine Pliocene and Lower Pleistocene sediments crops out in two measured sections: the Santerno River section and the West Santerno section.

The Santerno River section (Figure 11.3) has been sampled along the bed of the river except where alluvium or debris requires a lateral shift to the slopes of the valley. The sediments are poorly stratified gray-blue silty clays and sandy interbeds of turbiditic origin (Colalongo et al., 1982b). The sedimentary environment is bathyal in the lower and middle parts; at the top of the section, gray and yellow sands, more or less cemented, are indicative of a littoral–sublittoral environment (Colalongo et al., 1982b). The total thickness of the section is 2,600 m, according to Cremonini, Elmi, and Monesi (1969). Abundant macrofauna and microfauna have been reported in biostratigraphical studies by Ruggieri (1975), Colalongo (1968), Colalongo et al. (1974), Padovani and Tampieri (1970, 1974), Francavilla (1974), Raffi and Rio (1980a), Rio, Sprovieri, and Raffi (1984a), and Rio et al. (1991).

The Santerno River section is the most informative among the marine sections studied so far in northern Italy, and has been designated by Ruggieri and Sprovieri (1977) as the stratotype section of Santernian and Emilian chronostratigraphic units. The Santernian and Emilian, together with the Sicilian Stage of the Palermo area, were united as substages of the new Selinuntian Stage (Ruggieri et al., 1984), which those authors considered to be a replacement for the Calabrian Stage as a primary subdivision of the Lower Pleistocene.

The Pliocene–Pleistocene boundary traditionally has been traced in the Santerno section at the first appearance of the pelecypod *Arctica islandica*. According to the log of the Santerno section published by Cremonini et al. (1969), the level at which Ruggieri (1957) reported that event is 1,510 m above the base of the Pliocene, but subsequently Padovani and Tampieri (1970) showed the presence of this index fossil as low as 1,460 m. Actually, in the deep-water sediments of the Santerno River section, the shallow-water pelecypod *A. islandica* must have been reworked, and its appearance would have been delayed with respect to its first appearance in shallow water, such as in the Stirone section in western Emilia (Rio et al., Chapter 5, Figure 5.5, this volume). In fact, in the Stirone section, the first occurrence of *A. islandica* predates the first occurrence of *Gephyrocapsa oceanica s.l.*, whereas in the Santerno River section the first occurrence of *A. islandica* is recorded about 150 m above the first occurrence of *G. oceanica s.l.*

Rio et al. (Chapter 5, this volume) have reported the appearance of *G. oceanica* at around 1,300 m in the Santerno River logged section. According to M. L. Colalongo (unpublished data), the first occurrence of *Globigerina cariacoensis* and the last occurrence of *Globigerinoides obliquus extremus* are at about 1,440 m. These same three events are recorded close together, although in a different order, in the Vrica section just above the top of level *e*, the adopted definition for the Pliocene–Pleistocene boundary. Accordingly, the Pliocene–Pleistocene boundary of Vrica correlates in the Santerno River section somewhere around 1,300 m in the columnar log of Cremonini et al. (1969).

Magnetostratigraphy of the Santerno River section. The earliest analysis of the magnetostratigraphy of the Pliocene–Lower

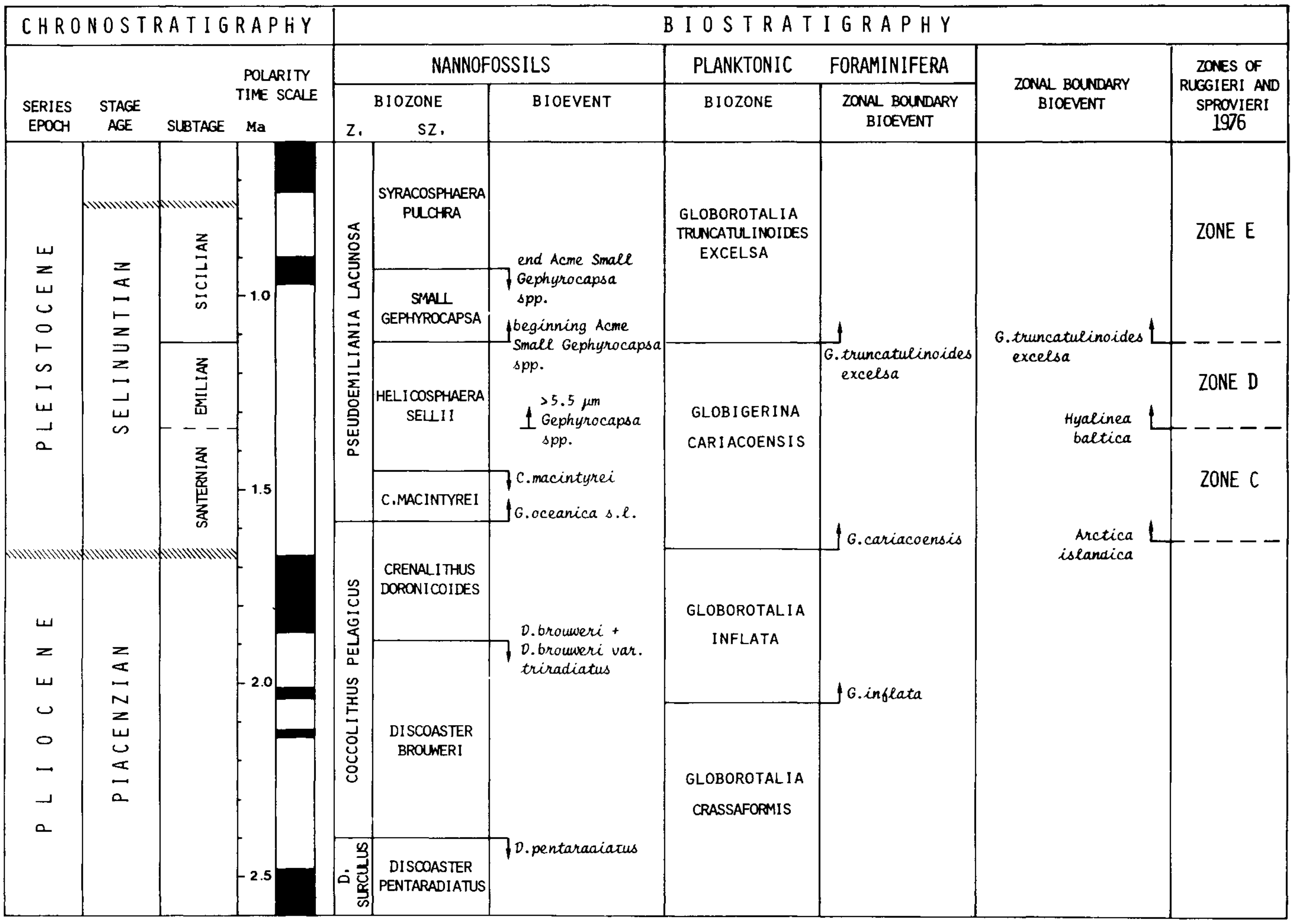

Figure 11.2. Chronostratigraphy and biostratigraphy adopted in this chapter.

Pleistocene portion of the Santerno River section was presented in several papers by Nakagawa and co-workers; see Nakagawa et al. (1975) and the sources cited therein. Kukla, Collins, and Bender (1979) proposed a reinterpretation of those data, but unfortunately the two teams had measured the total thickness of the section differently, and both sets of measurements differed from those reported by Cremonini et al. (1969), to which our biostratigraphic data are referred.

In Figure 11.9 we show the biostratigraphic data as interpolated between fixed points in the different versions of the log of the section (i.e., the first local occurrence of *A. islandica*, and the chronostratigraphic boundaries). Although this implies that the positions of the biostratigraphic events are rather approximate, it does not affect our discussion of the magnetostratigraphic interpretations.

The previous interpretations proposed by Nakagawa et al. (1975) and by Kukla et al. (1979) are compared in Figure 11.9. Apart from thickness problems, there are other discrepancies. For instance, Kukla et al. (1979) do not show the many short, normal-polarity events recorded by Nakagawa et al. (1975) in the reversed interval below SAN 1 and above SAN 2. This may, of course, be due to better cleaning to remove outcrop remagnetization, or it may indicate overlooked data.

Correlation of the paleomagnetism of this section to the standard paleomagnetic time scale is particularly difficult, as testified by conflicting opinions among workers (cf. Nakagawa et al., 1975; Kukla et al., 1979; Arias et al., 1980). Though we are unable to present here a definitive interpretation, we shall discuss the published interpretations in the light of the constraints suggested by biostratigraphy.

To begin with, the upper normal-polarity magnetozone, named SAN 1 by Kukla et al. (1979), was correlated to the Brunhes chron by Nakagawa et al. (1975) and to the Olduvai subchron by Kukla et al. (1979). Biostratigraphic evidence indicates than neither of those interpretations is tenable.

At the top of the Santerno River section, a vertebrate fauna assigned to the late Villafranchian Farneta unit has been recovered (Azzaroli and Berzi, 1970). This faunal unit is older than 0.8 Ma (Arias et al., 1980), which conflicts with possible correlation of the normal magnetozone SAN 1 to Brunhes (less

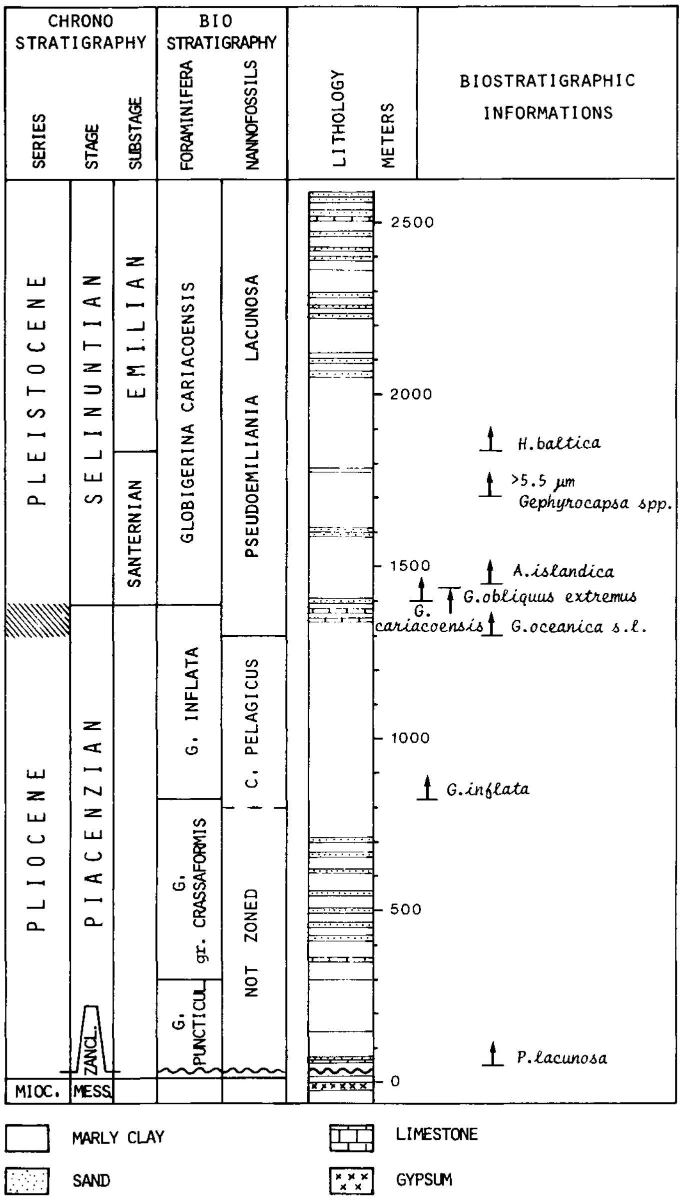

Figure 11.3. The Santerno River section.

than 0.78 Ma). On the other hand, below the SAN 1 magnetozone the successive appearances of *Gephyrocapsa oceanica s.l.* and "large" (more than 5.5 µm in diameter) *Gephyrocapsa* spp. have been detected by Raffi and Rio (1980a); both of those bioevents are known to have occurred in the late Matuyama, above the Olduvai and below the Jaramillo (Rio et al., Chapter 5, this volume). Taken together, these findings strongly conflict with the correlation of magnetozone SAN 1 to either the Olduvai or the Brunhes; the most probable correlation is, by default, the Jaramillo (ca. 1.0 Ma). The great thickness of the SAN 1 normal-polarity zone suggests the possibility, however, of secondary magnetization in this interval of the Santerno River section.

The biostratigraphic correlations reported in Figure 11.9 suggest a correlation between the SAN 2 magnetozone of Santerno River and the (upper) N3 magnetozone of the Vrica section. Above this level, a normal-polarity interval close to the first local appearance of *Arctica islandica* (Figure 11.9) was correlated to the Olduvai subchron by Nakagawa et al. (1975) and to the Réunion subchron by Kukla et al. (1979). Biostratigraphy indicates that both correlations are doubtful. In particular, we note that the first occurrence of *G. oceanica s.l.*, which has been observed above the top of the Olduvai (Rio et al., Chapter 5, this volume), is seen below the normal-polarity interval referred to in the Santerno River section. Furthermore, the first occurrence of *Globigerina cariacoensis* and the last occurrence of *Globigerinoides obliquus extremus*, also below the normal-polarity interval under discussion, are above the N1–N2 magnetozones of the Vrica section (Figure 11.9), which in turn have been definitively correlated to the Olduvai subchron (Tauxe et al., 1983; Pasini and Colalongo, Chapter 2, this volume).

Strangely, no long, normal-polarity interval correlatable to the Olduvai subchron has been identified below the first occurrence of *G. oceanica s.l.* in the Santerno section.

The lower part of the Santerno section, magnetozone SAN 3 of Kukla et al. (1979), shows normal paleomagnetic polarity and has been correlated to the Gauss chron. This correlation is in agreement with our biostratigraphic data. It is interesting that Nakagawa et al. (1975) reported a reverse-polarity interval at the base of the Santerno River section, which should be correlated to the uppermost Gilbert chron if the section is continuous. I. Raffi (unpublished data) recorded the first occurrence of *Pseudoemiliana lacunosa* a few meters above the base of the Pliocene in the Santerno River section, in the sampling of Colalongo et al. (1974); that event was calibrated to the top of the Gilbert by Rio (1982) and to the Gilbert–Gauss boundary by Poore et al. (1983).

West Santerno section. Colalongo et al. (1982b) studied this section in the watershed between the Santerno and Sellustra valleys. The main stratigraphic features are reported in Colalongo et al. (1982b). The Pliocene–Pleistocene boundary can be placed by correlation with the Pliocene–Pleistocene boundary-stratotype in the Vrica section by means of planktic foraminifera bioevents. In fact, the first occurrence of *Globigerina cariacoensis*, the last occurrence of *Globigerinoides obliquus extremus*, and the increase in abundance of left-coiling *Neogloboquadrina pachyderma* have been recorded close to each other and in the same relative order as in the Vrica section, where those events are close to the level adopted as the definition of the Pliocene–Pleistocene boundary.

Castell'Arquato section. In the Val d'Arda hills, near the towns of Vernasca, Lugagnano, and Castell'Arquato (Piacenza province), a thick sequence of Pliocene sediments is exposed that includes the Piacenzian (Upper Pliocene) stratotype. Pleistocene sediments crop out along the Riorzo and Bertacca streams west of Castell'Arquato. The complete section (Figure 11.4) has been

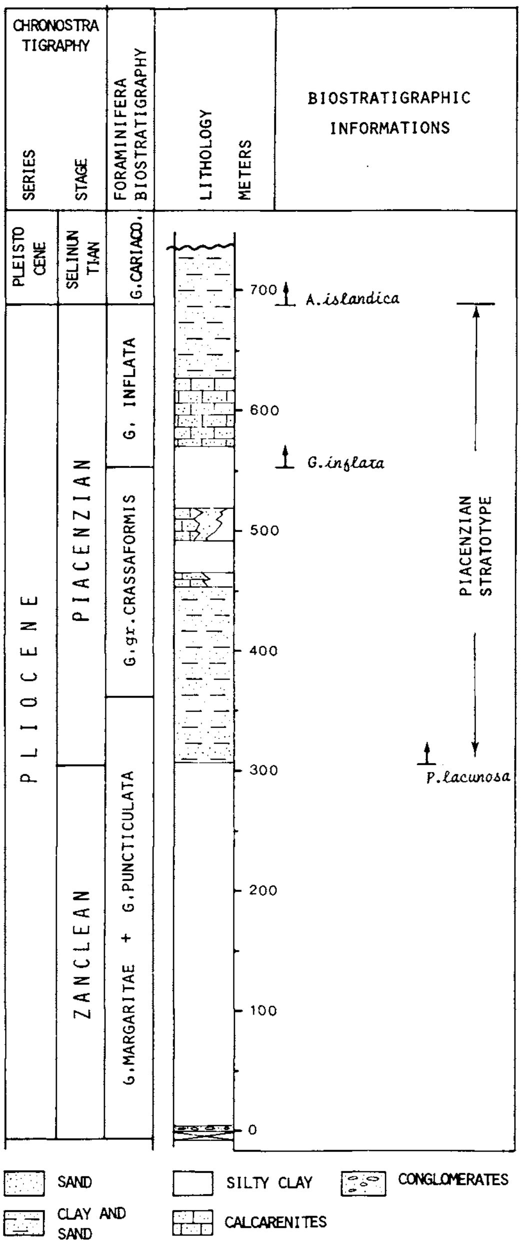

Figure 11.4. The Castell'Arquato section.

studied by several authors (Albertelli and Mazzei, 1963; Dondi, 1963; Barbieri, 1967, 1971; Colalongo et al., 1974; Rio et al., 1988).

The lithological sequence of the Castell'Arquato section, shown in stratigraphic order, is as follows:

sands (100 m)
calcarenites with interbedded silty clays (180 m)
clays and sands (140 m)
silty clays, with a basal conglomerate (ca. 320 m)

As a matter of fact, the Castell'Arquato section is a composite of different subsections: 9 according to Barbieri (1967), and 10 according to Colalongo et al. (1974). The correlations among these components have been locally difficult and may be incorrect.

Mollusca are very abundant in the outcrops on the left slope of the Arda valley (Monte Giogo, Monte Padova, Monte Falcone, Rio Cravalese), and corals, bryozoa, brachiopods, and echinoderms are also present. Benthic foraminifera are abundant throughout the whole section, while ostracodes and planktic foraminifera are much scarcer. Nannofossils are abundant, but mainly have been reworked (Barbieri and Rio, 1974; Rio et al., 1988). All things considered, it is difficult to obtain a detailed and reliable biostratigraphy based on calcareous microplankton. Different biostratigraphic zonations of the Castell'Arquato section have been published, based on the mollusks (Padovani and Tampieri, 1974), ostracodes (Colalongo and Russo, 1974), and benthic and planktic foraminifera (Barbieri, 1967; Colalongo et al., 1974). It is possible, nevertheless, that all those biostratigraphic efforts failed to identify hiatuses and correlation errors due to the desultory presence of the main index fossils (Colalongo et al., 1974).

According to Colalongo et al. (1974), the Pliocene–Pleistocene boundary in the Castell'Arquato section (i.e., the upper boundary of the Piacenzian Stage) is drawn in coincidence with the first occurrence of the "northern guest" *Arctica islandica*. As shown by Rio et al. (Chapter 5, this volume), this level seems to be at least roughly correlatable to the boundary-stratotype of the Pliocene–Pleistocene boundary in the Vrica section. A major unconformity, however, truncates the Piacenzian stratotype, below the shallow-marine deposits with Pleistocene fauna. According to Rio et al. (1994), this gap, which could be considered as an undefined Upper Pliocene chronostratigraphic interval, is represented by strata exposed at Monte San Nicola in southern Sicily that they have nominated as the stratotype of the new Gelasian Stage.

Central Italy

Vallebiaja section. Named for a farm on the hills southwest of Fauglia, Livorno district, this site is mentioned only because it is one of the typical sections of the Calabrian, according to Gignoux (1913, p. 300). Actually, Vallebiaja is not a section, but only a richly fossiliferous outcrop with a molluscan assemblage of Santernian age. As it happens, that particular outcrop no longer exists, and exposures in the area are so poor that the section may be considered completely unusable for biostratigraphic analysis.

Monte Mario section. The section at Monte Mario, on the right bank of the Trevere River in the city of Rome, is also one of the typical sections of the Calabrian, according to Gignoux (1913, p. 286). This section consisted of scattered exposures of fossiliferous sands belonging to the earliest Pleistocene, with the best outcrop at the locality called "La Farnesina." At the present time, the entire area has been built over because of the expansion of the city, and none of the old sites are accessible.

Musone section. The Musone is a small river near the village of Recanati, not far from Ancona. The section was synthesized by Selli (1954, p. 242), on the basis of a few outcrops, as a late Pliocene to early Pleistocene sequence. Even at the time of Selli's research, exposures in the area were generally poor, and in recent years several of the outcrops he described have been buried or destroyed.

Southern Italy

Monasterace section. This section crops out along the southern slope of Monte Rosito, a few kilometers north of Monasterace village on the Ionian side of Calabria, near Punta Stilo. It is noteworthy as one of the sections of the Calabrian described by Gignoux (1913, p. 48), but the local Plio–Pleistocene sequence is poorly exposed.

In the middle part of this section, where there are some scanty outcrops, about 20 m of blue marls can be sampled. From the presence of *Globorotalia inflata, Globorotalia truncatulinoides truncatulinoides, Globorotalia tosaensis,* and *Discoaster brouweri* these strata must be ascribed to the uppermost Pliocene. Greco, Ruggieri, and Sprovieri (1974) erroneously referred these marls to the Pleistocene because of the presence of *G. truncatulinoides,* which at that time was considered to have been restricted to the Quaternary.

Transgressively above the Upper Pliocene blue marls lies a sequence, about 100 m thick, that begins in a fossiliferous conglomerate with *Arctica islandica,* followed by interbedded sands and sandy clays, in which *G. truncatulinoides excelsa* is present. This transgressive sequence must be ascribed to a separate sedimentary cycle of late Emilian–Sicilian age (Figure 11.2), in a geologic and stratigraphic setting very similar to that in the Santa Maria di Catanzaro area (Sprovieri, D'Agostino, and Di Stefano, 1973). This being the case, we may conclude that the Monasterace section, even though indicated by Gignoux (1913) as one of the typical examples of the Calabrian, represents only the uppermost part of the Lower Pleistocene.

Capo Rossello section. One of the most spectacular Upper Miocene to Lower Pleistocene sections in the Mediterranean crops out at Capo Rossello, along the sea-cliff westward from Agrigento (Sicily). From bottom to top, the section (Figure 11.5) consists of white pelagic marly limestones (Trubi Formation), marls, marly clays, sandy clays (Monte Narbone Formation), and a thick turbiditic calcarenite bed (Agrigento Formation). In the classic Capo Rossello section, as described by Cita and Gartner (1973), there is a sedimentary hiatus at the Trubi–Monte Narbone transition. The missing part is well exposed nearby, however, at the locality of Punta Piccola (Rio et al., 1984a).

The sediments of this section, about 360 m thick, are bathyal in the lower part to neritic in the upper part and are very rich in microfossils, mainly foraminifera and calcareous nannofossils. The Capo Rossello section has been extensively studied in recent years by Cita and Gartner (1973), Cita and Decima (1975), Gartner (1977b), Sprovieri (1978), and Rio et al. (1984a), among others. Initially, Cita and Decima (1975) placed the Pliocene–Pleistocene boundary just below the base of a calcarenite marker where the lowest (resedimented) specimens of *Arctica islandica* occur in the section, at a level containing the first appearance of *Hyalinea baltica.* Subsequently, Sprovieri (1978) recognized the Pliocene–Pleistocene boundary by correlation with the Vrica section, on the basis of the first appearance of *Globigerina cariacoensis* and of the beginning of abundant left-coiling specimens of *Globoquadrino pachyderma,* about 80 m below the calcarenite bed. In addition, Raffi and Rio (1980a) and Rio et al. (1984a) noted the first occurrence of *Gephyrocapsa oceanica s.l.* about 65 m below that marker. In conclusion, the Capo Rossello

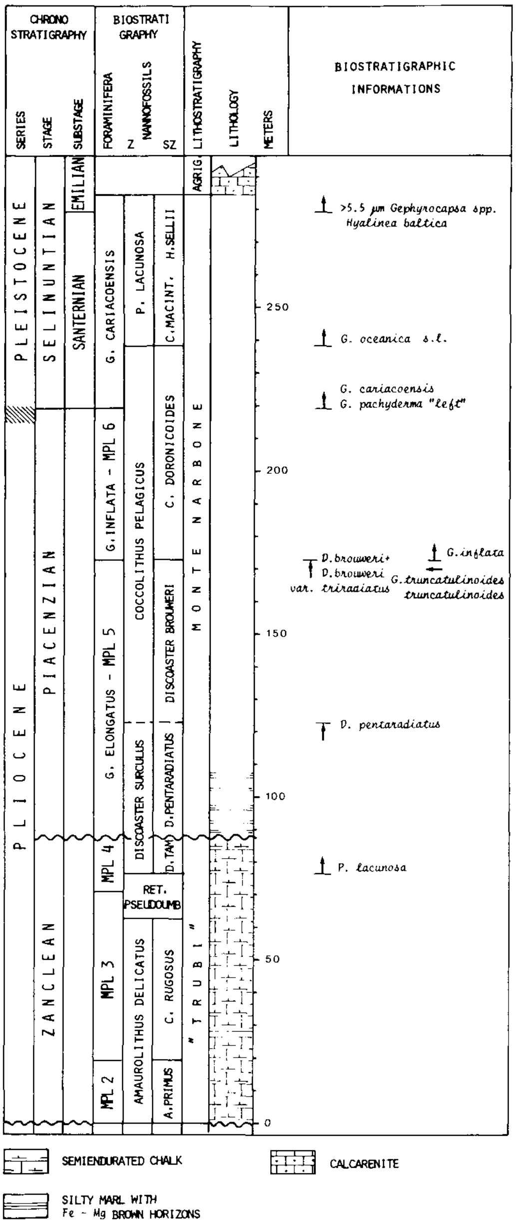

Figure 11.5. The Capo Rossello section.

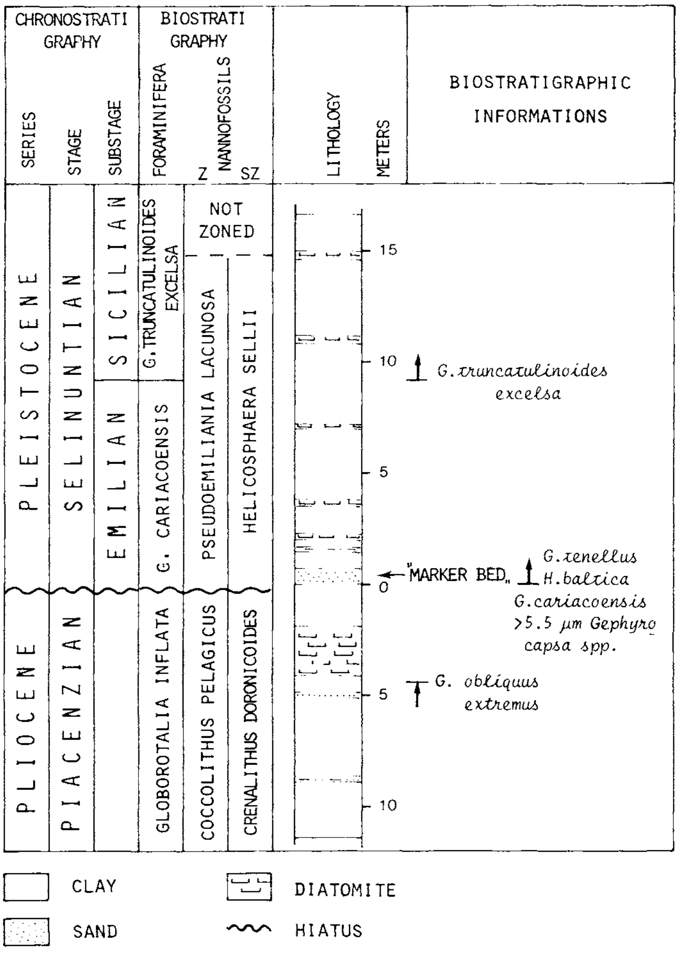

Figure 11.6. The Le Castella section.

section displays many very interesting features: (1) It represents an exceptionally long time interval, from the base of the Pliocene to the early Pleistocene, with continuous sedimentation in facies particularly suitable for paleontological studies. (2) It is directly correlatable with the Vrica section in the segment straddling the Pliocene–Pleistocene boundary. (3) According to preliminary studies, it seems to be suitable for paleomagnetic investigations.

Le Castella section. The Le Castella section (Figure 11.6) crops out in the Crotone sedimentary basin in south-central Calabria, about 20 km southwest of the city of Crotone. It is about 30 m thick, consisting of a clayey sequence, with some diatomaceous interbeds and two sandy beds, the thicker of which is well known in the literature as the "marker bed" (Emiliani, Mayeda, and Selli, 1961). Selli et al. (1977, p. 198) studied the part of the section "corresponding to the slope located near Telegrafo, comprising the 11 m under the sand marker bed and the 17 m above it. . . . As we get off this slope the exposure of the section becomes discontinuous, being frequently interrupted by several slumps and big blocks fallen from the overlying calcarenites; the area also presents some faults." For further discussion, see Colalongo et al. (1981, p. 105). Unfortunately, many other workers who have studied the Le Castella area have included those parts of the section, extending for several tens of meters below and above the "marker bed," that are distorted and duplicated by faults and outcrop slumps.

Earlier, the Le Castella section was selected by Emiliani et al. (1961) for studying the paleoclimatic variations near the Pliocene–Pleistocene boundary, located by those authors at the base of the "marker bed" and practically in coincidence with the first appearance in that section of the "cold guest" *Hyalinea baltica.* The Le Castella section was subsequently proposed (at the VII INQUA Congress, Denver, 1966) as a suitable section for representing the Pliocene–Pleistocene boundary-stratotype and was accepted as such by numerous authors (e.g., Rio, 1974; Haq, Berggren, and Van Couvering, 1977).

In recent years, however, the consensus has grown that the "marker bed" of Le Castella is unsuitable as a physical stratotype of the Pliocene–Pleistocene boundary (Pelosio, Raffi, and Rio, 1980; Colalongo et al., 1981, 1982b). The main reasons are (1) that the part of the section in the transition from Pliocene to Pleistocene is too short, (2) that the first local occurrence of *Hyalinea baltica*, which justifies the "marker bed" definition, is not the best criterion for recognizing the basal Pleistocene in Italy (Rio et al., Chapter 5, this volume), and (3) that biostratigraphic analyses (Raffi and Rio, 1980b; Colalongo et al., 1981) indicate the presence of a hiatus just below the "marker bed," as shown by a coalescence of biostratigraphic events that normally are widely spaced.

These findings indicate an Emilian age, probably late Emilian, for the "marker bed," in the chronostratigraphic subdivision of Ruggieri et al. (1984). In particular, the appearance of *Globorotalia truncatulinoides excelsa* 11 m above the "marker bed" (Colalongo et al., 1981) is indicative of the base of the Sicilian substage (Ruggieri and Sprovieri, 1977; Ruggieri et al., 1984).

Magnetostratigraphy of the Le Castella section. Nakagawa and co-workers (1975; Chapter 3, this volume) have reported on the magnetostratigraphy of a Le Castella section more than 300 m thick, sampled along a track where several long intervals are covered by debris and where faults have been recognized. As noted earlier, the Le Castella section is reliably continuous only over a short interval below and above the "marker bed" (Figure 11.6), and we therefore consider only the magnetostratigraphy of this interval (Figure 11.9). Biostratigraphic correlations (Figure 11.9) indicate that the normal-polarity interval (here named C2) below the "marker bed," in the interval above the last occurrence of *Discoaster brouweri* and below the first occurrence of *Gephyrocapsa oceanica s.l.*, can be correlated to the Olduvai subchron and to magnetozone N2–N3 of the Vrica section (Tauxe et al., 1983). The normal-polarity interval above the "marker bed" (here named C1) can be correlated to Vrica magnetozone N3 on biostratigraphic evidence, as shown in Figure 11.9. The short N3 magnetozone, as discussed by Tauxe et al. (1983) and Rio et al. (Chapter 5, this volume), cannot be

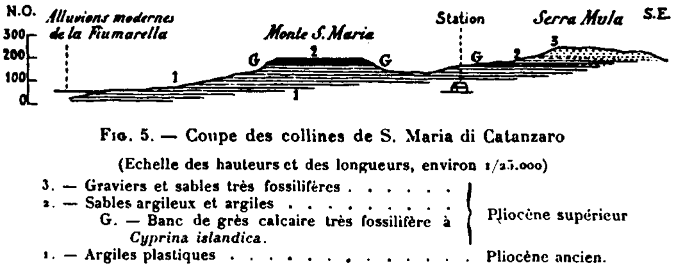

Figure 11.7. Stratigraphic cross section of the hills south of Santa Maria di Catanzaro, according to Gignoux (1913). In this drawing, the "Pliocène supérieur" is the equivalent of the Calabrian. (From Gignoux, 1913, p. 22.)

correlated to the Jaramillo subchron and might better be correlated to one of the short normal-polarity intervals recognized between the Olduvai and the Jaramillo by Cooke (1981), Mitsunashi et al. (1982), Nakagawa et al. (1982), and Suzuki and Manabe (1982).

Santa Maria di Catanzaro section. In the neighborhood of Santa Maria di Catanzaro, Calabria, the Plio–Pleistocene sequence is subdivided into a lower clayey unit (A) and an upper sandy unit (B). The upper unit includes a distinctive yellowish, macrofossil-rich calcarenite bed termed the G-G′ horizon by Gignoux (1913) (Figures 11.7 and 11.8).

Unit A is securely dated to the Pliocene. The "northern guests" *Arctica islandica* and *Hyalinea baltica* occur from the very base of unit B (Sprovieri et al., 1973), indicating that all of this unit should be ascribed to the Pleistocene (Figure 11.2). The two units have an unconformable contact described either as a transgressive surface (Sprovieri et al., 1973) or as a set of faults (Pasini, Selli, and Colalango, 1977). Unit A is not represented, however, in the classic "Santa Maria di Catanzaro section" of Gignoux (1913), which crops out along the road from the village of Santa Maria di Catanzaro up to the Colle Santa Maria (Pasini et al., 1977). The traverse crosses unit B and the G-G′ horizon, but nevertheless this section (Figure 11.7), like every other section in the vicinity, presents unfavorable features, including unexposed intervals, possible hiatuses, and tectonic disturbances. A synthesis is possible only through correlating a great many separate segments, with undue chances of error.

As noted, *Arctica islandica* and *Hyalinea baltica* are present from the base of the "classic" section. In that section, Ruggieri and Sprovieri (1977) reported that *Globorotalia truncatulinoides excelsa* occurs at least 50 m below the G-G′ bed, and on the basis of the biostratigraphic scheme in Figure 11.2, that leads to several conclusions. First, the whole "classic" Santa Maria di Catanzaro section is not only Pleistocene in age but also younger than the lowermost part of zone D of Ruggieri et al. (1976). This is confirmed by the presence, from the base of the section, of *Bolivinita quadrilatera* and *Coryphostoma karrerianum,* which in Italy always appear above the first appearance of *H. baltica.* Second, the G-G′ bed, which has been considered by most authors since Emiliani et al. (1961) to define the base of the Calabrian at Catanzaro according to the intentions of Gignoux (1913), is actually younger than the base of the Sicilian Stage, as marked by the first appearance of *G. truncatulinoides excelsa.*

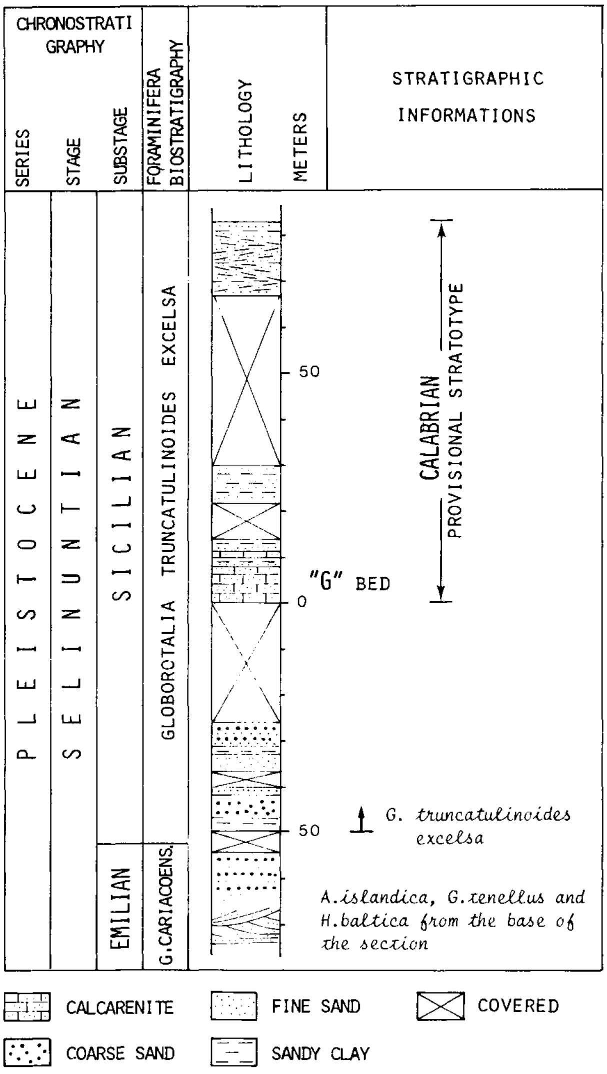

Figure 11.8. The Santa Maria di Catanzaro section.

Consequently, the Calabrian Stage, at least in the restricted sense of the Catanzaro "neostratotype" with the G-G′ bed at the base, as proposed by Selli (1971), is the same age as part of the Sicilian Stage, previously erected in 1872 by Doederlein. The term "Calabrian" under that definition is therefore equivalent to the earlier term for this chronostratigraphic interval in Italy and should be regarded as invalid. The question of replacing "Calabrian" as the name for the lowermost Pleistocene levels, which are not defined by the Santa Maria di Catanzaro neostratotype, or of amending the present definition is currently being debated (Rio, Sprovieri, and Thunell, 1991; Van Couvering, Preface, this volume).

Although Nakagawa and co-workers observed normal and reversed magnetozones in the Santa Maria di Catanzaro section, in work reviewed by Nakagawa (1981), Watkins et al. (1974) were unable to identify any variations in depositional polarity in samples from the same section, which appeared all normal. The latter authors demonstrated the presence of a precipitate phase, suggesting a secondary overprint, and as a result of this uncertainty we did not show magnetostratigraphy of this section in the correlation chart of Figure 11.9.

Vrica section. This section (Figure 11.9), about 4 km south of Crotone, is ideally suited for the Pliocene–Pleistocene boundary-stratotype and has been approved as such by the International Commission on Stratigraphy of the IUGS. The reader is referred to Aguirre and Pasini (1985) and to Chapters 2–5 in this volume for details.

Conclusion: the Italian marine record

Most of the Italian sequences reviewed in this chapter are unsuitable for defining the Pliocene–Pleistocene boundary-stratotype, for a variety of reasons: (1) because their exposures are of poor quality (Vallebiaja, Monte Mario, Musone, and Santa Maria di Catanzaro); (2) because they do not extend into the Pliocene (Vallebiaja and Monte Mario); (3) because they lack the lowermost Pleistocene (Monasterace, Le Castella, and Santa Maria di Catanzaro); (4) because they consist of several different component-sections of unreliable continuity and contain a poor planktic assemblage (Castell'Arquato). Only four of the reviewed sections (Vrica, Santerno River, West Santerno, and Capo Rossello) are sufficiently well exposed, fossiliferous, and continuous to be considered for the Pliocene–Pleistocene boundary-stratotype. The northern Italian Santerno sections, however, are highly terrigenous, and calcareous plankton (in particular, the nannofossils) are much less well represented than in the southern Italian sections.

Continental record

Villafranchian mammal stratigraphy and dispersal events

Detailed correlations of the typical Villafranchian faunas of Italy to European and Asian mammal faunas and to marine and paleomagnetic stratigraphy have been presented by Azzaroli (1977, 1983) and Azzaroli et al. (1982, 1986) (Figure 11.10). The Villafranchian has been subdivided into six faunal units, characterized from oldest to youngest by the sites at Triversa, Montopoli, Saint-Vallier, Olivola, Tasso, and Farneta, altogether representing the time from 3.0 Ma to approximately 1.0 Ma. Two major faunal turnovers – the "dispersal events" of Lindsay et al. (1980) – occurred at the beginning of the Montopoli unit and at the end of the Villafranchian. Less dramatic but still significant faunal changes distinguish other unit boundaries.

The Triversa unit, at the base of the Villafranchian, can be dated to the transition from early to middle Pliocene time, at the end of the *Globorotalia puncticulata* biozone in marine micropaleontology (De Giuli et al., 1983). Lindsay et al. (1980) estimated an age for this level at about 3 Ma, based on correlations to a short, normal-polarity paleomagnetic event between the Kaena and Mammoth reversed chrons in the Gauss epoch.

Montopoli event. The first major dispersal event took place during the early Villafranchian, between the Triversa and Montopoli faunal units. It was characterized by the disappearance of a warm-forest assemblage (*Mammut borsoni, Tapirus arvernensis, Ursus minimus, Sus minor*) and its replacement by an open-forest or bushland assemblage with the first appearances of *Equus, Gazella*, and *Mammuthus*. The Montopoli unit falls within the time interval of the *Globorotalia crassaformis* planktonic foraminifera zone (De Giuli et al., 1983). Lindsay et al. (1980) dated the Montopoli-type fauna to the beginning of the Matuyama paleomagnetic epoch, slightly younger than 2.6 Ma. That date finds support in a similar faunal assemblage at the Rincón 1 site, in the Júcar Basin of eastern Spain, which Alberdi et al. (1982) have dated to 2.5 Ma. The arrival of elephants in Italy, however, was somewhat earlier than in the type Montopoli fauna in the lower Valdarno of Tuscany, as evidenced by an occurrence in underlying strata at Laiatico.

The first dispersal event coincided with a marked cooling recorded in marine sequences at about 2.4 Ma, the most severe cooling of the Pliocene, in a series of steadily intensifying climate cycles leading up to the Pleistocene conditions (Thunell and Williams, 1983; Rio et al., 1991).

A second dispersal event, at the beginning of the Saint-Vallier unit, is less well marked. In Italy, that event is poorly represented because it coincided with the Aquatraversan phase of sea-level lowering (Ambrosetti et al., 1972), a time when the Italian peninsula was generally subjected to erosion. On the other hand, richer faunas have been described at Saint-Vallier and Chilhac in France and at La Puebla del Valverde in Spain. The Tegelen fauna of The Netherlands can also be ascribed to this unit, as can the older of the two faunas from the rich site of Senèze (Aguirre et al., Chapter 9, this volume). The transition from the Montopoli to the Saint-Vallier faunal unit is not sharp, but is marked by the replacements of *Dicerorhinus jeanvireti, Mammuthus gromovi*, and *Equus livenzovensis* by *Dicerorhinus etruscus*, primitive *Mammuthus meridionalis*, and *Equus stenonis*, respectively, in what may have been evolutionary successions. Other characteristic forms in the new fauna were *Cervus rhenanus* Dubois, 1905 (= *C. philisi* Schaub, 1941), *Eucladoceros teguliensis* (Dubois), 1905 (= *E. senezensis* Depéret and Mayet, 1910), and sporadic early appearances of *Sus strozzii*. The Saint-Vallier unit also contains the final forms of *Gazella*, the cervid *Croizetoceros*, and the canid *Nyctereutes* in European faunas. The absence of *Hipparion*, last known from Étouaires in the later Montopoli fauna, is also notable.

In the Tegelen clays of The Netherlands, pollen complexes indicating a generally warm and equable climate, with low-amplitude climatic fluctuations, span the early Matuyama from

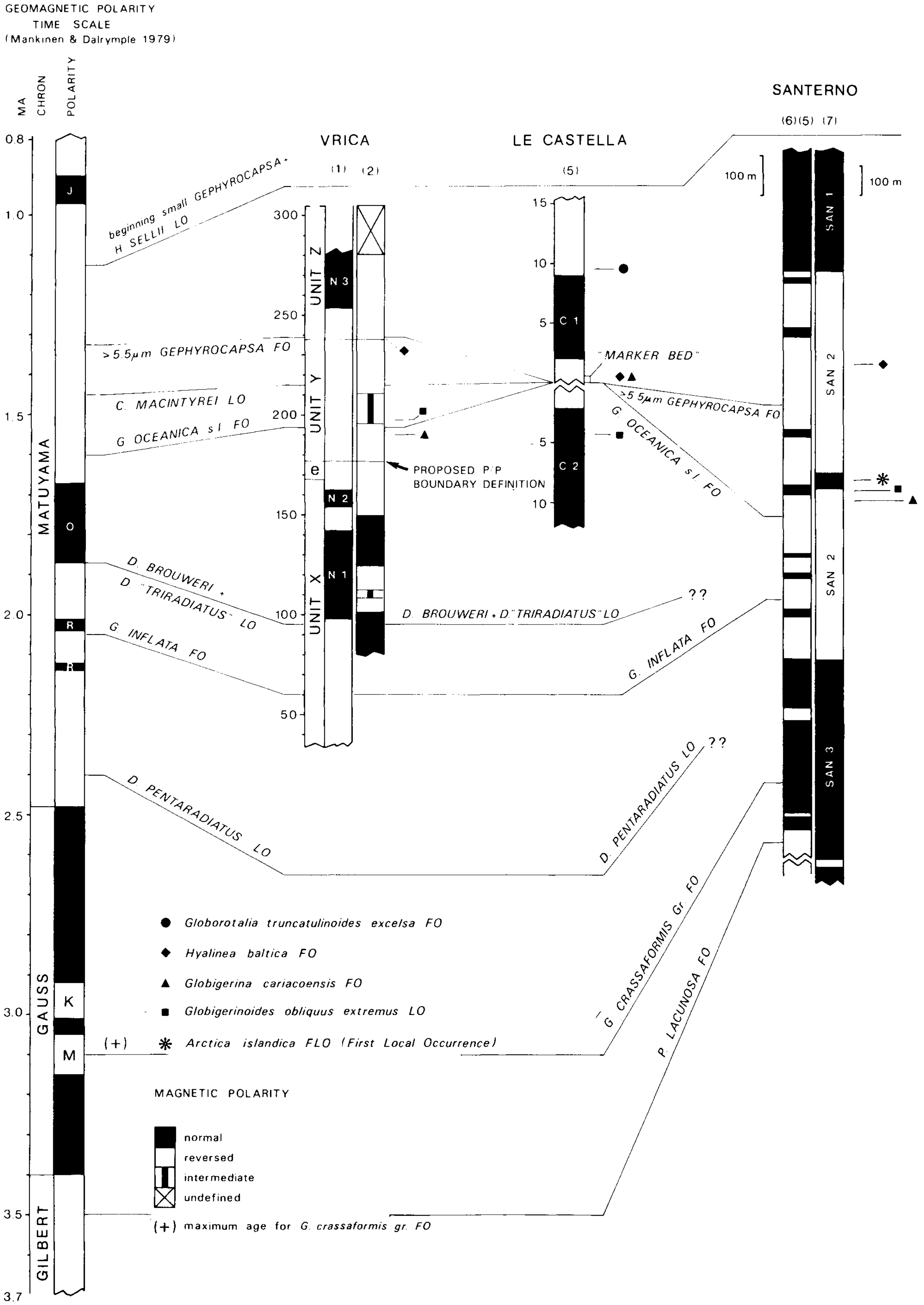

Figure 11.9. Biostratigraphic correlations for the Vrica section, the Le Castella section, and the Santerno section. FO, first occurrence; LO, last occurrence. The paleomagnetic logs are as follows: 1, Tauxe et al. (1983); 2, Nakagawa et al. (Chapter 3, this volume); 5, Nakagawa et al. (1977); 6, Nakagawa et al. (1975); 7, Kukla et al. (1979).

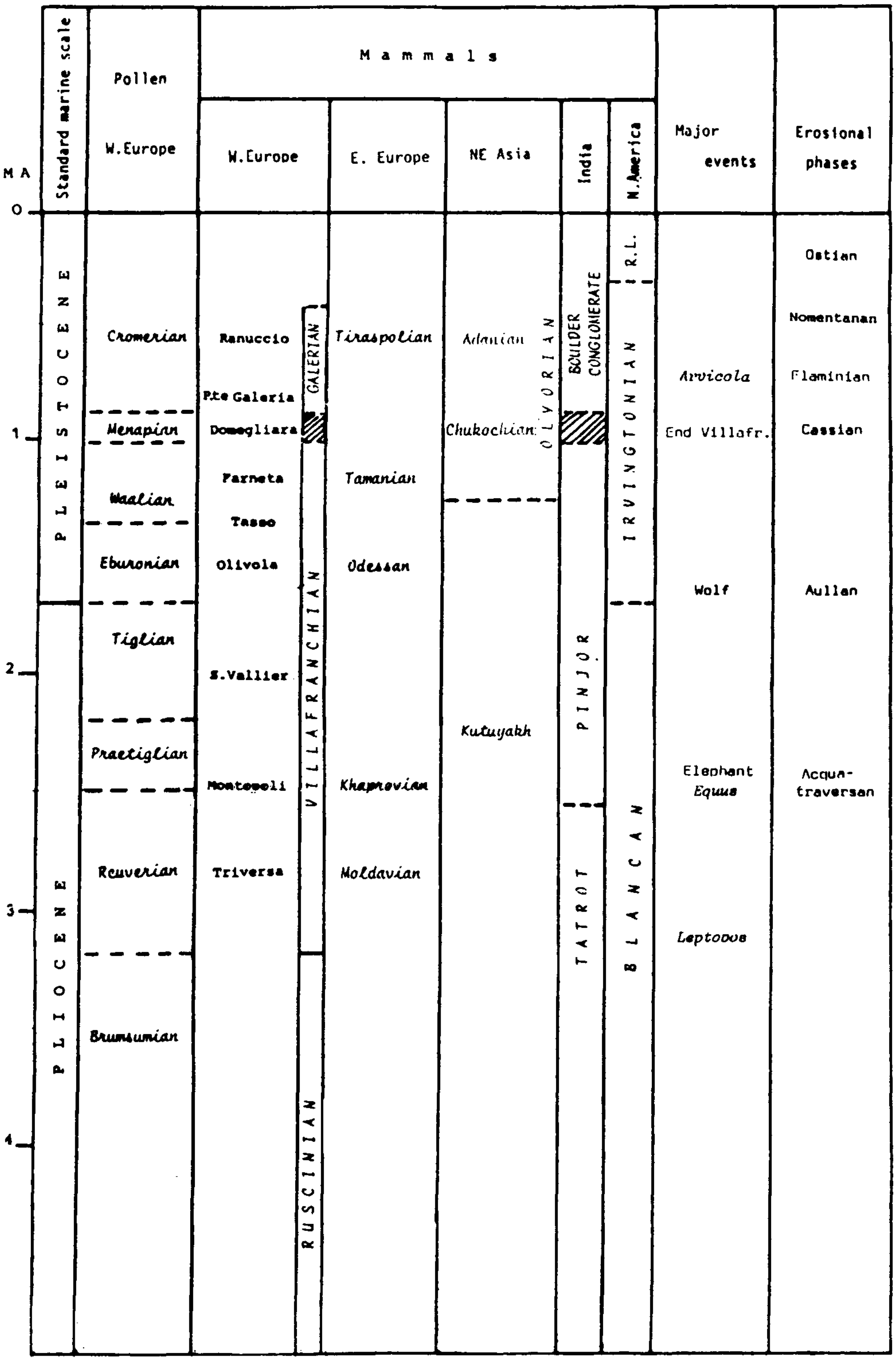

Figure 11.10. Chronostratigraphy of mammalian faunas, paleofloras, and physiographic events in Eurasia (uncorrected time scale).

about 2.2 Ma to the top of the Olduvai normal at 1.8 Ma (Suc and Zagwijn, 1983; Zagwijn, Chapter 16, this volume). The maar deposits at Senèze have a pollen spectrum with similar features and also contain a mammal fauna that shares many characteristic elements with the Saint-Vallier unit (e.g., *Nyctereutes megamastoides, Eucladoceros teguliensis, Cervus rhenanus*, and *Croizetoceros ramosus*). The Senèze fauna is generally considered to be younger than the fauna at Saint-Vallier (Méon et al., 1979), and in fact it includes more progressive taxa, which nevertheless are older than the comparative forms in the next (Olivola) unit. A date between 2.0 and 1.8 Ma is the only plausible age for the fauna of the Senèze maar deposits. It must be noted, again, that Bout (1970) reported fossils from slopes above the maar deposit, which accounts for the mixture of two distinct paleofaunas at this site.

Olivola event. The transition to the Olivola unit is more sharply marked, in a changeover termed the "wolf event" by Azzaroli (1983). This is characterized by the appearance of the "wolf" *Canis etruscus*, as well as *Pachycrocuta brevirostris, Leptobos etruscus*, and cervids such as *Dama nestii* and *Eucladoceros dicranios* (including *E. ctenoides*), the latter two as evolutionary successors to *Cervus rhenanus* and *Eucladoceros teguliensis*, respectively. Scattered occurrences of the mastodont *Anancus arvernensis* in the beds of the Montevarchi Group in the Upper Valdarno sequence provide evidence that this typically Pliocene species survived for a short time after the end of the Saint-Vallier time. The type specimen of *Mimomys pliocaenicus* (Major), 1889, comes from this unit in the Upper Valdarno or (less probably) from the overlying Tasso unit.

There has been no calibration of the Olivola in Tuscany itself, despite several attempts to secure paleomagnetic or pollen data, but sediments that yield the Olivola-type fauna in northwestern Tuscany are high-energy deposits indicative of cold, wet climatic conditions and lowered sea level in the Aullan erosional phase (Arias et al., 1980). The Aullan, in all likelihood, was coincidental with the Eburonian cold phase documented in The Netherlands (Zagwijn, Chapter 16, this volume), which succeeded the Tiglian warm climate at the end of the Olduvai normal-polarity chron. The beginning of the Olivola faunal unit, which in classic terms marks the beginning of the Middle Villafranchian (Azzaroli, 1977), thus closely approaches the Pliocene–Pleistocene boundary defined by the marine sequence at Vrica. De Giuli et al. (1983) proposed a correlation of the Olivola unit with the *Globigerina cariacoensis* biozone.

Tasso and Farneta events. Succeeding units of the later Villafranchian are less sharply defined. The next younger Tasso unit is characterized by the following immigration events: *Canis arnensis*, a jackal or coyote; *Canis falconeri*, possibly a primitive "hunting dog" or lycaonid; *"Leptobos" vallisarni*, a somewhat aberrant species of this genus with a massive skull; a poorly known ovibovine (Borselli et al., 1980); and *Hippopotamus antiquus*. The caballine *Equus stehlini* in that fauna may have evolved *in situ* from *E. stenonis*.

The Farneta unit is even less well known. A distinguishing element is the highly derived elephant, *Mammuthus meridionalis vestinus*, in Italy, which has its close relatives *M. m. tamanensis* in southern Russia and *M. m. cromerensis* in the East Anglian faunas of Great Britain. The local fauna at Imola, in northern Italy, is characterized by progressive subspecies of *Mammuthus* and *Eucladoceros* associated with *Hippopotamus*.

There are two similar faunas in the Massif Central of France, one at the Creux de Peyrolles, near Perrier, with *Eucladoceros tetraceros, Dicerorhinus etruscus*, and the still poorly known *Cervus perolensis* (Bout and Azzaroli, 1953). The other French Farneta-like fauna is the younger fauna at Senèze noted earlier, with *Cervalces gallicus, Equus bressanus, Canis arnensis*, and possibly *Equus stehlini. Megalovis latifrons* may also belong to this level, rather than to the older fauna in the maar deposits.

Fairly rich Upper Villafranchian faunas occur at several sites in the Forest Bed series of East Anglia, beneath the Cromer Forest Bed. Those sites have been divided into Pastonian and pre-Pastonian, but unfortunately it is impossible in virtually all instances to know from which level the collected material was recovered (Stuart, 1982). At East Runton, the Forest Bed fauna occurs alone, but at Bacton, Sidestrand, Overstrand, Mundlesey, and possibly Pakefield the fossils are found together with younger Cromerian (or, better, Galerian) material (Azzaroli, 1953; Aguirre et al., Chapter 9, this volume). *Dama nestii, Eucladoceros dicranios, Equus stenonis*, and *Mimomys pliocaenicus* indicate a general late Villafranchian age, while *Mammuthus meridionalis cromerensis, Eucladoceros tetraceros, Cervalces gallicus*, and *Equus bressanus* suggest latest Villafranchian (i.e., Tasso or Farneta) levels.

End-Villafranchian event. The most dramatic dispersal event occurred at the end of the Villafranchian, with the transition to the Galerian faunas of the early middle Pleistocene (Ambrosetti et al., 1972). That event, which affected not only Italy but all of Eurasia (Aguirre et al., Chapter 9, this volume), was the "end-Villafranchian event" of Azzaroli (1983). The faunal list compiled by Azzaroli et al. (1982) for Italy alone lists 17 extinctions and 17 new taxa at the beginning of the Galerian, with 4 lineages that survived the transition undergoing marked evolutionary changes, and only 4 that survived into the Galerian, if only for a brief time, with little or no change.

The extent of the end-Villafranchian turnover is clearly apparent from these statistics, but from these changes other changes are also apparent. The fact most deserving of emphasis is that the end-Villafranchian event was marked by new types of adaptations unknown in earlier faunas. The development of hypselodont (ever-growing) teeth in the microtine rodents *Microtus, Arvicola*, and *Pitymys* may have been no more than the natural outcome of a trend that had already begun in the Pliocene, but among the ungulates the appearance of giant herbivores, such as the megacerid *Cervalces latifrons* and heavy-bodied species of *Bison, Bos, Praeovibos*, and *Ovibos*, represents, in totality, an entirely new response to the environment.

The end-Villafranchian turnover was not instantaneous, but it took place within a relatively short time span that in Europe seems to have been confined to about 70 k.y., within or just above the Jaramillo subchron, between 1.0 and 0.9 Ma. It was accompanied by major changes in the vegetation of the region from Europe to Japan (Grichuk, Chapter 8, this volume) and great increases in loess deposition in central Asia and the Danube and Rhine basins. A major shift in stable-isotope values in marine environments of this age indicates a sharp cooling at approximately the same time (Shackleton and Opdyke, 1976; Thunell and Williams, 1983; Shackleton et al., 1995), together with sea-level lowering and erosion evidenced in the Cassian event in the Tiber delta (Ambrosetti et al., 1972; Azzaroli, 1983).

Summary of the continental record

In conclusion, the Villafranchian and Galerian continental faunas in Italy show evidence of two major turnovers, the elephant-*Equus* event at about 2.5 Ma and the end-Villafranchian event at about 0.9 Ma, both of which were approximately coeval with major climate and ocean-level minima. The Pliocene–Pleistocene boundary, coincident with the beginning of the Olivola unit in the Valdarno sequence, was also marked by significant changes in the land-mammal faunas, but those changes were not accompanied by changes in the environment comparable to the changes that accompanied the two major events.

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12 Stratigraphy of the Plio–Pleistocene sequence of the Mediterranean coastal belt of Israel and its implications for the evolution of the Nile Cone

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Introduction

The Plio–Pleistocene sedimentary sequence of the coastal plain and the continental shelf of Israel is an integral part of the Nile Cone, which has been building up since the end of the Messinian Event. The Yafo Formation of the Saqiye Group (Pliocene to early Pleistocene) and the Hefer Formation of the Kurkar Group (Pleistocene), which constitute this sequence, are subdivided into 15 correlatable members. Nine biostratigraphic datum levels have been established within the sequence, which permits correlations to the Plio–Pleistocene events in the central Mediterranean and Italian sequences.

Background

Our compilation of the marine Plio–Pleistocene sequence of the coastal belt of Israel is based mostly on our previous studies and on additional examinations of new boreholes. The lithostratigraphy and mapping, including correlations of wells and outcrops, interpretations of seismic surveys, subcrop mapping, and sedimentology, were published by Gvirtzman (1969a,b, 1970, 1983) and Gvirtzman and Buchbinder (1969). Other publications include studies of biostratigraphy and molluscan correlation by Moshkovitz (1961, 1963, 1968), studies of nannofossil biostratigraphy by Ehrlich and Moshkovitz (Ehrlich and Moshkovitz, 1978; Moshkovitz and Ehrlich, 1980), studies of foraminiferal biostratigraphy by Martinotti (1981a–c, 1986), and studies of combined biostratigraphy of foraminifera and nannofossils by Moshkovitz and Martinotti (1979). In addition, compilations and regional syntheses were prepared by Gvirtzman and Buchbinder (1977, 1978) and by Gvirtzman et al. (1984).

The Nile Cone

The Plio–Pleistocene sequence is an integral part of the Nile Cone (Figure 12.1). The eastern lobe of the cone is formed by sediments which are transported by the Nile River and are carried by anticlockwise longshore currents into the southeastern corner of the Mediterranean (Emery and Bentor, 1960; Emery and Neev, 1960; Neev et al., 1976; Nir, 1984). This lobe is distributed over the coastal plain, the continental shelf and the continental slope of Israel.

The present Nile Delta and Nile Cone began forming at the end of the Miocene, when the Mediterranean sea level was restored and sediment from the Nile began to backfill the deeply dissected canyon that had been incised to the Messinian (latest Miocene) base level. Thus, the sediments of the delta and cone overlie Messinian evaporites in most of the area. The sedimentary sequence of the delta and cone is composed of three successive complexes (Rizzini et al., 1978; Gvirtzman and Buchbinder, 1978; Said, 1981). In the deposits of the Israel coastal plain, these are recognized as the basal post-evaporite shallow-water Afiq Formation, the deep-water Yafo Formation, and the overlying shallow-water Hefer Formation (Figure 12.2). The Afiq is composed of coarse clastics that fill erosional channels, and it does not form a continuous sheet. This complex is found only at the southern margin of the area, near the continent. The middle complex, or Yafo, forms most of the volume of the delta and the cone and is distributed throughout the entire area. This complex includes open-sea and hemipelagic clays and silts in foreset beds. The Hefer strata at the top are composed of coarse, mainly sandy clastics distributed (in the same manner as the Afiq) only in the southern coastal plain and near the continent.

Four lithologic components make up the sediments (Gvirtzman, 1970):

1. Clays (smectite, kaolinite, and illite) and quartz particles, most of which were transported by the Nile River. This component accounts for most of the volume of the delta and the cone.
2. Biogenic detritus formed in the marine environment, including microfossil oozes (planktonic foraminifera and nannofossils) and shells of benthic foraminifera, siliceous microfossils, and mollusks.
3. Eolian dust from the Sahara and Sinai deserts.
4. Coarse detritus, including conglomerates, transported by small rivers other than the Nile.

Features such as salt tectonics, salt flows, and diapirs are common in the continental slope in areas where thick Messinian

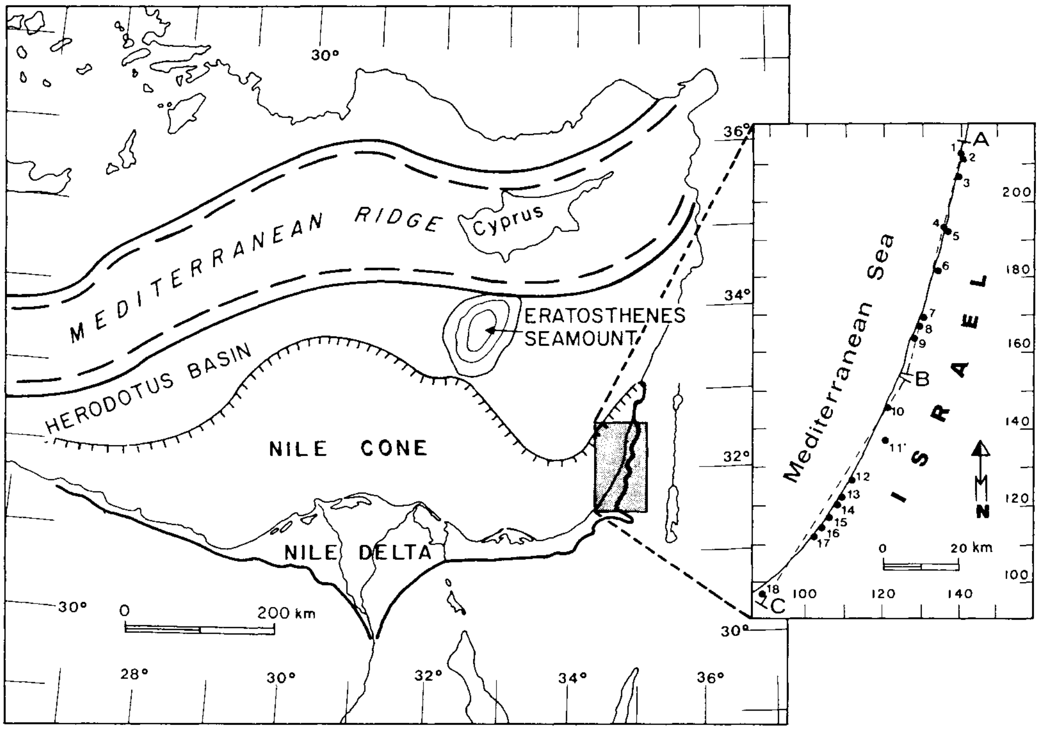


Figure 12.1. Physiographic elements in the southeastern Mediterranean (Ross and Uchupi, 1977; Gvirtzman and Buchbinder, 1978; Said, 1981) and location map of the studied area. Section A–B–C is illustrated in Figures 12.3 and 12.5. Data on boreholes 1–18 are given in the Appendix.

evaporites underlie the Nile Cone. These features cause huge growth faults, gigantic rotational slump structures, and collapse structures in the clayey sediments of the cone (Neev et al., 1976; Ross and Uchupi, 1977; Almagor and Garfunkel, 1979).

Gravitites (gravity-induced sediments) and suspensites (hemipelagic-sapropelic sediments) are being deposited today, as during the late Quaternary, over most of the cone area (Maldonado and Stanley, 1978). After a study of clay minerals from those sediments, Maldonado and Stanley (1981) concluded that the smectite, the dominant clay mineral, is deposited by the mass flow of water from the Nile River, whereas the kaolinite is windblown, originating in North Africa, and the illite and chlorite have a northwestern provenance. It is highly probable that sediments of the same type and composition have been deposited during the entire history of the Nile Delta and Cone.

Lithostratigraphy

The findings from our study are presented in stratigraphic sections of 18 representative boreholes or wells selected along a line following the present shoreline of the Mediterranean and parallel to the depositional strike (Figure 12.1). An up-to-date lithostratigraphic subdivision of these wells, together with distribution charts for selected index fossils, has been prepared from the borehole data shown in the Appendix. The new lithostratigraphic nomenclature is illustrated in Figure 12.2; detailed correlations of the upper part of the sequence, including the occurrences of the most significant fossils, are shown in Figure 12.3. The sequence from the Jaffa 1 borehole, which constitutes the reference section for the Plio–Pleistocene in Israel (Gvirtzman, 1970; Gvirtzman et al., 1984), is illustrated in Figure 12.4, and a regional correlation chart of the basin is shown in Figure 12.5.

Two stratal groups are included in the Plio–Pleistocene sequence: the Saqiye and the Kurkar. The Saqiye Group (Gvirtzman and Reiss, 1965; Gvirtzman, 1970) is dated from the late Eocene to the early Pleistocene and is composed mostly of open sea marls and clays. Its upper part (Figure 12.2), in ascending order, includes the Mavqiim Formation (Gvirtzman, 1970), composed of anhydrite, gypsum, and salt of late Miocene (Messinian) age, the Afiq Formation (Gvirtzman, 1970; Gvirtzman and Buchbinder, 1978), composed of conglomerates, variegated shales, sandstones, siltstones, and marls of latest Messinian (latest Miocene) age, and the Yafo Formation (Gvirtzman and Reiss, 1965), composed of marine marls and clays of Pliocene and early Pleistocene age. The Kurkar Group (Gvirtzman, 1969b; Gvirtzman et al., 1984), which ranges from early Pleistocene to Holocene in age (Figure 12.2), is composed of calcareous sandstone (known by the local name of “kurkar”), reddish clayey-silty sandstones (known by the local name of “hamra”), siltstones, marls, conglomerates, dark swamp claystones, and unconsolidated dune sands. In the west, near the present shoreline, this group is represented by the Hefer Formation (Gvirtzman et al., 1984), composed of marine calcareous sandstones, eolianites, hamras, siltstones, and marls. To the east, the Kurkar Group in the coastal plain is represented, in ascending order, by the Pleshet Formation (Issar, 1961, as emended by Gvirtzman and Buchbinder, 1969) composed of marine calcareous sandstones, by the Ahuzam Formation (Issar,

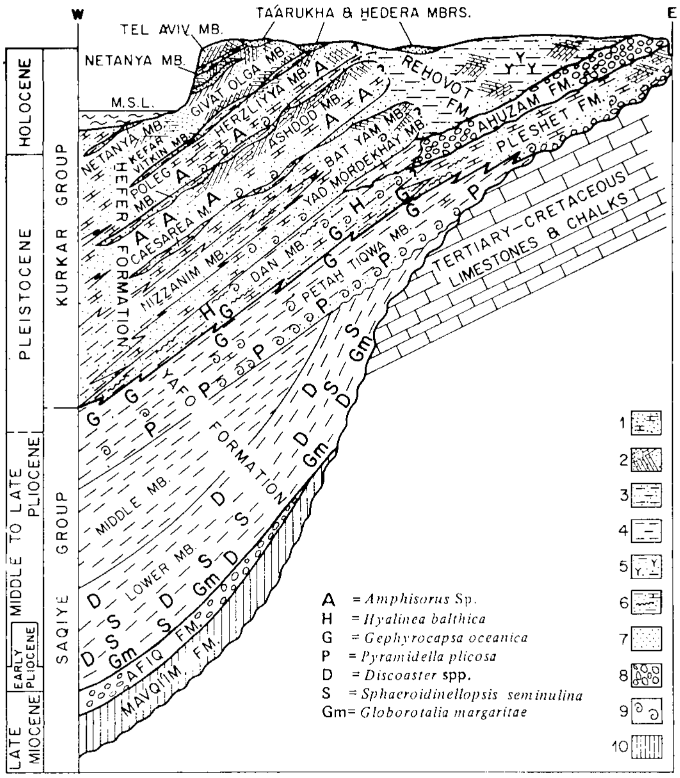


Figure 12.2. Stratigraphic nomenclature of the Plio–Pleistocene sequence of the coastal plain of Israel (modified from Gvirtzman et al., 1984). Lithology: 1, marine calcareous sandstones (kurkar); 2, eolianites; 3, clayey-silty reddish sandstones (hamra); 4, marine shale; 5, dark clayey swamp deposits with plant remains; 6, marine silts, clays, and calcareous sandstones; 7, loose dune sands; 8, conglomerates; 9, coquina beds; 10, anhydrites. Biostratigraphic markers: A, *Amphisorus* sp.; H, *Hyalinea baltica*; G, *Gephyrocapsa oceanica*; P, *Pyramidella plicosa*; D, *Discoaster* spp.; S, *Sphaeroidinellopsis seminulina*; Gm, *Globorotalia margaritae*.

1961) composed of fluvial conglomerates, and by the Rehovot Formation (Issar, 1961) composed of hamras, eolianites, and dark swamp clays. The Hefer Formation is a lateral equivalent of these three formations. The main difference between the Saqiye and the Kurkar groups is that the Saqiye is composed mainly of marls, clays, and silts, whereas the Kurkar is made up mainly of sandstones.

Yafo Formation

The Yafo Formation varies in thickness from a few tens of meters in the east to about 2,000 m in the continental shelf. The type section of the formation is in the Jaffa 1 borehole in the depth interval between 217 and 1,212 m (Figure 12.4). The marls of the formation are composed of smectite, kaolinite, and illite (20–70%) and of silt-size and fine-sand-size quartz particles (5–10%), most of which are derived from the Nile River (Gvirtzman, 1970). The marls also include biogenic calcite (25–75%), mainly derived from microfossils and nannofossils, some of which are recycled Cretaceous and Tertiary fossils. Seismic records from the western coastal plain and from the continental shelf show huge growth faults and rotational slump structures in the Yafo Formation (Almagor and Garfunkel, 1979). The formation is subdivided into three members.

Lower Member. The Yafo Lower Member, introduced herein, has its type section in the Jaffa 1 borehole in the depth interval between 650 and 1,212 m (Figures 12.2 and 12.5). It is composed of relatively carbonate-rich marls, with abundant planktonic foraminifera and calcareous nannofossils. Its thickness changes from about 100 m at the margin of the basin to about 700 m at its center (Figure 12.5). The member overlies the Mavqiim and the Afiq formations and in some places rests on a truncated surface of older formations. The top of the Lower Member is eroded along the margin of the basin. In the center of the basin, the member is overlain by the Middle Member of the Yafo Formation. The boundary between the Lower and the Middle members is somewhat transitional, and therefore it is not always well defined.

Middle Member. This member, also introduced herein, has its type section in the Jaffa 1 borehole in the depth interval between 310 and 650 m. It is relatively rich in clays and marls and low in carbonates, and while poor in planktonics, redeposited nannofossils are abundant (Moshkovitz and Ehrlich, 1980). Its thickness varies from zero at the margin of the basin to about 500 m in the center. The member conformably overlies the Lower Member and is overlain by the Petah Tiqwa Member, the latter horizon being marked by a strong seismic reflector just above the boundary (remark no. 25 in the Appendix). From seismic records it is evident that most of the growth faults and the slump structures are confined within the Middle Member. The seismic reflector near the base of the Petah Tiqwa Member appears to be a smoothed surface on the distorted blocks and rotated slabs of the Middle Member. Some of the faults, however, also displace the overlying Petah Tiqwa Member.

Petah Tiqwa Member. The Petah Tiqwa unit of Gill (1965) is the upper member of the Yafo Formation, with its type section in the Jaffa 1 borehole, in the depth interval between 217 and 310 m, and is composed of carbonate-rich marls. In most of the area, three distinct coquina beds composed of fine sands and carbonate-rich marls alternate with clayey beds, with a distinct signature that can be recognized clearly in the electric logs throughout the entire basin. The lowermost coquina bed is rich in loose, well-preserved mollusks. The thickness of the member varies from about 10 m to about 100 m.

Five electric-log markers in the Petah Tiqwa, known in ascending order as BPT, PT3, PT2, PT1, and A, can be followed throughout most of the area. These electric markers (Figures 12.3 and 12.4) have been used for precise correlation. With the Petah Tiqwa Member a regional shallowing of the entire basin is introduced. Minor sands presage the shift to dominantly sandy lithologies in the overlying members of the sequence.

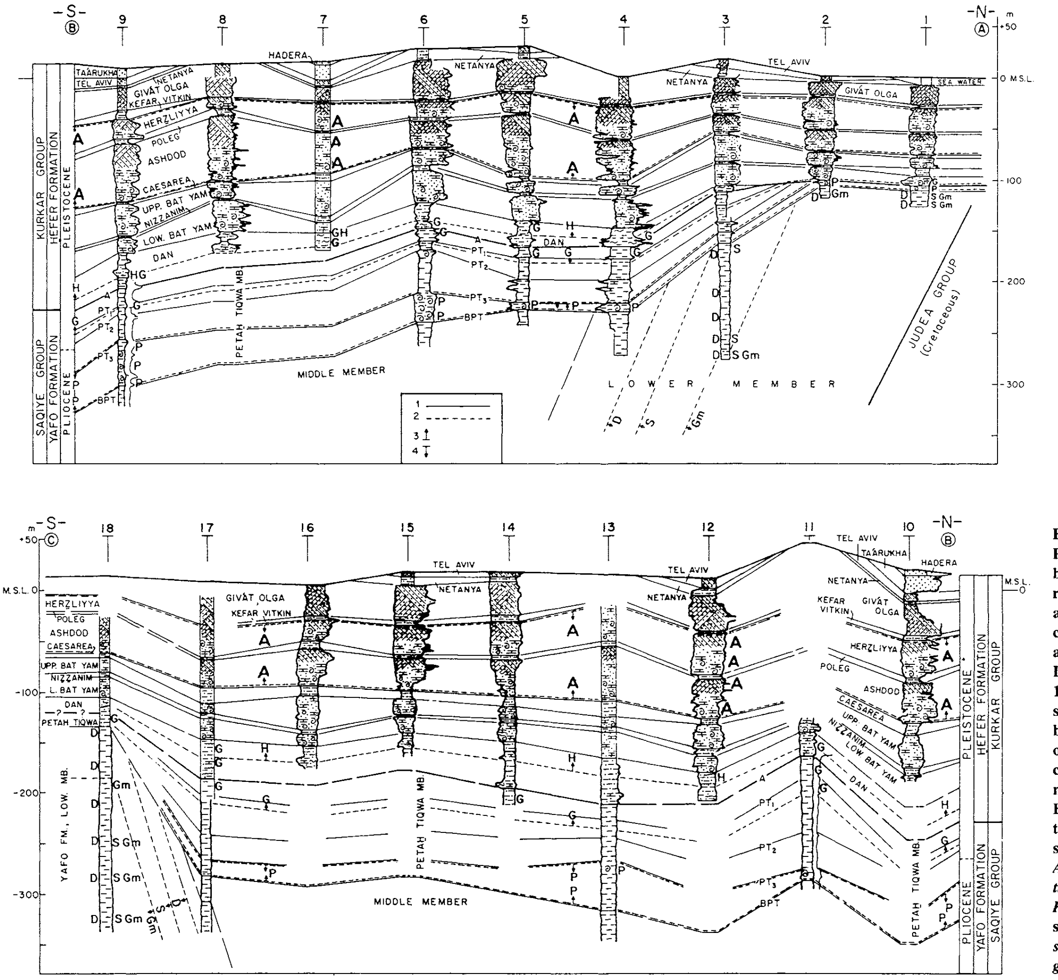
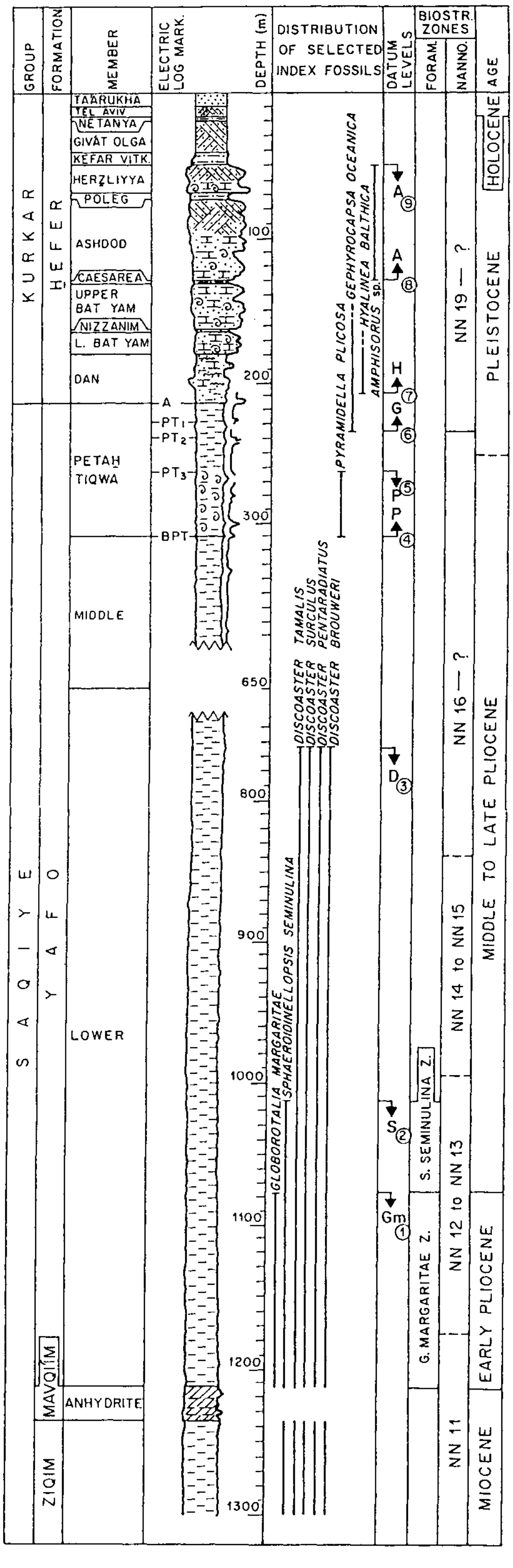


Figure 12.3. Upper Pliocene to Pleistocene correlations in boreholes along the shoreline of Israel (partly from Gvirtzman et al., 1984). See Figure 12.1 for locations of the correlation profiles, and Appendix for borehole data. Lithologic notation as in Figure 12.2. Boundary types: 1, lithostratigraphic boundary; 2, biostratigraphic boundary; 3, first occurrence (FO) datum; 4, last occurrence (LO) datum. The points marked A, PT1, PT2, PT3, and BPT are electric-log markers in the Petah Tiqwa Member. Biostratigraphic markers: A, *Amphisorus* sp.; H, *Hyalinea baltica*; G, *Gephyrocapsa oceanica*; P, *Pyramidella plicosa*; D, *Discoaster* spp.; S, *Sphaeroidinellopsis seminulina*; Gm, *Globorotalia margaritae*.



Hefer Formation

The type section of this formation (Gvirtzman et al., 1984) is in the Jaffa 1 borehole (Figure 12.4). The Hefer conformably overlies the Petah Tiqwa Member in the center of the basin and rests unconformably on older truncated formations in the distal parts of the basin, with a thickness varying from about 70 m in the east to about 180 m in the west. The formation is subdivided into 13 members (Gvirtzman et al., 1984) (Figure 12.2), one of which, the Yad Mordekhay Member, wedges out in the westernmost belt of the coastal plain and therefore is not included in the correlation chart of Figure 12.3.

The following lithologies are found in the various members of the Hefer Formation (Figure 12.2):

1. siltstone, marls, marine calcareous sandstones rich in benthic fauna and mollusks of the Dan Member
2. marine marls and clays of the Nizzanim Member
3. marine calcareous sandstones, including beach rocks and coquina beds, of the Lower Bat Yam Member, the Upper Bat Yam Member, the lower part of the Ashdod Member, and the lower part of the Herzliyya Member
4. eolianites of the upper part of the Ashdod Member, the upper part of the Herzliyya Member, the Giv'at Olga Member, and the Tel Aviv Member
5. reddish, clayey-silty sandstones ("hamra") of the Yad Mordekhay Member, part of the Caesarea Member, and layers in the Poleg, Kefar Vitkin, and Netanya members.
6. dark swampy clays with plant remains of part of the Caesarea Member

The Hefer Formation is a product of six sedimentary cycles (Issar, 1961; Gvirtzman et al., 1984) resulting from the eustatic oscillations of Quaternary sea levels. The sedimentary cycles, from bottom to top, are represented by the members as follows:

cycle 1: Petah Tiqwa–Dan–Yad Mordekhay
cycle 2: Lower Bat Yam–Nizzanim–Upper Bat Yam–Caesarea
cycle 3: Ashdod–Poleg
cycle 4: Herzliyya–Kefar Vitkin
cycle 5: Giv'at Olga–Netanya
cycle 6: Tel Aviv

The Ta'arukha Member (Horowitz, 1979) and the Hadera Member (Horowitz, 1979) are two generations of sand dunes which cover part of the coastal plain. The Ta'arukha Member has been partly stabilized since the Middle Bronze Age (Gvirtzman

Figure 12.4. Reference type section of the Plio–Pleistocene sequence of the coastal plain of Israel, in the Jaffa 1 borehole (no. 9 in Figure 12.1 and in the Appendix). The distribution chart of index fossils and biostratigraphic datum levels for the Plio–Pleistocene sequence are based on data compiled herein (see remarks 17, 18, and 19 in the Appendix) and data from Gvirtzman (1970), Gvirtzman et al. (1984), Martinotti (1981a), Moshkovitz (1963), Moshkovitz and Ehrlich (1980), Reiss (1964), and Reiss and Issar (1961). Nannofossil zones after Martini (1971).

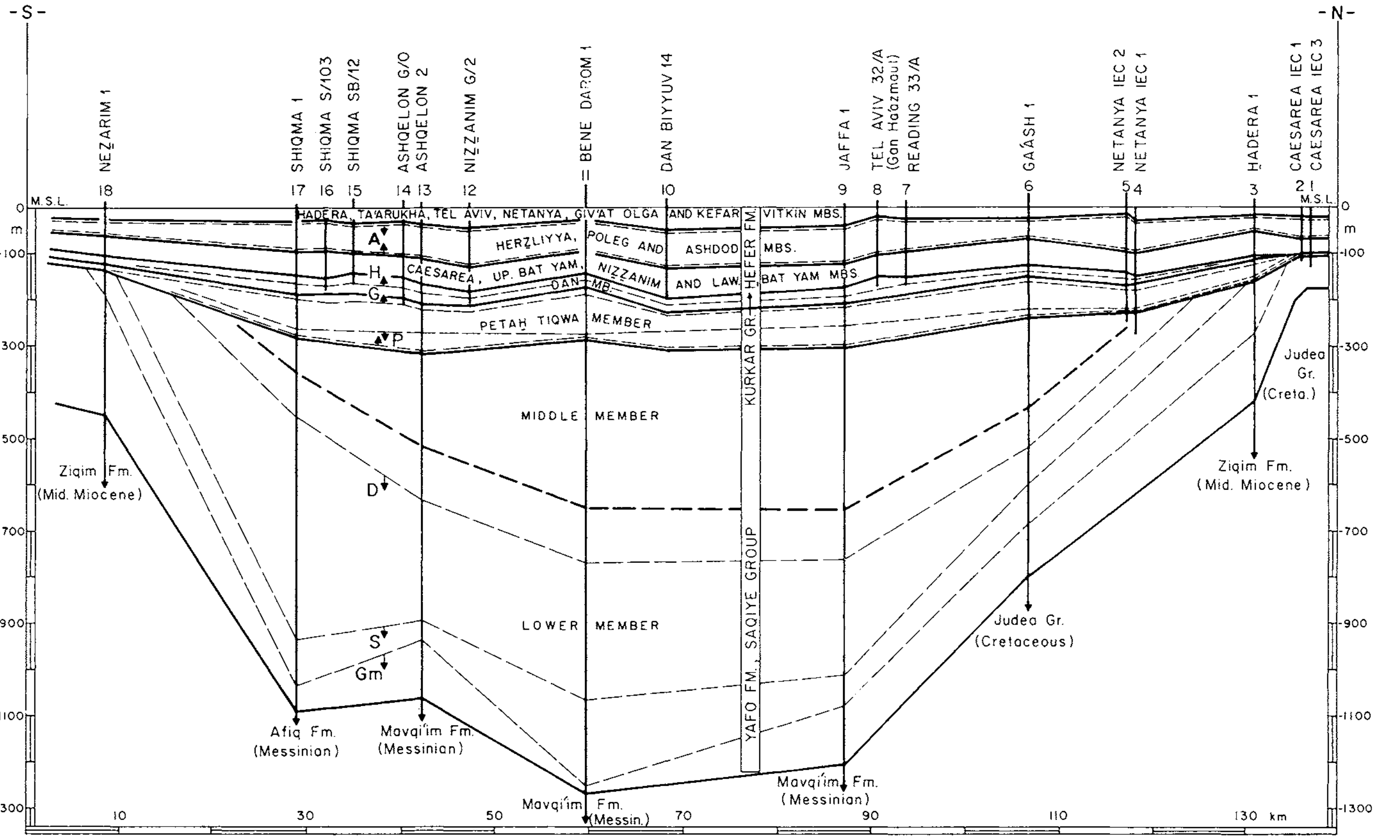


Figure 12.5. Cross section A-B-C of the full Upper Miocene to Pleistocene basinal section in the coastal plain of Israel. Locations of boreholes as in Figure 12.1, and lithostratigraphy, boundaries, and biostratigraphic markers denoted as in Figures 12.3 and 12.4. Note gaps and unconformities in the northern margin (Carmel block) and the southern margin (Gaza–Be'eri block) of the basin.

et al., 1984). The Hadera Member is a result of the rejuvenation of the sand migration since the time of the Byzantines and Crusaders (Gvirtzman et al., 1984).

Biostratigraphy

The distribution and zonations of the marine invertebrates of the Yafo and Hefer formations are well established in the Israeli coastal plain, based mostly on subsurface information from boreholes. The foraminiferal biozonation has been published by Reiss and Issar (1961), Reiss (1964, 1966), Perath (1965), Reiss and Gvirtzman (1966a,b), Derin and Reiss (1973), Moshkovitz and Martinotti (1979), and Martinotti (1981a,b). The biostratigraphy of the mollusks, which occur mainly in the Petah Tiqwa and Dan members, was established by Moshkovitz (1961, 1963, 1968), and nannofossil studies were published by Ehrlich and Moshkovitz (1978), Moshkovitz and Martinotti (1979), and Moshkovitz and Ehrlich (1980). One of the main problems encountered in all those studies was the lack of good continuous cores. The cutting samples from the boreholes were highly contaminated by uphole caving, and the first occurrences of index fossils could not always be accurately determined. Another problem was that the Upper Pliocene and the Pleistocene sequences in Israel are relatively poor in planktonics, and correlations between the eastern and western Mediterranean deposits of this age are not always well founded.

In this chapter, we refer only to occurrences of index fossils whose stratigraphic distributions have been well established, in spite of the previously mentioned limitations. These are as follows: the LO (last occurrence) of *Globorotalia margaritae*; the LO of *Sphaeroidinellopsis seminulina*; the LO of *Discoaster* spp. (including *D. tamalis*, *D. surculus*, *D. pentaradiatus*, and *D. brouweri*, all of which disappear from our area at about the same time because of local paleoecological conditions); the FO (first occurrence) and LO of *Pyramidella plicosa*, the FO of *Gephyrocapsa oceanica*, single occurrences of *Hyalinea baltica*, and the FO and LO of an *Amphisorus* sp. On the other hand, some fossils that occur only rarely and whose exact stratigraphic ranges still are in doubt were not included, in spite of their biostratigraphic importance. Those include *Globorotalia puncticulata*, *G. crassaformis*, *G. bononiensis*, *G. inflata*, *Reticulofenestra pseudoumbilicata*, *Pseudoemiliana lacunosa*, and *Cyclococcolithus macintyrei*, among others.

Datum levels

LO of Globorotalia margaritae. This datum level is well established in many wells in Israel and has been correlated with the global last appearance of the same taxon (Martinotti, 1981b): Lower Member of the Yafo Formation.

LO of Sphaeroidinellopsis seminulina. This well-established datum level is correlated with the last appearance of the taxon in the deepest basins of the Mediterranean (cf. Martinotti, 1981b, 1986): Lower Member of the Yafo Formation.

LO of Discoaster spp. In the subsurface of the coastal plain and continental shelf in Israel, this datum level is related to an ecological event, because all of the species of this genus disappear at a certain level in the sequence in all the boreholes studied in Israel (Ehrlich and Moshkovitz, 1978; Moshkovitz and Martinotti, 1979; Moshkovitz and Ehrlich, 1980, p. 13). The disappearance of the discoasterids from our area occurred during NN-16 and therefore not later than 2.6 Ma, according to the interpolated time scale of Rio, Sprovieri, and Raffi (1984, table II).

This disappearance level, which is found in the Yafo Lower Member, is interpreted in our area as being related to a major change in the discharge of the Nile River in the middle-to-late Pliocene. The extinction of discoasters coincides with a sharp reduction in planktic/benthic ratios and a marked increase in reworked nannofossils of Upper Cretaceous–Lower Tertiary age, together with the appearance of marine and fresh-water siliceous microfossils. This evidence points to abundant influx of alluvial material from the Nile (Ehrlich and Moshkovitz, 1978). At about the same level, we observe a strong increase of shallow-water benthic foraminifera, with "many *Ammonia* and elphidiids of robust size" noted by Perath (1965, p. 19), indicating a reduction in water depth. A parallel decrease or absence of discoasterids in the NN-16 zone in the western Mediterranean was noted by Müller (1979), who associated that phenomenon with the glaciation of the Northern Hemisphere. According to Rio et al. (1984), marked changes in the late Pliocene, at about 2.4–2.6 Ma, occurred in the western Mediterranean when discoasters faded out (with the exception of *D. brouweri*). Those floral events were also related to a severe cooling, in this case one of the precursor cycles (Rio et al., 1984, referring to Thunell and Williams, 1983). Initiations of major global climatic changes and high-latitude glaciation at about 2.4–2.6 Ma have been noted by many authors, both in the Northern Hemisphere (Berggren, 1972; Berggren and Van Couvering, 1974; Shackleton et al., 1984) and in the Southern Hemisphere (Stipp, Chappell, and McDougall, 1967), as well as in the western Mediterranean (Keigwin and Thunell, 1979; Thunell and Williams, 1983; Suc, 1984). Stipp et al. (1967) noted a major drop in sea level as a result of the advances of polar glaciers at that time, and Vail and Hardenbol (1979) also reported a fall in sea level during the NN-16 zone time interval.

Nevertheless, in several locations in the western Mediterranean, such as DSDP site 132 (Raffi and Rio, 1979), at Vrica (Backman et al., 1983; Rio, Raffi, and Backman, Chapter 5, this volume), and at Capo Rossello (Rio et al., 1984; Chapter 5, this volume), the discoasterids show successive extinctions which can be correlated with the extra-Mediterranean biozones. It therefore seems that the disappearance of the discoasterids from our area is a result of three factors which are locally interrelated: global cooling, a lowered sea level, and an increase of freshwater discharge, with simultaneous increase in the flux of continental detritus from the Nile. Thus, shallowing, increasing erosional forces, turbulence, and dilution in this region led, simultaneously, to disappearance of the discoasterids, a decrease of planktonic foraminifera, the dominance of some benthonic foraminifera, and better preservation of allochthonous siliceous microorganisms.

FO and LO of Pyramidella plicosa. This gastropod is the most significant taxon in a mollusk assemblage (Moshkovitz, 1963) that forms the widespread lower coquina bed in the Petah Tiqwa Member (Gvirtzman, 1970). It is always found between the BPT and PT3 E-log markers (Figures 12.3 and 12.4) and is a good tool for local correlations. The so-called *Pyramidella plicosa* zone includes macrofossils such as *Pyramidella plicosa* Brn., *Dentalium michelottii* Hoernes, *D. sexangulum acutangularis* Cocc., and *D. ?passerinianum* Coss., of typical Pliocene age. Because the assemblage also contains some forms representing extant lineages, it has been interpreted to be of late Pliocene age (Moshkovitz, 1963, 1968). Unfortunately, biozones of this kind cannot be correlated with the global stratigraphy of the Plio–Pleistocene (Berggren et al., 1980). The fact, however, that this molluscan assemblage is always found in layers above the NN-16 zone and below the first occurrence of the early Pleistocene *Gephyrocapsa oceanica* is confirmatory evidence of the late Pliocene age of *P. plicosa*. Furthermore, in the Ashqelon 2 borehole (no. 13, Figures 12.1, 12.3, and 12.5), the last occurrence of *P. plicosa* was found at about 304 m, significantly below the first occurrence of *Globorotalia inflata* at 250 m (Martinotti, 1981b). Although *G. inflata* is very rare in our area and has not been established as an index fossil, as discussed earlier, the possibility that these observations represent the regional last occurrence of *P. plicosa* below the regional first occurrence of *G. inflata* should not be excluded.

FO of Gephyrocapsa oceanica. This datum level was found during our investigations in most of the wells studied in the upper part of the Petah Tiqwa Member, slightly below the PT1 electric-log marker (Figure 12.3; see also remarks 13 and 19 in the Appendix). This datum level is correlated with the global first-appearance datum of *G. oceanica*. Detailed evolutionary studies of this species (Samtleben, 1980) and its significance for Pleistocene stratigraphy was recently summarized by Perch-Nielsen (1985, p. 511). In the Mediterranean basin, its first appearance was recorded in the Vrica section slightly above the Olduvai magnetic event (Rio et al., Chapter 5, this volume). Accordingly, it might prove to be a useful tool for both stratigraphic and paleoecologic purposes in our area.

Occurrences of Hyalinea baltica. Along the line of correlation shown in Figure 12.3, this fossil was found in only five boreholes, occurring only rarely (remarks 15, 18, 23, and 27 in the Appendix). In four of the boreholes, its specific stratigraphic position is in the upper part of the Dan Member, above the FO of *G. oceanica*. In the fifth, its stratigraphic position is unknown (remark 27 in the Appendix). This fossil is one of the so-called Ml-zone assemblage, found in boreholes near Tel Aviv (Reiss and Issar, 1961). Its absence higher in the section in the Lower Bat Yam Member might be related to a facies change. This fossil is the only representative in Israel of the "nordic guests" in the Mediterranean, and it is found in an assemblage characterized as the "Calabro-Sicilian" stage (Reiss and Issar, 1961). As in the western Mediterranean area, such as the Vrica section (Pasini and Colalongo, 1982; Ruggieri, Rio, and Sprovieri, 1984, p. 257), this fossil occurs in Israel slightly above the FO of *G. oceanica* in the upper Dan Member. Therefore, in spite of its rarity in Israel, it seems probable that its occurrence is very close to the first appearance of the species.

FO and LO of Amphisorus sp. Avinmelech (1952) included a *Marginopora* sp. and *Amphisorus hemprichi* in a detailed list of microfossils found in the Quaternary of the coastal plain. An unidentified species of *Amphisorus* was also recorded by Itzhaki (1960) from strata that he attributed to the Tyrrhenian. Reiss and Issar (1961) regarded *Amphisorus* to be a synonym of *Marginopora* and characteristic of their Mp and Mpa biostratigraphic zones. Those zones were later equated to the Ashdod and Herzliyya members of the Hefer Formation (Gvirtzman et al., 1984). Following Reiss and Issar (1961), this fossil was recorded as *Marginopora* (Michelson, 1970; Issar and Kafri, 1972; Gvirtzman et al., 1984), but Hottinger (1977, p. 100) and, more recently, Z. Reiss (personal communication, 1987) have restored the identity of this fossil as *Amphisorus*, an amendment followed here. *Amphisorus* is found in the marine calcareous sandstones of the aforementioned two members, usually in the lower part below eolianites. Issar and Kafri (1972) found the gastropod *Strombus bubonius* in the Galilee coastal plain, in *Amphisorus*-bearing calcareous sandstones that subsequently were named the Herzliyya Member (Gvirtzman et al., 1984). The sedimentary cycles of Ashdod (which contains only an *Amphisorus* sp.) and of Herzliyya (which contains both an *Amphisorus* sp. and *Strombus bubonius*) represent in Israel the "warm-water Senegalian fauna of the Tyrrhenian Stage," which is well known from Quaternary shorelines all through the Mediterranean basin. Thus, stages such as the "Tyrrhenien inferieur," "Eutyrrhenien," and "Neotyrrhenien," of Bonifay (1975), or the "shorelines with *Strombus* fauna" (Butzer, 1975), should be considered, in our opinion, as reflecting a circum-Mediterranean "Tyrrhenian Event," based on stratigraphic relationships, paleoclimatic considerations, and human artifacts.

Correlations between the Ashdod and the Herzliyya members and the oxygen-isotope stages 7 and 5, respectively, of Emiliani and Shackleton (1974) were recently suggested (Gvirtzman, 1980; Gvirtzman et al., 1984). From these correlations, it appears that the invasion into the eastern Mediterranean of the Senegalian warm-water fauna lasted during the time of stages 7 and 5.

Evaluation of datum levels and correlations

The nine datum levels defined by the distribution of seven index fossils in the Plio–Pleistocene marine record in the coastal belt of Israel are indicated in the type section of the Jaffa 1 borehole (Figure 12.4). Three of these datum levels are of global significance, namely: *G. margaritae* LO, *S. seminulina* LO, and *G. oceanica* FO (no. 1, 2, and 6, respectively). The other six datum levels are considered as ecologic, climatic, or sedimentologic "events" and can be used for intra-Mediterranean or intra-basinal correlations. The term "event" is used here for designating episodes in a sequence of significant physiographic changes in the huge "natural sedimentation laboratory of the Mediterranean Sea" (Stanley, 1972). This usage follows another well-established term, in the same context and for the same basin: the late Miocene "Messinian Event." Although all of the Plio–Pleistocene events are believed to have happened almost synchronously throughout the Mediterranean, they were not synchronous in the same sense as an evolutionary event. For instance, the local LO of *Discoaster* spp. (datum level 3) reflects an ecologic event related to global glaciation and changes in the Nile. Three other datum levels are related to circum-Mediterranean climatic events of faunal migration. The occurrence of *Hyalinea baltica* (datum level 7) is connected with the invasion of "nordic guests" from the North Atlantic Ocean into the Mediterranean Sea. The FO and LO of *Amphisorus* (datum levels 8 and 9) were related to a warm-water faunal invasion of the Mediterranean from the coasts of western Africa. Two other datum levels (4 and 5) are reflections of a limited local sedimentological event on the Levant shoreline, where a rather thick coquina bed is characterized by the presence of *Pyramidella plicosa* (FO and LO in the same bed).

The Pliocene–Pleistocene boundary

Two possibilities for a biostratigraphic identification of this boundary in Israel can be considered: (1) a boundary based on the first-occurrence datum of *G. oceanica*, as proposed at the INQUA Moscow Congress of 1982 (Pasini and Colalongo, 1982; cf. Pelosio, Raffi, and Rio, 1980); or (2) the level suggested in the draft proposal of INQUA Subcommission 1-d on the Pliocene–Pleistocene boundary (Aguirre and Pasini, 1985), based on the detailed study of the Vrica stratotype (Selli et al., 1977; Pelosio et al., 1980; Obradovich et al., 1982; Pasini and Colalongo, 1982; Chapter 2, this volume; Backman et al., 1983; Tauxe et al., 1983). That latter proposal defined the Pliocene–Pleistocene boundary at the base of the claystone layer overlying sapropelic bed *e* in the Vrica section, approximately 22 m below the FO datum of *G. oceanica* and close to the top of the Olduvai normal event.

The first option would locate the boundary in the Petah Tiqwa

Member, as seen in the Jaffa 1 type section (Figure 12.4), approximately between the PT1 and the PT2 electric-log markers, at a depth of about 234 m. The second option would place the boundary also in the Petah Tiqwa Member, between the PT2 and PT3 electric-log markers, at a depth of about 255 m.

The Mediterranean Pleistocene

In spite of numerous studies, the regional chronostratigraphy of the Mediterranean Pleistocene remains unresolved. The Calabrian, Sicilian, and Tyrrhenian stages have been difficult to identify and correlate in terms of modern global chronostratigraphy (Berggren et al., 1980), although some progress has been made recently (Preface, this volume). For the time being, however, it seems that *Gephyrocapsa oceanica, Hyalinea baltica*, and the *Amphisorus* sp. in our area are useful for intra-Mediterranean correlations. *G. oceanica* is the only fossil which can also be used for global correlation. The possible correlation of Pleistocene sedimentary cycles of Israel with the global oxygen-isotope stages, as mentioned earlier, seems to hold promise for future work in this regard.

Event stratigraphy and correlation with the Nile Delta

The lithostratigraphic sequences of the coastal plain of Israel, from the Messinian evaporites upward, in conjunction with the biostratigraphic datum levels and their chronostratigraphic interpretations, have permitted detailed correlation within the Nile Cone sequence in the coastal belt of Israel (Figure 12.5) and also, to some extent, within the entire Mediterranean basin. A sequence of events can be deduced from the regional correlation chart (Figure 12.5), which in turn will throw new light on the evolution of the Nile Delta and the Nile Cone. The reconstructed events, in their stratigraphic order, are summarized in the following scheme.

Evidences of desiccation, sebkha deposits, salt lakes, and evaporite accumulations are recorded in the Upper Miocene Mavqiim Formation in Israel (Figure 12.2). Late Miocene erosion, evidenced in a buried topography of canyons, channel fill, and drainage patterns, can be deduced from study of the Afiq Formation and from interpretation of seismic records for the coastal plain of Israel. The widespread early Pliocene transgression is connected with the beginning of the buildup of the modern Nile Delta and Nile Cone, in the form of basinal and hemipelagic muds, deposited as deltaic foreset beds, recorded in the Yafo Lower Member. A middle–late Pliocene (ca. 2.6 Ma) dilution of the planktonic biomass, together with the extinction of the discoasterids, probably resulted from global cooling and a change in the discharge of the Nile River. That extinction event is recorded in the upper part of the basinal foreset beds of the Lower Member of the Yafo Formation.

Late Pliocene differential movements, in which the Carmel and the Gaza-Beeri blocks were uplifted and eroded, while the central basin was downfaulted, took place prior to the deposition of the Petah Tiqwa Member (notably in the northern and southern marginal areas of the basin, Figure 12.5). A westward tilt of the entire coastal plain (Neev et al., 1978; Gvirtzman, Dickenstein, and Croker, 1980) and truncation in the eastern coastal plain resulted in unconformities between the Lower and Petah Tiqwa members of the Yafo Formation. Other results of the westward tilt were slumps and growth faults in the eastern lobe of the Nile Cone. Deformation occurred mostly in the Middle Member of the Yafo Formation. Alternations of Upper Pliocene coquina beds with sand units and muds, as a result of regional shallowing, are recorded in the Petah Tiqwa Member.

The dominance of clays and silts in the depositional supply, as seen in the Saqiye Group, gave way to a dominantly sandy supply for the Kurkar Group during the early Pleistocene. Though some sand units are found in the Petah Tiqwa Member, and some mud units are found in the Dan Member, the overall change in lithology is quite marked.

An early Pleistocene invasion of the Mediterranean by "nordic guests" from the North Atlantic coasts is documented in the upper part of the Dan Member by the occurrence of *Hyalinea baltica*. An early Pleistocene renewal of the earlier block movements and of the westward tilt is deduced from the erosional surface found below the Lower Bat Yam Member in the Carmel area. During that pre–Bat Yam erosional phase, the Ahuzam conglomerate, a fluvial sediment which covers the eastern coastal plain (Gvirtzman et al., 1984) and the lower Shefala and cuts valleys into the upper Shefala (Sneh and Buchbinder, 1984), was deposited. The Yad Mordekhay hamra was also deposited during the phase between the Dan and the Lower Bat Yam members (Gvirtzman et al., 1984) (Figure 12.2). A late Pleistocene invasion of West African "warm-water Senegalian fauna" into the Mediterranean, and its extinction thereafter, is documented by the occurrence of *Amphisorus* and *Strombus bubonius* in the Ashdod and the Herzliyya members.

The Plio–Pleistocene sequence of the Nile Delta, described by Rizzini et al. (1978) and by Said (1981), is correlated with the sequence of the coastal belt of Israel, using mainly lithostratigraphic criteria, as follows:

1. The Mavqiim Formation and the Rosetta Formation are time-equivalents, both composed of Messinian evaporites.
2. The Afiq Formation and the Abu Madi Formation represent the fluvial coarse clastics found at the base of the Nile Delta and Cone in the nearshore area.
3. The Lower and Middle members of the Yafo Formation and the Kafr el Sheikh Formation are the main body of the foreset muds in the Nile Delta and Cone.
4. The Petah Tiqwa Member and the El Wastini Formation represent the transition from the foreset mud units of the Nile Delta to the overlying topset sand units. Both formations are composed of alternating beds of muds and sands.
5. The Hefer Formation and the Mit Ghamr Formation represent the topset sand beds of the Nile Delta in the nearshore area.

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Appendix

The following remarks refer to the wells denoted in Figures 12.1, and 12.3–12.5. Numbers 1–18 refer to the 18 well sites studied for this chapter, and numbers 19–27 refer to details of the lithology and biostratigraphy of certain wells, as indicated.

1. Caesarea IEC 3. Lithostratigraphy after Gvirtzman et al. (1984).
2. Caesarea IEC 1. Lithostratigraphy prepared for this chapter.
3. Hadera 1. Biostratigraphy previously published.
4. Netanya IEC 1. Biostratigraphy according to the authors.
5. Netanya IEC 4. Biostratigraphy after Moshkovitz and Martinotti (1979).
6. Ga'asha 1. Biostratigraphy after Martinotti (1981c).
7. Reading 33/1A. Biostratigraphy after Martinotti (1981b).
8. Tel Aviv 32/A. Biostratigraphy after Martinotti (1981a).
9. Jaffa 1. Biostratigraphy after Moshkovitz (1963).
10. Dan Biyyuv 14. Biostratigraphy after Reiss and Issar (1961).
11. Bene Darom. Biostratigraphy after Reiss (1964).
12. Nizzanim G/2. Biostratigraphy after Perath (1965).
13. Ashqelon 2. Biostratigraphy after Moshkovitz and Ehrlich (1980), who noted a single occurrence of *Gephyrocapsa oceanica* in core no. 1 (205 m depth). However, as a result of recent detailed study, we noted the first occurrence of *G. oceanica* in this well at a depth of 224 m, very close to the electric-log marker PT1.
14. Ashqelon G/O. Biostratigraphy after G. M. Martinotti (in Moshkovitz and Ehrlich, 1980, table 2).
15. Shiqma SB/12. *Hyalinea baltica* was not found in this well. However, Z. Reiss (in Moshkovitz, 1961) found this fossil in the nearby Reading 33/0 well (410 m to the northwest) in the interval 164–181 m. That interval was projected to the correlative depth interval of this well, as illustrated in Figure 12.3.
16. Shiqma S/103. This well is included in this report to show the lithostratigraphic correlation according to the authors.
17. Shiqma 1. The presence of *Amphisorus* was not determined in this well. The presence of this fossil in the depth interval cited (Figures 12.3 and 12.4) was projected from several nearby wells, according to information from Reiss and Issar (1961) and also from the archives of the Geological Survey of Israel.
18. Nezarim 1. *Hyalinea baltica* was reported by Martinotti (in Moshkovitz and Ehrlich, 1980, table 2) as a single occurrence in core no. 1 (205 m). We were unable to find any other specimens at or below that level, despite careful examination.
19. In the Jaffa 1 well (no. 9), the ranges of distribution of *Gephyrocapsa oceanica, Hyalinea baltica*, and *Amphisorus*, as shown in Figure 12.4, are the maximum ranges according to correlations and compilations of all the studied boreholes, as shown in Figure 12.3.
20. In the Bene Darom well (no. 11), no electric log is available for the upper part or for the complete sequence of this well. The lithostratigraphic correlation is based only on well cuttings.
21. In the Ashqelon 2 well (no. 13), no electric log is available for the upper part of this well, and the well cutting samples are not reliable.
22. In the Ashqelon G/O well (no. 14), the presence of *Amphisorus* was not determined, but is assumed because of occurrences at equivalent depths in the two nearby wells Nizzanim 10/0 (980 m to the northeast) and OBS-1 (360 m to the northeast). Data from Ecker and Olsina (1978).

23. In the Nizzanim G/2 well (no. 12), *Hyalinea baltica* occurs at depths between 199 and 200 m and is also found in the nearby Nizzanim B-76–8 well (860 m to the southeast) at a depth of 192 m, both in the upper part of the Dan Member.
24. The Jaffa 1 well (no. 9) is the type section of the Yafo Formation (Gvirtzman, 1970) and of the Hefer Formation (Gvirtzman et al., 1984).
25. In the Shiqma 1 well (no. 17), acoustic and density logs for this well were converted into a synthetic seismogram (Klang and Gafsou, 1981). A significant and continuous seismic horizon from the Yafo Formation was identified in the synthetic seismogram near the base of the Petah Tiqwa Member in this well.
26. In the Nezarim 1 well (no. 18), *Sphaeroidinellopsis seminulina* was not found at the depth interval of 149–200 m in this well, probably because of ecological factors. However, the *S. seminulina* zone is tentatively recognized from the occurrence of *G. padana* without *G. margaritae* (Martinotti, 1981c).

13 The Pliocene–Pleistocene transition in the Iberian Peninsula

EMILIANO AGUIRRE

Introduction

The transition from the late Neogene to the Pleistocene in the Iberian Peninsula was characterized by widespread erosion on the Meseta and its margins because of structural deformation associated with subsidence over wide areas, both along the Mediterranean shorelands and in the intramontane basins along the Betic front. Marine deposition was rare and was restricted to littoral fringes. Volcanic activity in the southeast, and also in the southern Meseta, decreased after the end of the Miocene. The paleomagnetic record is extremely limited. However, there are significant sites in continental environments with both small and large vertebrates (Figure 13.1) that provide the material for interregional correlations.

Stratigraphic sequences

Bay of Cádiz

The Bay of Cádiz, in the area between the Guadalquivir depression and the western margin of the Subbetic realm, was first inundated during the late Miocene Tortonian transgression. The subsidence of the bay was related to the advance of the Subbetic olistostrome to the north-northwest, and its present configuration was influenced by a normal fault along an east-northeast axis.

Torre del Puerco section. This section (36°19′20″N; 2°29′45″W) begins with marls and marly limestones of the TP-1 (Torre Puerco Unit 1) containing *Globorotalia margaritae*, overlain by TP-2 marly sands and bioturbated offshore horizons with *G. crassaformis*. Near its top, *G. margaritae, G. crassaformis*, and *Turborotalia puncticulata* disappear, but *Globigerinoides extremus* and *Turborotalia humerosa* are still present. The TP-3 shallow conglomerates that close this cycle yield abundant rhodoliths of Lithotamniaeae, with *Flabellipecten planomedius, Pseudoamussium calcatur, Pecten jacobaeus*, and *Ostrea lamellosa*. The top of the series is continentally karstified, and it disconformably underlies a lower to early middle Pleistocene terrace with Olduwan-like pebble tools. The paleomagnetic record shows a gradual shift from reversed to normal and again to reversed polarity in the lowest unit (TP-1), which may correspond to the middle Gilbert chron.

Puerto Real section. The quarries beside the highway near Puerto Real (36°32′20″N; 2°29′W) offer the most nearly complete section in this area. At the base, a regressive sequence termed PR-1 (Puerto Real Unit 1) is seen, with offshore sands changing to offshore marginal calcarenites with *Globorotalia crassaformis*; this cycle ends with a shore-edge conglomerate. A roughly transgressive sequence, PR-2, follows, with minor oscillations dominated by lagoon and barrier-island facies; it ends with shallow-marine conglomerates. The upper part is affected by continental karstification, and that level is followed by a new cycle, PR-3, with conglomerates, sands, and lagoon deposits. Within this last unit, in a horizon of yellow sands, a fragment of a tooth of *Mammuthus meridionalis* has been found. The sequence ends in a karstified surface buried by an extensive regional terrace.

El Aculadero section. The Plio–Pleistocene sequence at El Aculadero (36°34′45″N; 2°33′50″W) starts with the facies known as the Ostionero (viz., "conglomerado ostionero" or oyster-bearing conglomerate). Common fossils in this basal sequence are *Chlamys varia, Chlamys opercularis, Chlamys glabra, Pecten jacobaeus, Pecten maximus, Ostrea lamellosa*, and *Flabellipecten planomedius*. The assemblage is indicative equally of late Pliocene or early Pleistocene age. The surface of the Ostionero is karstified, and pockets are infilled with red marine sands, with subaerial soils at the top. The section continues upward into marine–brackish lagoonal silts with interfingering sands (wash-over fans) indicating breakup of a barrier island during a transgressive oscillation. The end of that episode is marked by a marine conglomerate. During the following hiatus, a new karst was developed, infilled and overlain by a terrace deposit or glacis containing Early Paleolithic tools of Olduwan tradition (Querol and Santonja, 1983, figure 1, A–I; figure 5, LP). The top layer is fersialithic soil.

Cantera de la Florida. In the exposures of an abandoned quarry at La Florida, near El Puerto de Santa María (Cádiz), Zazo et al.

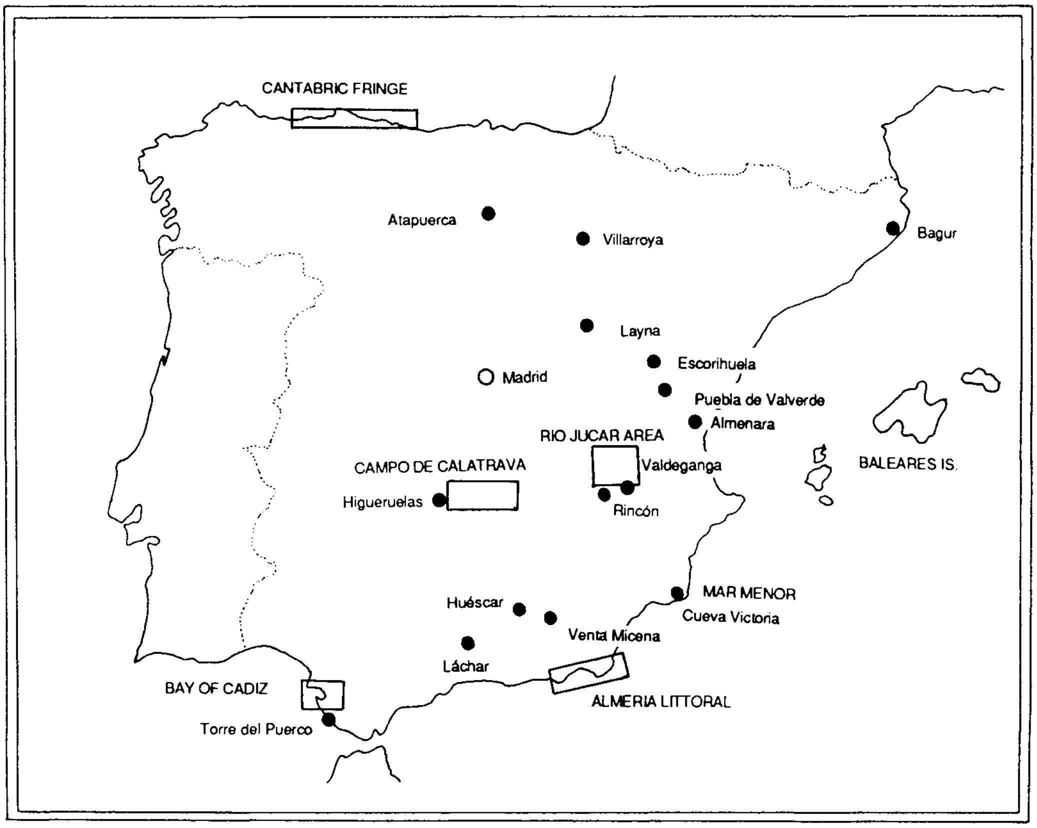

Figure 13.1. Index map of Iberian Pliocene–Pleistocene study areas.

(1985) identified the paleochannels of a former southern outlet of the Guadalquivir River, to which the Guadelete was tributary in early Pleistocene time. The earliest quartzite pebbles deposited by the Guadalquivir in fluviomarine strata at the La Florida mouth, according to those authors, were very near the base of the Pleistocene.

Cerro de los Mártires. Diaz, Parra, and Benot (1988) recognized the Pliocene–Pleistocene boundary in marine strata cropping out at Cerro de los Mártires, south of San Fernanda, Cádiz, where alternating yellowish sands and blue clays contain two planktonic foraminifera faunas. The lower fauna is characterized by an assemblage of *Globorotalia tosaensis, Globorotalia inflata, Globorotalia crassula crassula, Globorotalia crassula viola, Globorotalia* ex gr. *G. crassaformis, Globorotalia praehirsuta, Globorotalia bononiensis, Turborotalia quinqueloba, Globigerina uvula, Globigerina apertura, Globigerina* ex gr. *G. bulloides, Globigerinoides conglobatus, Globigerinoides elongatus, Globigerinoides obliquus obliquus, Globigerinoides obliquus extremus* (very rare), *Globigerinoides ruber, Neogloboquadrina acostaensis, Neogloboquadrina humerosa, Neogloboquadrina pachyderma* (dextral), and *Globorotalia truncatulinoides*. The upper fauna includes *Globigerina hessi, Globigerina* aff. *G. calida, Globigerina* aff. *G. digitata, Globigerinoides tenellus, Neogloboquadrina dutertrei*, and sinistrally coiled *Neogloboquadrina pachyderma*. The Pliocene–Pleistocene boundary is correlated to the transition between the two assemblages, that is, just below the first appearance of *Globigerina hessi*.

Sanlúcar de Barrameda and Moguer. At several localities near Moguer, Huelva, and Sanlúcar de Barrameda, Torcal et al. (1991) have studied the connection between the Pliocene–Pleistocene boundary and climate changes, taking into account sedimentological evidence and local stratigraphy in the Bay of Cádiz. Within the Arenas de Bonares, correlated to the regional Arenas Rojas or "Red Sands" formation, those authors identified a lower horizon with illite, smectite, and dominant calcalkaline feldspar, representing subtidal deposition, and an upper sandy horizon with dominant potassic feldspar and kaolinite, without smectite, which represents an intertidal regime. The Pliocene–Pleistocene boundary is close to the transition between these two suites, which could be attributable to local diastrophism or to a climate change (Torcal et al., 1991), so that the epoch boundary could in fact be located within the upper unit, near its base.

Summary of major features. In the Cádiz area, sedimentation was discontinuous during the time in question, and a hiatus in deposition is almost universal in the upper part of the sequence. The Upper Pliocene (Piacenzian) probably is represented by the lower Ostionero and the succeeding erosional interval. The overlying Arenas Rojas is known all along the Atlantic shore of the southern Iberian Peninsula, from Algarve, Portugal, to Algeciras on the Mediterranean side of the Gibraltar Strait. The formation may represent estuarine bars developed at the same time as some of the lower Ostionero; it should be noted that on those estuarine sands a primary red soil is developed. The succeeding transgressive marine sediments, including correlatives such as the TP-4 and PR-2 (and, locally, later Ostionero-facies deposits), probably are early Pleistocene in age.

A final regressive episode, marked by the deposition and karstification of the lower conglomerate at El Aculadero and the upper conglomerate at Puerto Real, represents another major feature. I suggest, with some reservation, that those terminal processes belonged to the general lowering of sea level and the severe climate changes that marked the early middle Pleistocene (i.e., Menapian) glacial maxima. A change from tensional to compressive style in regional tectonism is conspicuous near the top of the lower Ostionero beds. The new compressive phase, which has lasted to the present, is characterized by strike-slip faults, both right- and left-lateral (Benkhelil, 1976; Zazo et al., 1977; Zazo, 1989).

The evidence strongly suggests that the Pliocene–Pleistocene boundary, as defined in the Vrica section (Aguirre and Pasini, 1985), can be correlated to a lowering of the sea level that is reflected in the karstic erosion of the late Pliocene Ostionero Conglomerate and in the change from shallow-marine to estuarine facies in a number of localities. In other locations, the Pliocene–Pleistocene boundary appears to be correlated to a level within the Arenas Rojas, the base of which is time-transgressive in the Bay of Cádiz region from Huelva to Algeciras (Zazo, 1989).

Mediterranean shorelines

The Cantabric fringe. According to Hoyos Gomez (1989), the transition from Pliocene to Pleistocene in the Cantabric coastal region was linked to the development of the extensive erosion surface known as the Rasa. The main erosional phase took place in the later Pliocene, as indicated by the fact that the final surface of the Rasa is closely echoed in the earliest Pleistocene drainage and seems to have lasted through most of the early Matuyama, between 2.6 and 2.0 Ma. After formation of the Rasa surface, the initial incision of the present river systems and the development of a karstic erosion surface, with abundant decalcification clays and stream deposits in caves, date from a period clearly antecedent to the severe climate shifts that began in the middle Pleistocene. Hoyos Gomez (1989) concluded that the Pliocene–Pleistocene boundary cannot be traced closer than this very approximate indication, because of the absence of more precise dating.

Iberian coastline. In the littoral of the east coast of the Iberian Peninsula and in the Balearic archipelago, the highest well-preserved marine terraces are at elevations above MSL (mean sea level) ranging from +65 m in Alicante to +90 m in Almeria. According to Goy et al. (1992), the ages of these highest terraces appear to be early Pleistocene. They are preceded either by regressive Pliocene marine deposits or by major unconformities of tectonic origin. An exception is the Torrevieja–Mar Menor district, where terraces are not evident because of continuous subsidence throughout the Pleistocene.

The Spanish Meseta

The best correlations between the Pliocene–Pleistocene boundary-stratotype at Vrica and the continental strata of Spain are found in the raña terrace formations on the margins of the Meseta, in the Orce deposits of the Baza Basin, and (potentially) in the cave fillings of Casablanca. In those deposits, paleomagnetic and paleontologic evidence and rare radiometric age determinations have been integrated in a framework of regional geodynamic history.

La Raña. The local lithofacies known as La Raña is a cobble piedmont sheet with pseudogley soil. Although the perfect synchrony of all raña sheets can be debated, it is currently agreed that the raña lithology is related to neotectonic uplifts that ended basinal sedimentation in the Meseta region (Pérez González and Gallardo, 1987). Tentatively, La Raña is dated older than 2 Ma; it is followed in the Meseta by two or more erosional events marked by terraces at +200 and +180 m ASL, probably of latest Pliocene age. Following the incision of these transitional forms, the drainage of the Meseta region to the ocean began, and the graded river-valley terraces started to develop. The number of terrace levels attributed to the early Pleistocene in the major valleys of the Meseta is four to six, starting with landforms at elevations of +155 m above present grade; the next lowest, a terrace of +80–70 m, is usually judged to be pre-Cromerian.

Raña deposits occur in separated areas of the Spanish Meseta, always in marginal areas bordering the neighboring mountain ranges. According to Molina (1975), in the Campo de Calatrava of the Guadiana River basin the middle Pliocene (S-1) surface was moderately deformed prior to the development of a new surface at lower elevations during a period of warm climate with marked seasonality. That new surface was covered by raña materials laid down in a series of anastomosing channels under even more pronounced seasonal climate conditions, as indicated by caliche crust formed on the deposit surface. That surface, named S-2 by Molina (1975), was then dissected, and the new landscape was covered by *glacis* (wind-sorted gravel terraces), considered by Molina (1975) to represent the earliest deposits of the Pleistocene. The V-III eruptive episode in the Calatrava volcanic field appears to have begun during the raña deposition and caliche formation, but after reaching a maximum during the dissection of the S-2 surface the eruptions ceased before deposition of Lower Pleistocene alluvial sediments (López-Ruiz et al., 1993). In the volcanic edifice of Cabezo de Segura, magnetostratigraphic studies by Calvo Sorando et al. (1992) have shown that a deposit correlative to the local raña is interbedded between two volcanic layers with reversed remanent polarity and that the lower volcanic layer rests on a surface that corresponds to the upper "polygenetic surface" of the Meseta and the Iberic Chain.

A geomorphologic analysis of raña formations along the southern margin of the Sierra de Somo and the Sierra de Ayllón in the Cordillera Central led Pérez Gonzáles and Gallardo (1987) to conclude that repeated raña erosional and depositional processes followed the Ibero-manchega II uplift and shortly preceded the earliest (highest) valleys of the Jarama and Sorbe river systems. Pérez Gonzáles and Gallardo (1987) offered no conclusion as to the relationship of the raña to climate change, but in my opinion the concentration of quartzite cobbles in the raña deposits is best explained as the result of active gelifraction on tectonized quartzites in the newly uplifted sierras. According to M. Hoyos Gomez (personal communication, 1991), the general mode of raña deposition is in the form of a regional *glacis*, not as debris cones or alluvial fans; this is logically explained as a reflection of specific climate conditions, as suggested by Molina (1975), and not as the result of tectonism alone.

After accounting for the effects of local diastrophism, the raña deposits described in the two studies just cited would appear to be products of arid, highly seasonal climates in which soil formation and fluvial erosion were less active, and deflation was more active, than in prior or following regimes. The conditions for raña formation ended with renewed erosion and indications of a moister climate, which could be related to global cooling. Such events at about 2.1 Ma, and at about 1.8 Ma (the end of the Olduvai normal-polarity subchron), may have been implicated, and a correlation with the Aullan unconformity of Tuscany (Italy) is thus a tentative possibility. In this hypothesis, La Raña of Calatrava and the foothills of the Somo and Ayllón sierras would mark the very end of the Pliocene, just prior to the Pliocene–Pleistocene boundary.

There is no evidence that rañas in other areas on the Meseta borderlands differ significantly in age from the foregoing. The only difference is that these accretion surfaces were built up with materials other than quartzite cobbles, but that must be expected in areas where the source rocks were different. Correlation of rañas deserves detailed study as a means of tracing the Pliocene–Pleistocene boundary in a large area of central Spain.

The Mediterranean shoreline

In the Mediterranean littoral of the Iberian Peninsula and in the Balearic Islands the uppermost traces of sea level are found at +65 m ASL in Alicante and Murcia, and at +90 m ASL in Almeria. These beach lines represent early Pleistocene sea levels, being younger than regressive deposits and erosion of Pliocene age. In the Torrevieja–Mar Menor basin, on the other hand, Lower Pleistocene deposits were laid down under conditions of continuing subsidence (Goy et al., 1992).

Radiometric dates

The stratigraphic sections studied in the Iberian Peninsula for the purposes of contributing to IGCP Project 41 are not well documented with radiometric dates. Among the very few datable volcanic layers that could be of interest for the period in question are the San Isidro Basalt (2.8–2.6 Ma) at Rambla de la Auia and Venta del Lirioh near Cartagena, Murcia province, and the Hostalrich (3.08 Ma) and San Corneli (1.99 Ma) basalts at La Selva, Gerona province. Unfortunately, the stratigraphic positions of these dated rocks in the respective regional sequences are as yet quite uncertain.

In Campo de Calatrava, a region of late Cenozoic basaltic volcanism in Ciudad Real province, the Higueruelas mammal site yields remains of *Anancus arvernensis, Hipparion rocinantis, Cervus* cf. *C. cusanus, Cervus* cf. *C. perrieri*, and *Gazella* cf. *G. borbonica*, in a fluvial deposit that includes coarse volcanic debris. No dates have been obtained for the co-deposited volcanic fragments, but the fauna corresponds to zone MN-16, with approximate age limits of 3.5 and 2.6 Ma. In support of this estimate, basalt flows that unconformably overlie the Higueruelas beds have been given a K-Ar age of 2.3 ±0.4 Ma (R. E. Drake and J. A. Van Couvering, personal communication). More recently, M. Hoyos Gomez (personal communication, 1992) dated the Higueruelas beds to the Gauss chron, according to a radiometric age of about 3 Ma on the overlying lavas, accompanied by paleomagnetic analyses that indicated normal polarity in the fossil bed and reversed polarity in strata immediately below. As noted earlier, Molina (1975) and López-Ruiz et al. (1993) concluded that these Upper Pliocene Calatrava volcanics had been erupted during the development of the regional erosion surface that is capped by the terminal Pliocene raña deposits, whereas Calvo Sorando et al. (1992) have suggested that some of these volcanics were of the same age as the raña formation. Evidently, the reduction of the landscape and the gradual accumulation of raña carapace were parts of a long-term geomorphological process that in the Calatrava region was periodically interrupted by volcanism.

Plio–Pleistocene mammalian faunas

Elephants and horses appear for the first time in the strata of the Júcar River valley at levels corresponding to the small-mammal horizons Valdeganga I and II (Mein, Moissenet, and Truc, 1978). *Equus stenonis* and *Mammuthus* ex gr. *meridionalis* are in this site associated with *Mimomys medasensis, Mimomys capettai, Stephanomys balcellsi, Castillomys crusafonti crusafonti*, and *Sus strozzii*, an assemblage which corresponds to MN-16 age (i.e., from 3.5 to 2.6 Ma). Rincón 1 appears to be similar in age, with *Oryctolagus* sp., *Eliomys intermedius, Mimomys capettai, Castillomys crusafonti crusafonti, Nyctereutes megamastoides, Pachycrocuta* sp. (of larger size), *Equus stenonis, Dicerorhinus* sp., *Gazella borbonica*, and other artiodactyl genera. Paleotemperature determinations and correlation of the Rincón section with the normally magnetized lower third of the red clays and

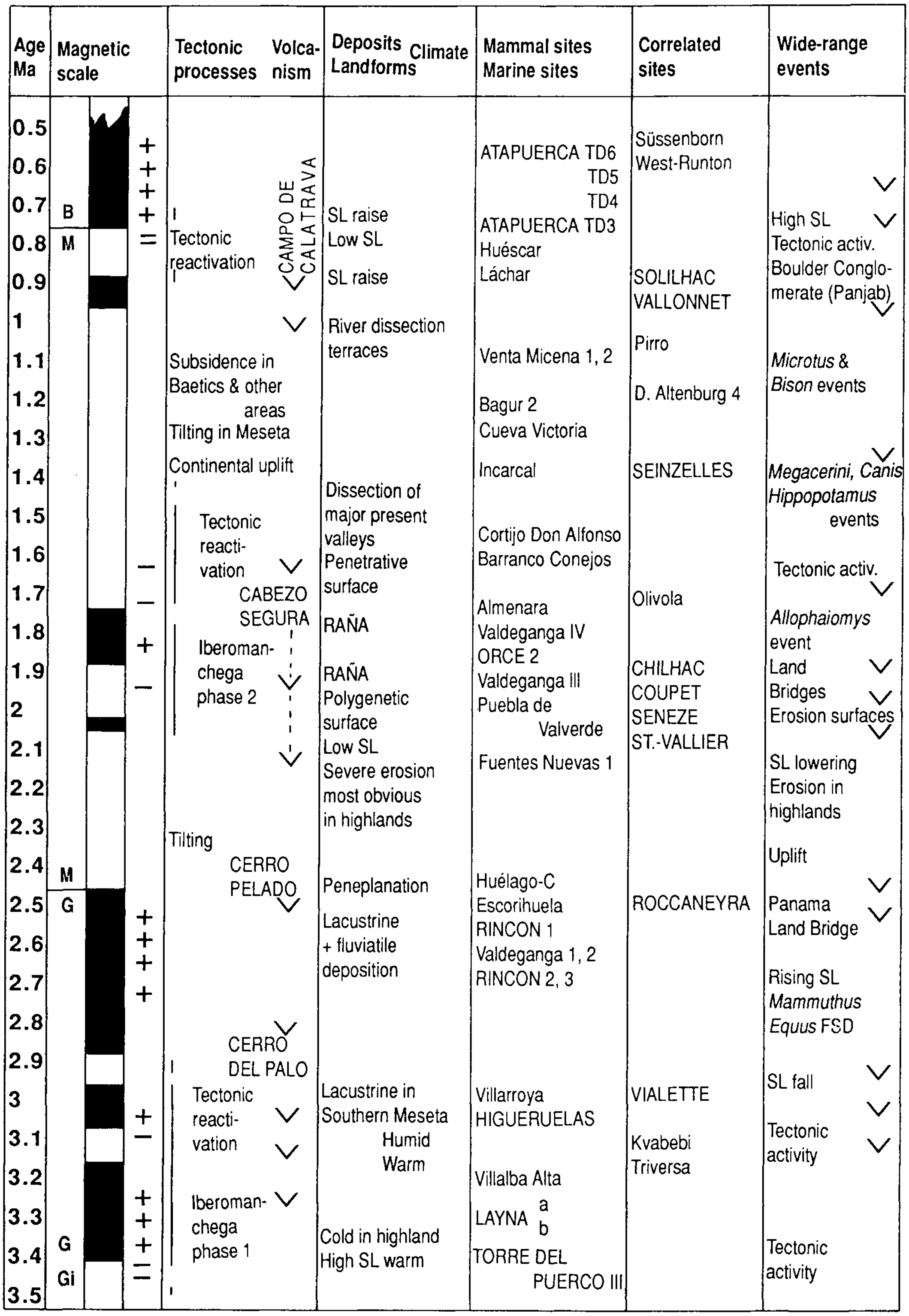

Figure 13.2. Correlation of main vertebrate fossil localities and geologic features of Spain, from the Lower Pliocene (lower Gauss) to the Middle Pleistocene (Matuyama–Brunhes boundary).

sandstones of Villalgordo suggest a maximum age in the very latest Gauss chron (Alberdi et al., 1982) for Rincón 1 and Valdeganga I–II faunas, in the later part of MN-16 (Figure 13.2).

A later regional fauna, of MN-17 age, is typified in Spain by the Puebla de Valverde local fauna (Heintz, 1978). That site, in Teruel province, yields *Paradolichopithecus arvernensis, Macaca* sp., *Nyctereutes megamastoides, Vulpes alopecoides, Ursus etruscus, Hyaena perrieri, Chasmaporthetes lunensis, Megantereon*

megantereon, Acinonyx pardinensis schaubi, Lynx issiodorensis, Mammuthus meridionalis, Dicerorhinus etruscus, Equus stenonis vireti, Croizetoceros ramosus, "Cervus" philisi, Eucladoceros senezensis, Gazella borbonica, Gazellospira torticornis, and *Gallogoral meneghini*. This is a typical Middle Villafranchian assemblage (Heintz, 1978), or later Villanyan, in the modern usage.

In the upper horizons of the Júcar canyon, Albacete province, the Valdeganga III site yields *Mimomys* aff. *M. medasensis, Mimomys rex, Stephanomys progressus, Castillomys crusafonti* subsp. indet., *Micromys* aff. *M. minutus, Apodemus* aff. *A. dominans, Eliomys* aff. *E. quercinus, Parapetenya hibbardi, Sorex subminutus, Desmana nehringi, Prolagus calpensis, Oryctolagus* aff. *O. laynensis*, and *Equus stenonis* subsp. indet. (Mein et al., 1978). The assemblage appears to correlate best to the characteristic later Villanyan faunas of Villány 3 and Kisláng.

The karst filling in the Casablanca quarry, near Almenara, Castellón province, has yielded abundant large- and small-mammal fossils, including *Prolagus calpensis, Desmana inflata, Myotis* cf. *M. myotis, Miniopterus* aff. *M. schreibersi, Rhinolophus* cf. *R. mehelyi, Eliomys* sp., *Stephanomys progressus, Apodemus* cf. *A. mystacinus, Apodemus* aff. *A. occitanus, Castillomys crusafonti* subsp. indet., *Mimomys tornensis, M. medasensis, Mimomys* aff. *M. rex, Ursus etruscus, Pachycrocuta brevirostris, Felis* sp. indet., *Dicerorhinus* cf. *D. etruscus, Equus stenonis* subsp. indet., *Cervus philisi, Gazellospira torticornis*, and Ovibovini gen. indet. (Soto and Morales, 1985; Esteban Aenlle and López Martínez, 1987). The rodent assemblage is comparable to that of Villány 5 and Kadzielna, and rather more to the latter, while the large-mammal assemblage is closely correlative to that of Olivola.

The MN-17 mammal assemblages are followed by a new faunal complex, characterized in Bagur 2 karst filling by *Allophaiomys pliocaenicus, Lagurus pannonicus, Pliomys episcopalis, Ungaromys* sp., *Castillomys crusafonti, Apodemus* aff. *A. mystacinus, Apodemus* aff. *A. sylvaticus, Eliomys* aff. *E. intermedius, Prolagus* cf. *P. calpensis*, and *Oryctolagus* cf. *O. lacosti* (López Martínez, Michaux, and Villalta, 1976). *Allophaiomys* is also recognized in younger horizons of the Baza basin and in the Venta Micena local fauna discussed later.

A very extensive detritic formation in the central basin of the Río Guadiana is found at the present river level, east and west of Ciudad Real. From a stock-pond excavation, this deposit yielded *Mammuthus meridionalis, Eucladoceros dicranios, Leptobos* sp., *Hippopotamus major*, and *Equus mosbachensis*. This association can be considered transitional to the early middle Pleistocene, or Cromerian–Galerian faunas. The identification of *Mammuthus meridionalis tamanensis* at a quarry on a terrace near Fuensanta, in the Júcar Valley, together with *Hippopotamus major* (Aguirre, 1989b), conveys the possibility of recognizing the Tamanian stage in Spain.

The lower half of the known cave-fill section at Atapuerca Burgos (Atapuerca I of Aguirre et al., 1990) yields *Mimomys savini, Pliomys episcopalis*, and *Pitymys gregaloides*. Atapuerca I also yields the large mammals *Crocuta intermedia, Panthera gombaszoegensis, Dicerorhinus hemitoechus, Equus* cf. *E. stehlini*, Megacerini, *"Cervus"* sp., and *Bison schoetensacki*, an assemblage similar to that of the lower Galerian of Italy. Near the base of this fossiliferous sequence, a geomagnetic-polarity reversal identified as the Matuyama–Brunhes transition has recently been identified (Garacedo, Soler, and Chicharro, personal communication, 1991).

Plio–Pleistocene faunas of the Betic depressions

The Guadix and Baza basins are linked tectonic depressions presently drained by the Guadalquivir River. In late Neogene and early Pleistocene time they were closed endorheic basins that filled with lacustrine deposits. Large and small mammals are found throughout the sequence, and in the Baza Basin in particular they constitute an almost continuous record from the early Pliocene to the early middle Pleistocene.

In the time span relevant to the Pliocene–Pleistocene boundary, the sites with the richest fossil records are (in ascending chronostratigraphic order) as follows: Carretera de Huélago (correlative to Montopoli and Étouaires); Fuentes Nuevas 1 (correlative to Puebla de Valverde, in Teruel province, and to Saint-Vallier) (Aguirre et al., Chapter 9, this volume); Orce D and Orce 1 (correlative to Senèze, Chilhac, and Le Coupet); Orce 2 (correlative to Olivola, Montoussé, and other sites placed in the lower *Mimomys ostramosensis* zone of Chaline, Chapter 14, this volume); Barranco de los Conejos (correlative to sites in the upper *ostramosensis* zone), which Agustí, Moyà-Solà, and Pons-Moyà (1987b) have equated with Casa Frata; Venta Micena 1 and 2 (correlative to early Biharian faunas at Betfia 2, Betfia 13, and Altenburg 2); Barranca Léon, Fuentes Nuevas 3, and Cañada de Murcia 1 (correlative to levels slightly younger than Venta Micena); Huéscar 2 and 3 and Puerto Lobo 1 (correlative to Le Vallonet on the basis of *Allophaiomys nutiensis*); Huéscar 1 (correlative to Atapuerca TD3–TD6, with small-mammal faunas assigned to the zone of *Pitymys gregaloides* and *Mimomys savini*); Cúllar de Baza (correlative to later Cromerian and later Galerian on the basis of *Arvicola cantiana* and *Mammuthus trogontherii*).

Also to be mentioned are the Cueva Victoria fauna from La Unión, near Cartagena, which is slightly older than the Venta Micena fauna and correlates to the Sinzelles faunal level (1.3 to 1.4 Ma), and Láchar in the Granada Basin, which is well correlated to the Solilhac faunal level.

The Plio–Pleistocene faunal succession of the Betic basins begins with Carretera de Huélago, from which Alberdi and Bonnadonna (1989) reported *Mammuthus meridionalis, Equus stenonis livenzovensis, Dicerorhinus* cf. *D. etruscus, Leptobos cf. L. elatus, Gazella borbonica, Gazellospira torticornis*, Ovibovini cf. *Hesperoceros merlae, Croizetoceros ramosus, Eucladoceros* cf. *E. senezensis*, Cervidae gen. indet., *Mimomys* cf. *M. reidi*, and *Stephanomys* sp.

The Fuentes Nuevas 1 site yields *Mimomys* cf. *M. reidi, Castillomys crusafonti, Apodemus* cf. *A. dominans, Equus stenonis* aff. *E. s. vireti*, and *Gazella borbonica*. The two teeth and metatarsal of the equid from Fuentes Nuevas resemble most closely those of the variety found at Puebla de Valverde and Saint-Vallier.

At Orce 2, the fossil list includes *Drepanosorex* sp., *Mimomys ostramosensis, Mimomys pusillus, Castillomys crusafonti, Microtus* (cf. *Allophaiomys*) sp., *Apodemus* aff. *A. sylvaticus, Apodemus mystacinus, Eliomys* aff. *E. quercinus, Galemys kormosi, Gazellospira torticornis*, and *Leptobos etruscus*. The co-occurrence of the last two taxa is known elsewhere only at Olivola. Sediments at Orce 2 show normal paleomagnetic polarity (Agustí et al., 1987a). An assemblage similar to Orce 2 is recorded at Barranco de los Conejos from beds 8a and 9, corresponding to the lower part of unit *b* in the informal notation of Anadón et al. (1987).

The extensive outcrops at Venta Micena 1 and 2 yield fossils from stratigraphic unit *c* of Anadón et al. (1987). As described by Agustí et al. (1987a), the list from Venta Micena 2 includes *Desmana* sp., *Microtus (Allophaiomys) pliocaenicus, Apodemus* aff *A. mystacinus, Castillomys crusafonti, Eliomys intermedius, Hystrix major, Prolagus calpensis, Oryctolagus* cf. *O. lacosti, Ursus etruscus, Canis etruscus mosbachensis, Vulpes preglacialis, Cuon priscus, Xenocyon sp., Homotherium latidens, Megantereon cultridens adroveri, Pachycrocuta brevirostris, Panthera* cf. *P. gombaszoegensis, Lynx* sp., *Meles* sp., *Mammuthus (nec Archidiskodon) meridionalis, Equus stenonis granatensis, Dicerorhinus etruscus brachicephalus, Hippopotamus incognitus, Praemegaceros solilhacus, "Cervus" elaphoides, Praeovibos* sp., *Capra alba, Soergelia minor, Bison* sp., Caprini gen. indet., *Testudo* sp., and *Lacerta* sp.

Barranca Léon has yielded fossils from at least three levels: L2 and L3, in unit *d* of Anadón et al. (1987), and L1, which is in upper *d* and lower *e*. Another site in the lower part of unit *e* is the rich horizon of Orce 7. The fossils from all these levels may be regarded as a single assemblage consisting of *Microtus (Allophaiomys) pliocaenicus, Apodemus mystacinus, Apodemus* aff. *A. sylvaticus, Castillomys crusafonti, Eliomys intermedius, Hippopotamus* sp., *"Cervus" elaphoides, Bison* sp., *Capra alba, Soergelia minor, Equus stenonis*, and *Mammuthus meridionalis*. Orce 3, at a slightly higher stratigraphic horizon, is relatively poor, but it includes a *Galemys* sp. and a *Mimomys* aff. *M. savini*.

An unresolved question is whether or not *Mimomys pusillus* has been identified correctly from Orce 2, in unit *b*, and the correlative site of Valdeganga IV. If so, its appearance is anomalous with regard to the first occurrence of *Microtus (Allophaiomys) pliocaenicus* in Venta Micena at the lower boundary of unit *c*. In regions to the north of the Pyrenees, the appearance of *M. pusillus* is delayed until after the first appearance of *M. pliocaenicus*, at a level that is recognized as equivalent to the top of the Olduvai normal-polarity subchron (Chaline, Chapter 14, this volume). The apparent presence of *M. pusillus* in normally polarized (Olduvai?) strata at an older level than *M. pliocaenicus* indicates that one or the other had a different time range in southern Spain than in the regions farther north, and it is possible that this district is beyond the paleogeographic limit for exact correlation of the Pliocene–Pleistocene boundary on the basis of central European microtine evolution.

Outside of the Guadix–Baza basin, but also in the southeastern part of Spain, the site of Cueva Victoria is a cave complex that was completely filled with hyena debris in the early Pleistocene. Most of the infilling bone-breccia was later eroded, but the remaining material includes remains of *Crocidura* sp., *Erinaceus* sp., *Prolagus calpensis, Oryctolagus* cf. *O. lacosti, Myotis* sp., *Rhinolophus euryhale, Rhinolophus* cf. *R. mehelyi, Miniopterus* sp., *Allophaiomys chalinei, Eliomys quercinus, Apodemus mystacinus, Vulpes* sp., *Canis etruscus, Equus stenonis, Equus* sp., *Mammuthus meridionalis, Dolichodoriceros savini, "Cervus" elaphoides, Hemitragus* sp., Ovibovini gen. indet., Caprini gen. indet., Bovini gen. indet., and Papionini cf. *Papio* sp. The presence of *Homo* sp. has been reported from Cueva Victoria, and from Venta Micena as well, but identification remains doubtful (Gibert-Clols et al., 1989). Cueva Victoria, on present evidence, is best correlated to Sinzelles; Venta Micena is not much younger.

The fauna at Huéscar 1, in the upper part of the Baza Basin sequence, includes many bird fossils. Mammals listed by Alberdi et al. (1989) are Soricidae gen. indet., *Eliomys quercinus, Apodemus* sp., *Castillomys crusafonti, Mimomys savini, Pitymys gregaloides, Microtus brecciensis, Oryctolagus* sp., *Lepus* cf. *L. granatensis, Canis etruscus, Panthera gombaszoegensis, Homotherium* sp., *Elephas (Palaeoloxodon) antiquus, Equus stenonis, Equus suessenbornensis, Dicerorhinus etruscus brachycephalus, Hippopotamus major*, and *Praemegaceros* cf. *P. solilhacus*. This assemblage is comparable to the early Galerian, or, in other words, close to the older faunas at Atapuerca, and thus to the transition between early and middle Pleistocene. The fauna at Láchar, in the Granada Basin, is a slightly older assemblage and includes *Mammuthus meridionalis, Equus stenonis granatensis, Dicerorhinus etruscus, Bison* sp., Bovidae gen. indet., *Praemegaceros* sp., *Capreolus* sp., and *Dama* cf. *D. clactoniana* (Aguirre, 1989b; B. Azanza and B. Sánchez, personal communication, 1991). The Láchar fauna is very similar to that of Solilhac, which is to say close in time to the transition from the terminal early Pleistocene paleofauna, such as Vallonet, to earliest Cromerian and Galerian assemblages.

In sum, the abundance of rich fossil sites, representing in one stratigraphic sequence almost every recognized faunal event straddling the Pliocene–Pleistocene boundary, suggests the value of the Guadix–Baza basin for illustrating the Pliocene–Pleistocene transition in the Iberian Peninsula and the ease with which this biostratigraphic succession can be correlated in the regional synthesis. I therefore expressly propose here that the basin-filling sequence of the Guadix–Baza basin should be adopted as a parastratotype area for the Pliocene–Pleistocene boundary in continental sediments, with the expectation that this will encourage stratigraphers working in this region to identify and propose a formal reference section.

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14 Biostratigraphy and calibrated climatic chronology of the Upper Pliocene and Lower Pleistocene of France

JEAN CHALINE

Introduction

In the years following the consensus on the Pliocene–Pleistocene boundary in France (Chaline and Michaux, 1972a) and the observation (Chaline, 1977) that the appearance of *Allophaiomys pliocaenicus* in European mammal faunas is synchronous with the Neogene–Quaternary boundary proposed by IGCP-41 at the top of the Olduvai event, there have been several further developments. In this chapter we review, in succession, the following: the new "rodent biostratigraphy" and the correlations that it implies; geochronologic data permitting new calibrations; new palynological and faunal data of significance for paleoclimatic correlations in northern Europe; and a reevaluation of the classic sites with respect to the Neogene–Quaternary boundary.

"Rodent biostratigraphy"

In the classic sites of Plio–Pleistocene age, the large mammals have provided the basis for a biostratigraphy that approaches a succession of assemblage-zones or cenozones. In practice, however, the true affiliations of the large mammal species from the various levels are seldom known for certain, because it is only rarely that sufficient numbers of individuals are found in a given assemblage to permit reliable determinations of the intraspecies and interspecies variations. The same is not true for rodents, which are notably very abundant in cave fillings, but can also occur in equal numbers in fluviolacustrine deposits. The best-documented rodent lineage in the interval under discussion is that which began with *Mimomys occitanus*, with seven species ranging from the middle Pliocene to middle Pleistocene (Chaline and Michaux, 1972a,b).

The Mimomys occitanus davakosi–Mimomys savini lineage

The phylogeny of the *Mimomys* taxa that make up the evolutionary succession *occitanus davakosi–occitanus occitanus–occitanus hajnackensis–pliocaenicus polonicus–pliocaenicus pliocaenicus–pliocaenicus ostramosensis–savini* is based on a biometric study using multivariate analysis of detailed dental morphology (Chaline, 1983; Chaline and Laurin, 1986), which shows that these species-level taxa correspond in their successive degrees of evolution to a single anagenetic or diachronic lineage evolving in place in Eurasia, that is, western Europe to central Siberia. This allows, for the first time, establishment of a high-resolution mammal biostratigraphy, with the advantages that the remains are abundant in sites very close to one another in time and space, and that geochronologic data are available to calibrate this evolutionary sequence and to discriminate variable evolutionary rates (Chaline and Laurin, 1986; Chaline and Farjanel, 1990).

Geochronological data

Four sites have yielded radiometric dates which calibrate evolutionary stages in the *Mimomys occitanus davakosi–savini* lineage.

Perrier-Étouaires (Auvergne). This site has yielded a tooth of *Mimomys pliocaenicus polonicus* from the mammal fauna found between the Roccaneyra Basalt and the chalky layer *np.1*. The earliest analyses, by Curtis (Bout, 1966), suggested dates of 3.9 and 3.3 Ma, respectively, for the enclosing strata. A subsequent date obtained by Lippolt (Bout, 1966) raised the age of this fauna to 3.1 Ma.

More recently, Chambaudet and Couthures (1981) obtained an age of 3.14 Ma based on fission-track analysis of volcanic sphene. In a subsequent study, which took into account new stratigraphic observations, Pastre, Chambaudet, and Couthures (1983) proposed a revised age for the fauna of 2.25 ±0.25 Ma. Considering the results of paleomagnetic analyses, which show a normal polarity for the *np.1* level (Sémah and Biquand, 1978), the Perrier-Étouaires fauna most probably should be situated at the end of the Gauss normal chron at about 2.6 Ma. Note that recent palynological correlations (discussed later), which indicate that the ancestral *Mimomys occitanus* dates from no more than 3.0 Ma, support this estimate.

Le Coupet (Auvergne). This site has been dated at the locality of Saint Georges d'Aurac. A few teeth of *Mimomys pliocaenicus* s.

str. recovered from a volcanic tuff beneath the Coupet basalt flow were dated by Savage and Curtis (Chaline and Michaux, 1972a) at 1.92 Ma. The petrography of the fossiliferous sediment indicates that it is virtually contemporaneous with the flow that overlies it.

Valerots à Nuits-Saint-George (Côte d'Or). This cave filling has yielded a fauna with *Mimomys savini*, together with advanced *Allophaiomys pliocaenicus* (Chaline et al., 1985).

Courterolles (Yonne). A fauna similar to the preceding, with evolved *Allophaiomys pliocaenicus* (Brochet, Chaline, and Poplin, 1983), is also found in cave deposits. These faunas with evolved *A. pliocaenicus* are best placed between the "Waalian" and "Cromerian I" interglacials, in the Menapian cold-climate interval above the Jaramillo normal subchron.

Climate chronology of the late Pliocene and early Pleistocene of France

Calibrated biostratigraphy can be linked with contemporaneous climate history. By incorporating data from sedimentology, faunal analysis, and palynology, a climatochronology can be constructed in which the geologic histories of northern and southern Europe can be compared. The presence of rodent remains together with those of large mammals in classic assemblages (Perrier-Étouaires, Chagny, Saint-Vallier, Le Coupet, etc.) makes it possible to replace the earlier chronology of the Villafranchian with a new system and to reinterpret the faunas from a paleoecological point of view.

Stratigraphic evidence. From the stratigraphic point of view, studies in the south of France by Michaux (1980) and Suc (1980) and in the Bresse basin, between Dijon and Lyon, by the "groupe Bresse" have provided information concerning the later Pliocene. The discovery of a fluvioglacial gravel interstratified with the Bresse marls (Senac, 1981), situated below the Vonnas horizon with *Mimomys pliocaenicus*, is of particular importance. This gravel could be related either to the Pretiglian or to the Eburonian, but in any event it represents the most ancient well-dated indication of fluvioglacial conditions in western Europe.

Again in Bresse, at Montagny-les-Beaune, beds with *M. pliocaenicus pliocaenicus* and remains of tortoises are overlain by a stratum with strongly wind-sculpted calcareous blocks, implying a period of denuded and arid soils, which in these latitudes would occur during cold phases (i.e., Eburonian).

Palynological evidence. Drill cores in the Bresse basin have produced rodent remains (Chaline, 1984) from within palynological sequences that are readily correlatable with those of northern Europe (Farjanel, 1985). A core from Servignat-Montmain yielded a tooth of *M. pliocaenicus* in a paleofloral context of interglacial type, probably Tiglian. A core from Labergement-les-Seurre at La Mare-du-Bois produced a tooth of *M. pliocaenicus ostramosensis* from a bed overlying a Tiglian flora and below one of Eburonian age. Finally, a core of Simard-Meix-Pernod contained a tooth of *M. p. ostramosensis* or *M. savini* in a bed above the Eburonian level, in a paleofloral context that may represent the beginning of the Waalian (Chaline and Farjanel, 1990).

For the southern part of western Europe, Suc (1980) has proposed correlations between micromammal zones according to palynological evidence (Suc and Zagwijn, 1983). Zone MN-15 of Mein (1975), with *Mimomys occitanus*, would thus range between 3.0 and 2.65 Ma, with the fossil mammal site of Sète at 2.8 Ma (Aguirre et al., Chapter 9, this volume; Azzaroli et al., Chapter 11, this volume).

Faunal evidence. At Perrier-Étouaires, the well-known large-mammal fauna (Bout, 1960) includes a tapir, a forest pig, and numerous cervids (Heintz, 1970), together with a *Taxodium* flora that suggests a partly forested landscape of interglacial (Reuverian?) type, as further indicated by the presence of *Hystrix*. At Saint-Vallier (Drôme), the loess fauna (Viret, 1954), of (upper?) Pretiglian age, contains an incisor of *Ochotona* associated with remains of *Mimomys pliocaenicus polonicus*. Faunas come from various levels at Chagny, but the tapir is unquestionably from a basal layer (Reuverian?), whereas the rodents come from the upper levels, with teeth of *M. pliocaenicus ostramosensis* transitional to *M. savini* and teeth of a species referred to *Allophaiomys*; the correlation is possibly to the final part of the Eburonian.

The site of Montoussé 5 (Clot et al., 1975) contains, together with *M. pliocaenicus ostramosensis*, some teeth of a lemming in the genus *Lemmus* (the most southerly known occurrence) and of *Macaca florentina*. The correlation is to the end of the Eburonian glacial-climate phase or the beginning of the Waalian interglacial.

In the Bresse basin, sediments at Commenailles, Magny, Montagny, and Vonnas contain remains of desmanids, which imply running water. At Montagny, the tortoises suggest a temperate climate. The correlation of Commenailles is to the Pretiglian, Montagny and Magny to phase C4 of the Tiglian, and Vonnas to the Eburonian (Chaline, 1984; Chaline and Farjanel, 1990). At Trevoux-Reyrieux, in the southern part of the Bresse basin, the presence of a southern vole, *Mimomys capettai*, indicates an interglacial phase (Reuverian?).

Other elements of climate correlation in Europe

In the stratigraphy of the *Mimomys occitanus davakosi–savini* lineage, evidence from other parts of Europe can be applied to the developments in France. At Wölfersheim (Tobien, 1952) the fauna has a tropical aspect, with a tapir (*Tapirus arvernensis*), flying squirrels (*Petaurista*), a cercopithecid (*Macaca*), *Prolagus*, and tortoises. All of these elements are of interglacial type; according to Brunnacker and Boenigk (1976), Wölfersheim is dated to the Gauss normal chron. According to Kowalski (1960), the type site of *Mimomys pliocaenicus polonicus*, at Rebielice-Krolewski, also contains the most primitive known lemming

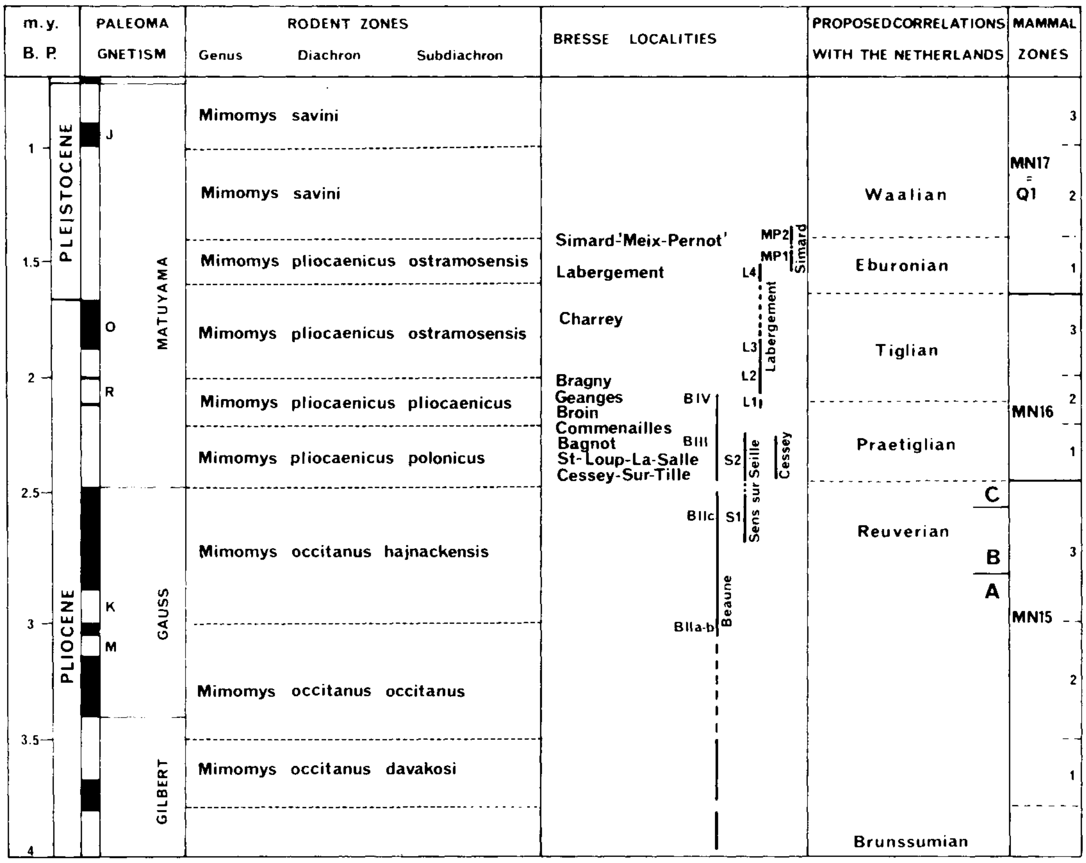

Figure 14.1. Eurasiatic Plio–Pleistocene correlations between rodents of the *Mimomys occitanus–M. savini* lineage and palynostratigraphy. (Adapted from Chaline and Farjanel, 1990.)

(*Synaptomys europaeus*), the spermophile *Citellus polonicus*, and the most primitive member of the avian genus *Lagopus*, attesting to a cold-steppe regime, probably Pretiglian.

In The Netherlands, the interglacial-type fauna at Tegelen, with *Macaca* and *Mimomys pliocaenicus*, is in the Tegelen C5 level, in reversed-polarity sediments between the normal paleomagnetic events attributed to the Réunion subchron and the Olduvai (as "Gilsà") by Freudenthal, Meijer, and Van der Meulen (1976; see also Zagwijn, Chapter 16, this volume), and thus about 2.0 Ma in the orbitally calibrated time scale. Deposits correlated to the earliest Eburonian at Brielle yield primitive *Allophaiomys pliocaenicus* and the collared lemming *Dicrostonyx* (Van der Meulen and Zagwijn, 1974).

In central Europe, Janossy and Van der Meulen (1975) have described the type of *Mimomys pliocaenicus ostramosensis* at Osztramos, Hungary, together with *Allophaiomys pliocaenicus* and *Lemmus lemmus*, in an assemblage correlative with Eburonian conditions. Finally, at Neuleiningen II (Malec and Tobien, 1976) and at Schernfeld (Dehm, 1962), a similar association with *Lemmus* indicates the same correlation, despite the fact that *Allophaiomys* has not yet been discovered at Schernfeld.

Chronobioclimatic synthesis in Europe and the Quaternary boundary

The European record of the *Mimomys occitanus davakosi–savini* lineage (Chaline and Laurin, 1986) supports a chronologic, biologic, and climatic schema that covers not only French territory but also all of Europe and northwestern Asia as far as western Siberia (Zazhigin, 1980). In the latter region, the cited *Mimomys* species and subspecies are considered to be taxonomic synonyms of the indicated characterizing species of the biozones.

The Upper Pliocene and Lower Pleistocene European biozonation (Figure 14.1) is founded on the evolutionarily successive morphospecies *Mimomys occitanus*, *M. pliocaenicus*, and *M. savini*, which are considered to represent three diachrons (a spatiotemporal chronomorphocline unit characterized by a degree of evolution). These diachrons are further divided into nine subdiachrons (Chaline, 1983; Chaline and Farjanel, 1990) as follows:

Mimomys occitanus davakosi biozone. Type locality: Ptolemais 3 (Macedonia, Greece).

Localities: Spain (Villalba Alta, Arquillo); France (Serrat d'en Vacquer) (P. Mein and J. Michaux, personal communication); and Russia (Ob Plateau of Siberia, with the type of *M. antiquus*, a synonym of *M. occitanus*).

Age range: 4.0 to 3.6 Ma.

Mimomys occitanus occitanus biozone. Type locality: Sète (Hérault, France).

Localities: Slovakia (Ivanovce B, with the type of *Mimomys hassiacus atavus*, a synonym of *M.*

occitanus); Hungary (Osztramos 9, with the type of *Mimomys silasensis*, a synonym of *M. occitanus*).
Age range: 3.6 to 3.0 Ma.

Mimomys occitanus hajnackensis biozone. Type locality: Hajnácka (Slovakia). Because the name *stehlini* used previously was ambiguous with respect to lineage, it was decided at the Rohanov conference (Czechoslovakia) in 1987 to replace it with the name *hajnackensis*. This taxon is without doubt related to the preceding lineage as indicated.

Localities: Spain (Escorihuela); Italy (San Giusto); Germany (Wölfersheim); Poland (Weze); Hungary (Csarnota 2); Moldavia (Kotlovina); Ukraine (Kryzhanovka, Kujalnick, Uryv 1); Russia (Akkulaevo, in Bashkiria, with the type of *M. gracilis akkulaevae*, a synonym of *M. o. hajnackensis*).
Age range: 3.0 to 2.6 Ma.

Mimomys pliocaenicus polonicus biozone. Type locality: Rebielice 1 (Poland).

Localities: France (Perrier-Étouaires, Saint-Vallier, and Bresse Valley sites of Cessey-sur-Tille, Beaune BIIIa–BIVa, Commenailles, Broin, Chagny 2, Bagnot, and Satin-Loup-la-Salle); Italy (Arondelli); Moldavia (Kotlovina); Ukraine (Kujalnick, Kryzhanovka, Uryv 2); Russia (Liventsovka, in the Don basin, with *M. livenzovicus*, a synonym of *M. pliocaenicus polonicus*).
Age range: 2.6 to 2.3 Ma.

Mimomys pliocaenicus pliocaenicus biozone. Type locality: Castelfranco (Italy).

Localities: England (Norwich and Weybourne Crag); The Netherlands (Tegelen); France (Saint Georges d'Aurac and Bresse Valley sites of Vonnas, Labergement-les-Seurre, Comblanchien, Montagny-les-Beaune, Charrey, Bragny, and Géanges); Poland (Kadzielna, Kielniki); Moldavia (Kotlovina); Russia (Platovo 1 in the Russian plain; Lebiaje, Podpusk, and Kizhikia in southern Siberia).
Age range: 2.3 to 2.0 Ma.

Lower Mimomys pliocaenicus ostramosensis biozone. Type locality: Osztramos 3 (Hungary).

Localities: France (Montoussé 5 and the Bresse Valley site of Demigny); Germany (Schernfeld, Neuleiningen); Poland (Kamyk).
Age range: 2.0 to 1.8 Ma.

Upper Mimomys pliocaenicus ostramosensis and Allophaiomys deucalion biozone. Type locality: Villány 5 (Hungary).

Localities: The Netherlands (Brielle); France (Senèze, Balaruc 1, and Bresse Valley sites of Chagny 1, Servignat-Montmain, and Simard-Meix Pernod); Austria (Deutsch Altenburg 2C); Slovakia (Mokra 1); Hungary (Osztramos 3); Romania (Betfia 2); Poland (Kamyk, Zaba Cave).
Age range: 1.8 to 1.4 Ma.

Lower Mimomys savini and Allophaiomys pliocaenicus biozone. Type locality: Püspökfürdö (Romania).

Localities: France (Mas Rambault, La Sartanette); Spain (Bagur 2); Austria (Deutsch Altenburg 4B–4C); Romania (Betfia 1, 3, 7).
Age range: 1.4 to 1.1 Ma.

Upper Mimomys savini and Microtus burgondiae–Allophaiomys nutiensis biozone. Type locality: Les Valerots (France).

Localities: England (Westbury 1); France (Courterolles); Spain (Venta Micena, La Victoria); Italy (Soave, Monte Peglia, Selva Vecchia); the former Yugoslavia (Marjan); Czech Republic (Chlum 6, Vcelare 4, Holjstein); Poland (Kozi Grzbiet); Ukraine (Nogaisk, Khairy); Russia (Tichonovka, near Vladivostok).
Age range: 1.1 to 0.8 Ma.

Identification of the Quaternary boundary

On the basis of studies in the Calabrian marine sequence of Italy by INQUA Subcommission 1-d, "Pliocene–Pleistocene Boundary," and by IGCP Project 41, "Neogene/Quaternary Boundary" (see Foreword, this volume), the physical stratotype of this boundary in the section at Vrica has been defined at a level which equates with the lowest (first) appearance of species indicating a cold climate. According to current calibrations, that cold-climate phase commenced at 1.8 Ma, slightly below the top of the Olduvai event (Pasini and Colalongo, Chapter 2, this volume). It follows that this cold phase in the European continental sequences must be identified with the Eburonian paleoclimatic stage of the lower Rhine basin (Zagwijn, Chapter 16, this volume).

In accepting this correlation to the Eburonian as a first approximation, subject to the support of complementary data, the Neogene–Quaternary limit in France and in Eurasia should therefore be identified in small-mammal biochronology primarily on the basis of the occurrence of *Allophaiomys pliocaenicus*, alone or jointly with *Mimomys pliocaenicus ostramosensis* (Chaline, 1977). In France, that was the period when the infilling of the Bresse basin was completed, with the overlying deposits showing the effects of the climate change to the Waalian interglacial, as for instance at Chagny 1 (Chaline and Farjanel, 1990).

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15 The Plio–Pleistocene of England and Iceland

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Plio–Pleistocene in England

The so-called Crag formations of East Anglia are the only British examples of a sedimentary sequence that is partly Pliocene and partly early Pleistocene. They are composed of marine nearshore and estuarine shelly sands and clays, in some horizons with good pollen assemblages. The oldest unit, the Coralline Crag, has always been accepted as Pliocene. Early attempts to locate the base of the Pleistocene in that sequence were guided by the belief that its position would be indicated by evidence of a sharp drop in temperature. Thus, at the International Geological Congress in 1948, it was recommended (King and Oakley, 1949) that in England the Neogene–Quaternary boundary should be placed in the East Anglian sequence at the base of the Butleyan Red Crag, the highest and youngest of Harmer's (1902) three Red Crag "zones." A number of authors (e.g., Boswell, 1952) subsequently proposed that it should be lowered to the base of the Red Crag, on the assumption that the transition from the underlying Coralline Crag indicated a major climatic deterioration. On the premise that the boundary was associated with the Olduvai normal paleomagnetic event, Hey (1977) and Funnell (1977) independently concluded that in East Anglia it could lie above the top of the Red Crag.

It must be emphasized that the paleomagnetic data available from the East Anglian succession are scanty (Van Montfrans, 1971, pp. 100–101), and materials suitable for radiometric dating appear to be absent. Any attempt to locate the boundary in the East Anglian succession must depend on paleontological correlations with the succession in the Rhine basin, which is both more nearly complete and more securely related to the paleomagnetic time scale (Zagwijn, Chapter 16, this volume; von der Brelie et al., Chapter 17, this volume).

Age of the Red Crag

A borehole at Stradbroke, 30 km north-northeast of Ipswich, passed through nearly 70 m of marine sediments, of which the uppermost 45 m were assigned palynologically to the Ludhamian Stage, whereas the lower beds, resting directly upon the Chalk Formation, yielded a pollen sequence unlike any recorded elsewhere in Britain (Beck, Funnell, and Lord, 1972). Funnell (1977) found that the foraminifera of the pre-Ludhamian beds indicated a correlation, on the one hand, with the typical Red Crag and, on the other hand, with foraminifera zone FA2 of The Netherlands. This is the equivalent of the nonmarine Reuverian Stage, which appears to lie entirely within the Gauss normal epoch and has long been accepted as Pliocene (Zagwijn and Doppert, 1978, figures 2 and 3; Zagwijn, Chapter 16, this volume). The pollen of the Stradbroke pre-Ludhamian suggests a similar correlation (R. G. West, personal communication, 1981). That evidence made it virtually certain that the whole of the Red Crag should be relegated to the Pliocene.

An objection to that conclusion arose from the claim that the Red Crag had yielded remains of the elephant *Mammuthus ("Archidiskodon") meridionalis* (Stuart, 1974). In the Dutch succession, that species appears to be unknown below the upper Tiglian, almost two full stages above the Reuverian (Azzaroli, 1970, p. 113; Zagwijn, 1974, p. 86). In Italy, similarly, it is known only from the Upper Villafranchian (Azzaroli et al., Chapter 11, this volume). Such evidence suggests that *M. meridionalis* first appeared in Europe only in the latest Pliocene, not before 2 Ma, in which case its presence in the Red Crag would imply that the unit should be of Pleistocene age. Dr. A. J. Stuart (personal communication) has, however, pointed out to the writer that the determination of fragmentary remains of *"Archidiskodon"* is a matter of some difficulty and that the identification of the Red Crag material should be considered provisional.

In the Dutch succession, according to the latest available information (Gibbard et al., 1991; Zagwijn, Chapter 16, this volume), the top of the Olduvai, and hence the Neogene–Quaternary boundary, lies at the base of the Eburonian Stage, if not in the uppermost part of the underlying Tiglian. In East Anglia, the Crag sediments immediately succeeding the Red Crag (e.g., Norwich Crag, Icenian Crag, etc.) have been divided by West (1961) into a number of stages (from lowest to highest): Ludhamian, Thurnian, Antian, Baventian, Beestonian, and Pastonian. On palynological grounds, Zagwijn (1974) correlated zone TC3 of the mid-Tiglian with the Ludhamian, whereas the foraminifera evidence suggested to Funnell (1977) that the Ludhamian correlated to zone FA1 in the foraminifera zonation of The Netherlands, which is placed at the top of the Tiglian (Zagwijn and Doppert, 1978). Those two lines of evidence therefore combine to suggest that the Ludhamian should join the Red Crag in the Pliocene. The most recent correlation, however, between the British and Dutch stages, as summarized by Gibbard et al. (1991, figure 8), indicates that the horizon equivalent to the base of the Eburonian lies in the erosional interval which everywhere marks the transition between the base of the Beestonian and the underlying deposits. Thus, not only the Red Crag but also most (or perhaps all) of the Norwich Crag would be of Pliocene age.

The Pliocene–Pleistocene boundary in Iceland

Rocks described as Plio–Pleistocene, dated to between 3.1 and 0.7 Ma, appear at outcrop over some 25,000 km^2 of Iceland (Saemundsson, 1980). They are predominantly volcanic, but incude minor intercalations of sediments, including tillites. Several local successions have now been described in detail, with paleomagnetic and some radiometric dates, and these studies suggest that at least six glacier-forming episodes had already taken place in Iceland before 2 Ma, and thus before the beginning of the Pleistocene (Albertsson, 1981; Einarsson and Albertsson, 1988).

The Plio–Pleistocene sequence of the Tjörnes peninsula in northeastern Iceland is unique in comprising considerable thicknesses of fossiliferous marine strata. The sequence begins below the base of the Gauss normal chronozone and continues upward into the Brunhes. Paleomagnetic data are plentiful, but reliable K-Ar ages are regrettably few, the basalts in the sequence being deficient in potassium and in many cases being altered as well. The stratigraphy of the relevant portion of the Tjörnes sequence is as follows, after Albertsson (1978, figure 2):

6. Máná basalts
5. Breidavík Beds (4 marine sedimentary beds, 4 tillites, 3 basalt flows)
4. Strangarhorn basalt
3. Furuvík Beds (1 marine sedimentary bed, 2 tillites, 1 basalt flow)
2. Furuvík lower basalt
1. Hvalvík basalt

A reasonably reliable age of 2.4 Ma has been obtained for the Hvalvík basalt, which has reversed remanent polarity and evidently lies within the lowest part of the Matuyama chronozone. A less reliable isochron age of 1.2 Ma for the base of the Máná basalts at the top of the sequence (Albertsson, 1978) suggests that the Neogene–Quaternary boundary may lie between those two levels. The lowest basalt of Furuvík is normally polarized, and thus could lie either within the Réunion subchronozone or, less probably, in the Olduvai subchronozone. The first interpretation is favored by Einarsson and Albertsson (1988), who show that it is consistent with the estimated average rate of accumulation in Tjörnes during the Matuyama and Brunhes chrons. It may be noted that the lowest basalt of Furuvík is directly overlain by the lower tillite of the Furuvík Beds, which not only is the oldest tillite in the Tjörnes succession but also appears to be the earliest evidence of sea-level glaciation in Iceland.

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16 The Neogene–Quaternary boundary in The Netherlands

WALDO H. ZAGWIJN

Introduction

The lithostratigraphy, biostratigraphy, and chronostratigraphy of Plio–Pleistocene deposits of The Netherlands can be related to paleoclimatic evidence for recognizing the Pliocene–Pleistocene boundary in this area. Such data have contributed to correlation between the stratigraphic sequence in The Netherlands and the proposed Neogene–Quaternary boundary-stratotype in the Vrica section of Calabria, Italy.

The Plio–Pleistocene sequences of strata in The Netherlands are apparently the most nearly complete within the southern North Sea basin, as they were deposited in an area of persistent strong subsidence. Moreover, interfingering of marine and nonmarine beds is common in this region.

The marine deposits contain microfauna that derived mostly from rather shallow water, with planktonic foraminifera being virtually absent, thus preventing direct correlations with the Mediterranean region or deep-sea sediments. The mollusk faunas are likewise quite different from those encountered in more southern regions of Europe. Mammal remains characteristic of the Villafranchian have been reported from some of the marine beds, especially from those of East Anglia, but again, those discoveries do not permit precise correlation with the Mediterranean area.

From about the middle part of the Pliocene strata upward, the marine faunas in the North Sea basin show reductions in the number of species, but increases in the numbers of cold-water specimens. During that time, periodic immigrations of boreal–arctic mollusks that today do not occur in the southern North Sea basin played an important role. The continental sequences, however, have provided the most detailed paleoclimatic and paleomagnetic information, which has led to a better understanding of the changes in the marine environment and to better correlations with the Mediterranean realm.

The Netherlands continental sequence

Since the beginning of the century, the Reuver Clay and the Tegelen Clay, two clay units of fluvial origin, have been recognized in the southeastern part of The Netherlands. The latter, in particular, has attracted much attention because of its abundant faunal remains (Dubois, 1905). In addition, both clay units have proved to be rich in fossilized fruits and seeds (Reid and Reid, 1907, 1915). From this evidence, a Quaternary age was proposed for the Tegelen Clay, a point of view initially disputed (Tesch, 1909; cf. Van Baren, 1920), but eventually accepted by all parties (Tesch, 1928, 1934).

The paleobotanical data, in particular, have contributed to clarification of the stratigraphy of these beds. The macroflora of the Reuver Clay is rich in species, about 50% of which are related to present-day floras in eastern Asia and North America, according to modern interpretations (Grichuk, Chapter 8, this volume). The flora from the Tegelen Clay also contains a number of extra-European exotics, although much fewer (i.e., hardly 15%). Moreover, quite a few modern species of Holarctic distribution, which are absent from the Reuver Clay flora, are common in the younger Tegelen flora (Reid and Reid, 1915; Zagwijn, 1959).

This change in floral composition can be traced through pollen analysis, as well as in macrofloral paleobotany. The palynological evidence shows that the Reuver Clay contains many taxa (*Sequoia, Taxodium, Sciadopitys, Nyssa, Liquidambar*, and others) that are absent from the Tegelen Clay. In the latter, only *Tsuga, Pterocarya, Carya, Eucommia*, and some other exotics that are relicts from the Upper Tertiary appear to have survived (Florschütz and Someren, 1950; Zagwijn, 1960, 1963).

These two paleofloras represent temperate, summer-rainfall associations, suggesting intervals of relatively warm climate, which have been termed *Reuverian* and *Tiglian*. Pollen analysis has shown that the marked changes in paleofloral composition between these two clay formations were caused by a severe cold-climate phase, the *Pretiglian*, during which all warmth-loving trees disappeared temporarily or permanently from the area, and the forest became more open, with the appearance of heaths and sphagnum bogs (Zagwijn, 1960). Although it has now been recognized that other cold-climate phases preceded that stage, none of them had such marked effects on the regional flora, nor did the forest actually open up to any notable extent. In other words, the Pretiglian was the first time during the late Cenozoic that the timberline was temporarily shifted south of The Netherlands.

After the Tiglian, which was a rather long interval with several alternating cooler and warmer phases (Zagwijn, 1963), a second phase of deforestation associated with an intense cold climate followed, the *Eburonian* cold stage. The Eburonian was followed by another relatively moderate interglacial complex with at least one phase of minor cooling, the *Waalian*, and subsequent to that came the *Menapian* cold stage, which is very well expressed in palynological data.

Following the Menapian, the two interglacials of the Bavelian age are characterized by occurrences of *Pterocarya, Carya, Eucommia, Tsuga*, and other trees which were also commonly present during the preceding Tiglian and Waalian temperate forest stages (Zagwijn and Doppert, 1978; de Jong, 1987). Beds representative of those interglacials are overlain by beds of the "Cromerian Complex," considered to be of middle Pleistocene age.

Paleomagnetic investigations. Studies by Van Montfrans (1971) and Boenigk, Koci, and Brunnacker (1979) have indicated that below the Matuyama–Brunhes boundary, which is in the lower part of the "Cromerian Complex," the reversed remanent magnetization of the Matuyama is predominant in the sediments down to the Reuver Clay. That clay has proved to be of normal polarity throughout, even where it is at its thickest, and therefore it is considered to date to the Gauss normal-polarity chron. Some consistent subzones of normal polarity can be observed in the predominantly reversed-polarity interval between the Reuverian and the "Cromerian Complex." One normal-polarity zone in the upper part of the Tiglian and the lowermost part of the Eburonian is identified with the Olduvai normal-polarity subchron, which means that the Eburonian cold stage began at about 1.8 Ma, in the present calibration. The normal-polarity intervals correlative with the Réunion normal subzones are in the lower part of the Tiglian (Van Montfrans, 1971; Urban, 1978). Further investigations by Boenigk et al. (1979) and Urban (1978) indicate that the Gauss–Matuyana boundary (2.60 Ma) is within the Reuverian C zone, the uppermost and coolest part of the Reuverian warm-climate paleoclimatic stage. This means that the Pretiglian cold stage can be dated between about 2.5 and 2.3 Ma (shown with older calibration in Figure 16.1), in synchrony with observations of late Pliocene cooling in the marine Piacenzian Stage of Italy (Preface, this volume; Azzaroli et al., Chapter 11, this volume).

Mammalian paleontology. The large-mammal fossils of the Tegelen Clay, the most significant of which are *Anancus arvernensis, Elephas (Mammuthus) meridionalis, Dicerorhinus etruscus, Equus robustus, Sus strozzii*, and *Leptobos*, have been recovered primarily from the upper part of the Tiglian (pollen zones TC5 and TC6). Some species, including *Anancus arvernensis*, have been found in beds dated as young as earliest Eburonian (Kortenbout van der Sluys and Zagwijn, 1962).

A rich fauna of small mammals has been found in beds of TC5 age (Freudenthal, Jeijer, and Van der Meulen, 1976). Biostratigraphically, the fauna of these beds, as well as those of earlier Tiglian age, belong to the *Mimomys pliocaenicus* zone or zone MN-17 of Mein (1975). In Eburonian deposits of the western Netherlands, *Allophaiomys pliocaenicus* and *Dicrostonyx* have been found (Van der Meulen and Zagwijn, 1974). This zone is considered to be indicative of earliest Pleistocene age in the view of many vertebrate paleontologists (Aguirre et al., Chapter 9, this volume; Chaline, Chapter 14, this volume).

The Netherlands marine sequence

In 1896, Harmer introduced the Amstelian Stage, based on a mollusk fauna considered to be younger than that of the Scaldisian in Belgium and older than that of the Icenian Norwich Crag. In the Amstelian, an immigration of boreal–arctic species is widely recognized, among which are *Serripes (Cardium) groenlandicus, Yoldia arctica*, and others. At that time, the Amstelian was considered to be of Upper Pliocene age. Tesch (1934), however, concluded that a marine fauna containing such boreal–arctic mollusks, underlying the "pre-glacial fluviatile beds" of the western Netherlands, should be included in the Pleistocene. He referred those deposits to "Icenian" age, while explicitly stating that the "true Amstelian" (because it was supposedly Pliocene) should be below those beds. Nevertheless, that usage clearly equated the greater part of Harmer's "Amstelian" with The Netherlands "Icenian," and that resulted in confusion as subsequent investigators continued to use "Amstelian" in a different way than originally intended (Figure 16.2). In view of this confusion, it is preferable to abandon the stage name Amstelian.

At present, four mollusk assemblage zones (Spaink, 1975) are identified in the Plio–Pleistocene marine beds of The Netherlands (in stratigraphic order, with oldest below):

Mollusk Zone A: Mya arenaria–Hydrobia ulvae assemblage zone. Beds of this zone are poor in species, and many of those species are presently living in the area under discussion. Influxes of land and fresh-water species, combined with shallowing marine conditions, characterized conditions in that interval.

Mollusk Zone B: Serripes groenlandicus–Yoldia lanceolata assemblage zone. This zonal assemblage contains many boreal–arctic species that are lacking in older zones; the two nominate species are predominant.

Mollusk Zone C: Nassarius propinquus–Lentidium complanatum assemblage zone. This zone is characterized by a very distinctive assemblage. Many species are abundant only in this zone, and several become extinct in the overlying zone.

Mollusk Zone D: Turritella triplicata–Yoldia semistriata assemblage zone. This zone is extremely rich in species, though relatively few of them are restricted to this zone.

According to present knowledge, Mollusk Zone B essentially coincides with the Amstelian, as defined by Harmer, though Harmer also included beds whose warmer-climate faunas we would assign to Mollusk Zone A.

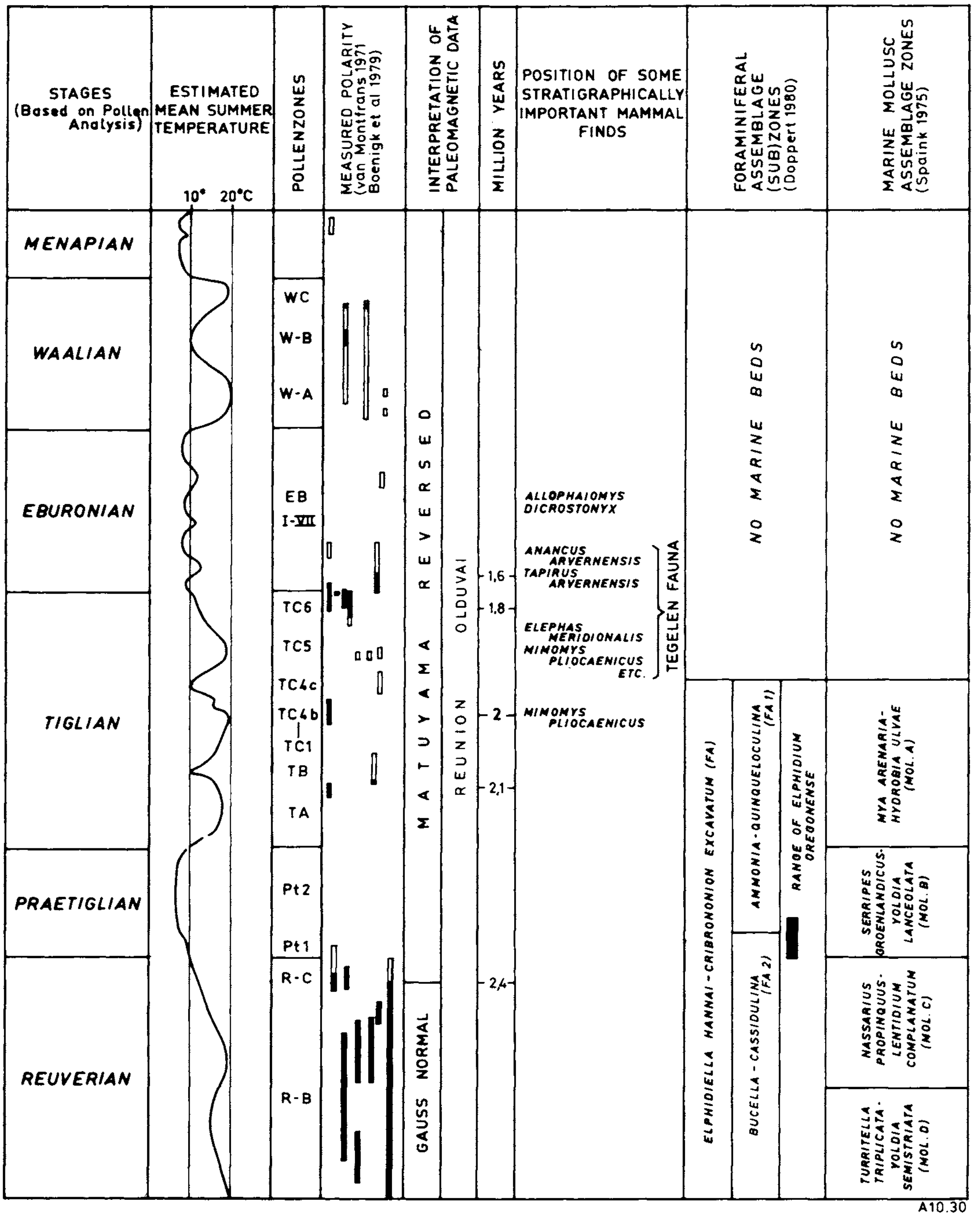

Figure 16.1. Marine and nonmarine stratigraphy of the southern part of the North Sea basin in the Upper Pliocene to Lower Pleistocene interval, from onshore sites in the Rhine and Scheldt deltas of The Netherlands. Note that the acme of the cold-adapted foraminiferan *Elphidium oregonense* in the Pretiglian interval corresponds to the first well-documented glacial influences on deep-sea deposits in the northern Atlantic, just above the Gauss–Matuyama transition. The level of the Pliocene–Pleistocene boundary at Vrica, at the top of the Olduvai normal-polarity interval, corresponds to the beginning of a later cycle of cold-climate conditions at the base of the Eburonian palynostage.

In the foraminifera assemblages, the first appearance of *Elphidiella hannai* (Cushman), which previously was suggested to represent the Pliocene–Pleistocene boundary (Lagaaij, 1952; Van Voorthuysen, 1957) or *Elphidiella hannai–Cribrononion excavatum* zone (Doppert, 1975, 1980), is definitely lower than the base of Mollusk Zone B (i.e., the Amstelian, *sensu* Harmer).

According to Doppert (1975), zone FA can be subdivided into two subzones, FA1 and FA2. The younger of these, FA1, is definitely poorer in species than FA2. The beginning of the narrow zone of the arctic species *Elphidium oregonense* in the middle part of zone FA straddles the FA1/FA2 boundary. That event has been proposed as the marine equivalent to the base of the palynologically established Pretiglian cold stage (Van Voorthuysen, Toering, and Zagwijn, 1972; Zagwijn, 1974a) and also correlates approximately with the base of Mollusk Zone B, but Van Voorthuysen et al. (1972) believed that the beds belonging to zone FA1 were of Pleistocene age.

Correlation of marine beds in the North Sea basin

In Figure 16.2, marine beds and biozones are compared for different parts of the North Sea basin. As stated earlier, recent

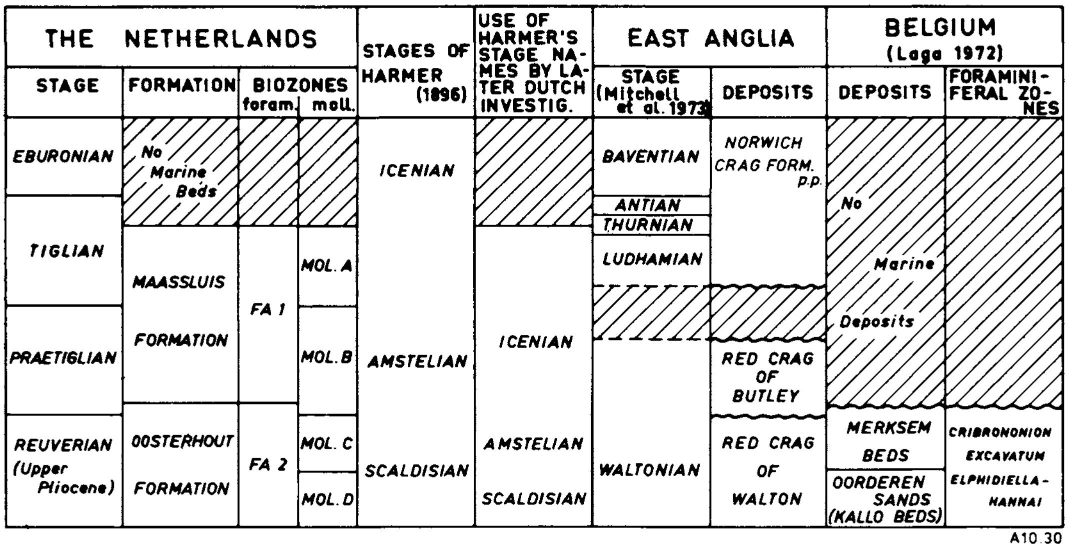

Figure 16.2. North Sea basin correlations. The Netherlands Plio–Pleistocene succession is compared with earlier stratigraphic concepts and with other successions in eastern England and northern Belgium. Note that the correlation of the East Anglian succession, shown here according to the state of knowledge in 1984, has since been revised (Gibbard et al., 1991) to move the Norwich Crag and the equivalent climate stages below the Eburonian, which is represented by an erosional interval (see also von der Brelie et al., Chapter 17, this volume).

investigators in The Netherlands have used Harmer's stage names of 1896, but in a somewhat different manner. Part of the Red Crag of East Anglia, at Butley, has a mollusk fauna characteristic of the *Serripes groenlandicus–Yoldia lanceolata* Mollusk Zone B of The Netherlands. A Pretiglian age, therefore, seems likely. Other Red Crag deposits, in particular the Walton Crag, are older (Harmer, 1896).

The Ludhamian and Baventian stages were established within the "Icenian Crag" according to pollen analysis (West, 1961), and each reflects a warmer-to-colder trend. According to Zagwijn (1974b), they can be correlated with the upper Tiglian (pollen zone TC3 to TC5) and the Eburonian (Spaink and Norton, 1967), rather than with the Tiglian, Eburonian, Waalian, and Menapian (West, 1961).

The Norwich Crag (Formation), which is another unit containing a major component of arctic mollusk species, according to present views dates to the Baventian and probably also to the older Ludhamian (West, 1980). It also includes deposits correlated to a younger cold stage, or stages, called Pre-Pastonian. Until recently, the accepted palynological correlations (Zagwijn, 1974a) indicated that the Ludhamian should be correlated to the middle Tiglian (Figure 16.2), so that the Baventian-age deposits seemed to be of the same age as the Eburonian. In a study that became available only while this chapter was being written, Gibbard et al. (1991) showed that the most reasonable correlation of the Eburonian to the East Anglian sequence is to the cold-climate deposits of Beestonian age, which unconformably overlie the typical Norwich Crag wherever they are in contact. In that view, the Baventian and Tiglian are considered equivalent.

The Belgian sequence, and particularly the foraminifera assemblages, were reexamined by Laga (1972), who established that there is good agreement with the sequence in The Netherlands. The palynological results (Hacquaert, 1961) clearly indicate that the Merksem Beds should be included into the topmost part of the Reuverian (C). The Oorderen Beds, formerly named Kallo Beds, are of Reuverian B age (Zagwijn, 1959; Hacquaert, 1961).

Correlation with the Mediterranean area and the Neogene–Quaternary boundary-stratotype

Palynological investigations by Suc (Cravatte and Suc, 1981; Suc and Cravatte, 1982) have revealed the existence of an Upper Neogene steppe-flora phase in the western Mediterranean, beginning just before the beginning of the *Globorotalia inflata* planktonic foraminiferal zone (2.5–2.4 Ma), and followed by a Mediterranean forest phase. Because Pleistocene cold-climate phases in more northerly parts of Europe clearly correlate with intervals of steppe conditions in the western Mediterranean, it can be considered certain that the first steppe phase of the western Mediterranean, coinciding with the lower part of the *G. inflata* zone, correlates with the Pretiglian of The Netherlands (Suc and Zagwijn, 1982). In both areas, the cyclic climatic changes of Pleistocene type evidently began in the lowermost Matuyama magnetozone, around 2.5 Ma, coincident with a marked cooling event in late Piacenzian time in the world oceans, and from a strictly paleoclimatic point of view it might be preferable to draw the lower boundary of the Pleistocene at that level (Zagwijn, 1992).

The boundary-stratotype section adopted at Vrica, however, was chosen according to a different view, which includes other stratigraphic criteria (Pasini and Colalongo, Chapter 2, this volume). The Neogene–Quaternary boundary at the top of marker *e* is definitely younger than the base of the *Globorotalia inflata* zone, which is more than 228 m below the proposed boundary in the Vrica section. If the paleomagnetic data are used for correlation, according to current evidence (Zijderveld et al., 1991) the proposed N/Q boundary is just below the top of the Olduvai normal magnetic subchronozone and is therefore coeval with the lower boundary of the Eburonian cold stage in The Netherlands.

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17 The Tertiary–Quaternary Boundary in western Germany

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Introduction

Deposits of late Pliocene and early Pleistocene age are found in several parts of western Germany (Figure 17.1), with marine strata principally in the northern coastal plain of northwestern Germany, and estuarine and nonmarine deposits in the Rhine Graben and other structural depressions in the North German Plain. Among these the marine record concerns mostly Pliocene deposits, whereas the lower Rhine Basin contains a lengthy and apparently continuous record of sedimentation during that time period. In the upper Rhine Graben and other inland areas, the Plio–Pleistocene record is not as comprehensive, being represented by localized and irregularly condensed sequences, which still afford some important correlation data.

The authors were responsible for the sections of this chapter as follows: K. Brunnacker, geological and paleomagnetic data; the late Professor H. Tobien, vertebrate paleontology; G. von der Brelie, palynology; W. Hinsch, marine paleontology.

Geological background

Northwestern Germany. In the coastal sequence, the stratigraphic units of most interest are the Kaolinite Sand (Weyl, Rein, and Teichmüller, 1955) and the Lieth series (Menke, 1975). Particular sections have been described from Oldenswort (Menke, 1975), Weser-Bergland (Benda et al., 1968), and Ostfriesland (Meyer, 1981).

Lower Rhine Basin. In this region, the stratigraphic sequence, divided into cyclic units of transgressive–regressive couplets, is denominated A/a to D/d, from oldest to youngest (Boenigk et al., 1974), and appears to correspond with the record from the Rhine delta in The Netherlands (Zagwijn, 1974; Zagwijn, Chapter 16, this volume).

The lower strata, unquestionably Pliocene in age, have been correlated according to molluscan faunas and palynology, analyzed petrographically, and inserted into the paleomagnetic scale by Boenigk and co-workers (Boenigk et al., 1974; Boenigk, 1978a; Boenigk, Koci, and Brunnacker, 1979). Two heavy-mineral suites are clearly distinguished (Table 17.1). The deposits in the A1 beds (clay horizon A of the Ville district, as well as the overlying peat layer) contain a suite of stable heavy minerals, such as tourmaline, zircon, and staurolite, and correlate with the upper Reuverian (Reuverian B–C) of The Netherlands climatostratigraphic sequence (Zagwijn, 1974). The younger A2 clays are characterized by unstable heavy minerals (epidote, garnet, green hornblende, and alterite). An important molluscan fauna has been found in the A2 horizon (Boenigk et al., 1974), which clearly represents a Pliocene assemblage. The A2 horizon has normal paleomagnetic polarities in the lower part and reversed polarities in the upper part.

The succeeding stratum, termed the Oldest Pleistocene in historical literature, is the b1 Gravel (Figure 17.2), the lithology of which resembles the Pliocene quartz gravels, or *Kieseloolithschotter* (Schnütgen and Brunnacker, 1977). The B1 Clay follows above, with evidence for relatively warmer conditions described as the Fortuna Oscillation by Urban (1978). The upper part of this sequence records a change back to normal paleomagnetic polarity. At the top of this sequence, a major erosional surface is the main discordance in the Plio–Pleistocene sequence of this area.

The overlying b2 Gravel (Figure 17.2) has the more variegated appearance of the typical Quaternary “colored gravels” (Schnütgen and Brunnacker, 1977). Nevertheless, both the b1 and b2 gravels are assigned to the Pretiglian cold-climate interval. The overlying B2 Clay contains a “Quaternary” (i.e., modernized) warm-climate molluscan fauna (Lozek, in Boenigk, Kowalczyk, and Brunnacker, 1972) and evidence of a corresponding warmer-climate flora, including fruits and seeds (E. K. Kempf, in Boenigk et al., 1972). The presence of *Fagus* (beech) pollen indicates the Tiglian-A warm period (Urban, 1978). The B2 Clay shows reversed paleomagnetic polarity.

Above that level, the variegated quartz-poor c Gravel is succeeded by the C Clay with another thermophilic molluscan fauna and flora (Lozek, in Boenigk et al., 1972; Urban, 1978). This clay also shows reverse magnetization (A. Koci, in Boenigk et al., 1979).

In the d Gravel horizon, the relatively sand-rich deposits contain the earliest glacially striated erratics. In the overlying D Clay, the heavy-mineral association is similar to that of the B

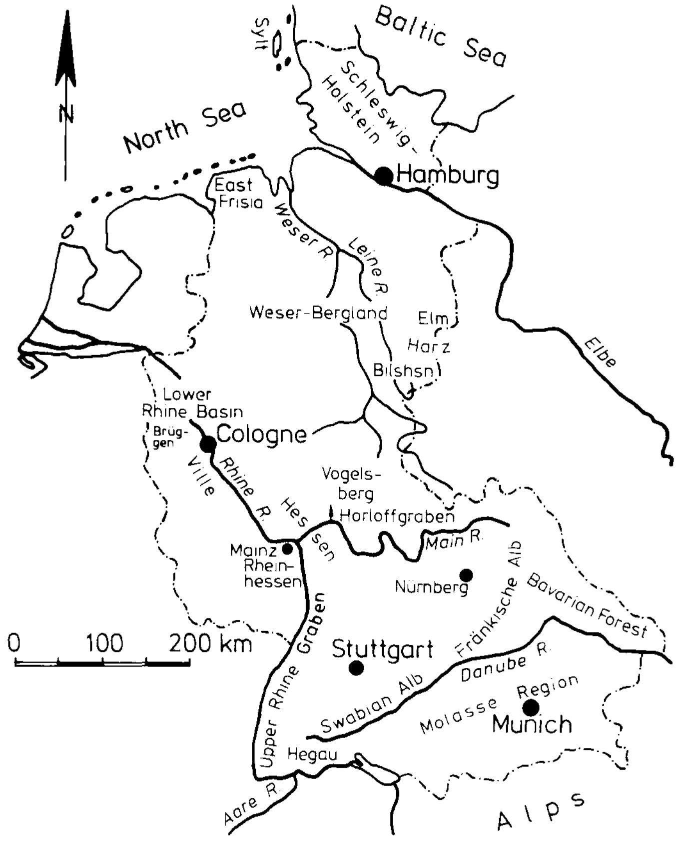

Figure 17.1. Geologic provinces of western Germany.

horizon, and the molluscan fauna is still of the warm "Quaternary" nature (Lozek, in Boenigk et al., 1972); the paleoflora is noteworthy for the presence of some holdover "Tertiary" elements (Kempf, in Boenigk et al., 1972; Urban, 1978). Samples of the D Clay from the Ville sequence have shown normal paleomagnetic polarities.

Petrographic analyses (Schnütgen, 1974; Boenigk, 1978b) provided the basis for a separate classification of the gravel layers as low-stand terraces, or "main terraces," classified under the terms HT1 through HT4. The molluscan fauna of the Upper Pliocene in the B1 Clay is sharply different from the warm-climate fauna in the B2 Clay horizon, which according to pollen analysis (Urban, 1978) can be correlated to the lowest horizon of the Tiglian complex in The Netherlands sequence (Zagwijn, 1960, 1974; Zagwijn, Chapter 16, this volume).

Vogelsberg. According to geobotanical results, the Liegendton ("under-clay") and the lignites of the Wohnbach mining area, situated in the Horloff Graben, belong to the lower part of the Upper Pliocene (either uppermost Brunssumian or Reuverian A) and therefore are younger than earlier assumed (Boenigk et al., 1977). The mammal fauna at Wölfersheim, found at the top of the Liegendton, is classified as late Ruscinian, MN-15 (as discussed later), and is therefore older than the lower Villafranchian (e.g., Étouaires) and younger than the typical Ruscinian (e.g., Montpellier). Evidence for several alternating cold- and warm-climate periods has been found in the overlying clay-silt layers, the earliest of which corresponds to the normally magnetized Pretiglian horizons (Boenigk et al., 1974).

Table 17.1. *Stratigraphic correlation of the Pliocene and Quaternary in western Germany*

Pleistocene	**Toringian**	**MQ2**	
	Biharian	**MQ1**	Körlich E and F Hohen-Sülzen Weissenburg 7 Neuleiningen 15
Pliocene	**Late Villanyian**	**MN17**	Gundersheim 2 Erpfingen 2 Moggaster Höhle Schernfeld Deinsdorf, Schumbach
	Early Villanyian	**MN16**	Frechen
	Late Ruscinian	**MN15**	Gundersheim 3, 4
	Early Ruscinian	**MN14**	Herbolzheim
Late Miocene	**Turolian**	**MN13**	

Other inland sections. The Upper Pliocene and Lower Pleistocene strata have also been studied in Nordhessen Buchenau (Leschick, 1951), in the Kelsterbach Terrace complex of the Rhine/Main area and the Upper Rhine Graben (Semmel, 1974; von der Brelie, 1974), and in other deposits of the Upper Rhine Graben (Bartz, 1976, 1982). The age of the Steinbach deposits has been discussed by Kolumbe (1963) and Frenzel (1973), and that of Jockgrim by Guenther and Mai (1977), Peters (1965), and Koci, Schrimer, and Brunnacker (1973).

Biostratigraphy

Mammalian faunas

Mammalian local faunas in the interval of the Pliocene–Pleistocene boundary are categorized in the mammal ages Ruscinian, Villanyan, and Biharian in the current biostratigraphic concept for central and western Europe. The Villafranchian continental stage corresponds, more or less, to the Villanyan and Biharian together, as discussed later.

In this concept, the Ruscinian and Villanyan represent the Pliocene, and the Biharian the lower part of the Pleistocene. The Pliocene–Pleistocene boundary is drawn between the Villanyan and the Biharian at a level close to the top of the Olduvai normal-polarity subchron, corresponding to the decision of the Neogene–Quaternary Boundary Commission (Fejfar and Heinrich, 1989), with an age of 1.81 Ma (Pasini and Colalongo, Chapter 2, this volume). The biochronological subdivisions of these ages are based on the mammal zones proposed by Mein (1989, p. 73), in which the Ruscinian is equivalent to Mammalian Neogene (MN) zones MN-14a, -14b, -15a, and -15b, and Villanyan to MN-16a, -16b, and -17. The Biharian is equivalent to the Mammalian Quaternary (MQ) zone MQ-1. The terminal Miocene faunas are assigned to the Turolian mammal age, zone MN-13 (Table 17.1). For details, see Fejfar and Heinrich (1989, figures 1 and 2).

Table 17.2 shows the classic Villafranchian Stage of the Italian continental sequence in place of the mammal ages Villanyan and Biharian. Correlation between the Villafranchian subdivisions in Italy and these mammal ages, which are typified mainly in central and eastern European faunas, is obviously not perfect, given the present state of knowledge. According to Masini and Torre (1989), the Italian Plio–Pleistocene biostratigraphy should be compared with that of the regions to the north as seen in Table 17.2.

Ruscinian. The Herbolzheim local fauna, of early Ruscinian age, has long been known from a fissure filling in Dogger limestones north of Freiburg im Breisgau, at the eastern border of the Rhine Graben (Figure 17.1) (Tobien, 1951). The association includes 11 small and large taxa, among which are *Anancus arvernensis, "Sus" minor, Canis donnezani, Sciurus* n. sp., *Vulpes* sp., and *Lutra* n. sp., which indicate MN-14 (early Ruscinian) age (Mein, 1989). The riverside forest environment indicated by *Sciurus, Lutra*, and a suid is similar to the typical Montpellier asssemblage itself (Tobien, 1975), although Herbolzheim was listed erroneously under "Csarnotian" or late Ruscinian.

Table 17.2. *Succession of mammalian local faunas in western Germany*

Biharian	Upper Villafranchian: Farneta, Tasso, Olivola faunas (= MQ-1)
----------------- N/Q boundary -----------------	
Late Villanyan	Upper Middle Villafranchian: Saint-Vallier fauna (= MN-17)
Middle Villanyan	Lower Middle Villafranchian: Montopoli fauna (MN-16b)
Early Villanyan	Lower Villafranchian: Triversa fauna (MN-16a)
Late Ruscinian	Perpignan (Csarnotian): Serrat d'en Vaquer (= MN-15)

In Rheinhessen (Figure 17.1), east of Eppelsheim, fissure fillings in Upper Oligocene limestones at Gundersheim (Heller, 1936) have yielded two local faunas of small mammals of different ages. The older is known from an isolated block from the quarry floor, Gundersheim-Findling (Storch and Fejfar, 1989) or Gundersheim 4 (Fejfar and Storch, 1990), and also from samples in "Fissure nr. 4" at Gundersheim 3 (von Koenigswald and Tobien, 1990, p. 234). The Findling block has produced a rich micromammal fauna with 27 taxa, correlated to MN-15b (late Ruscinian) age with eastern European and Siberian affinities.

A similar attribution is indicated for the rich collection from Wölfersheim in the Vogelsberg area, which contains large and small mammals (tapirs, several deer, *Dicerorhinus, Anancus, Mammut borsoni, Sus minor*, and several beavers, as well as microtines such as *Mimomys occitanus* and *M. gracilis*, sciurids, cynomorphs, and insectivores, among them *Desmana*). The assemblage points to a swampy forested milieu (Tobien, in Boenigk et al., 1977, p. 65; von Koenigswald and Tobien, 1990, p. 235). As noted earlier (Table 17.2), the palynological age of the Wölfersheim bone-bearing bed in the upper Liegendton is uppermost Brunssumian or Reuverian A, whereas the late Ruscinian mammal faunas of Gundersheim 1 and Wölfersheim belong to zone MN-15 (Fejfar and Heinrich, 1989; Mein, 1989, p. 79).

Villanyan. An early Villanyan level, characterized by the MN-16 marker taxon *Mimomys polonicus*, is represented in river deposits on top of lignites in the open-pit quarries of Frechen, near Köln (Figure 17.1). The fossiliferous level is close to the Gauss–Matuyama paleomagnetic reversal at 2.48 Ma (Preface, this volume). According to palynological analysis, the lignite deposit corresponds to Reuverian B (Van Kolfschoten and Van der Meulen, 1986; von Koenigswald and Tobien, 1990). The boundary between Reuverian A and B may thus be correlated to the Ruscinian–Villanyan boundary (Table 17.2), of later Pliocene age.

Numerous late Villanyan sites are known. At Gundersheim, yet another micromammal assemblage, initially described by Heller

(1936) and later separated into two units, Gundersheim 1 and 2, by Kretzoi (1962), has been revised by Storch and Fejfar (1989). There are two views on the age of this assemblage: either it is an early MN-17 fauna, with persistent primitive arvicolids, or it is a mixture from two different levels (MN-16a and MN-17). In any event, the former age assignment of late Ruscinian (MN-15) age for Gundersheim 1 (Kretzoi, 1962; Tobien, 1980) is invalid. Fejfar and Heinrich (1989, figure 2) have placed the entire assemblage, as "Gundersheim 2," in MN-17 (i.e., late Villanyan).

A cave deposit at the entrance of the Bären Höhle near Erpfingen, 50 km south of Stuttgart in the Swabian Alps, is termed Erpfingen 2, not to be confounded with the nearby, younger fissure fillings Erpfingen 1 and 3 of Heller (1958). This site has produced large and small mammals that document a late Villanyan age (Mein, 1989; von Koenigswald and Tobien, 1990). The assemblage is closely comparable to the late Villanyan Tegelen local mammal fauna in The Netherlands (Zagwijn, Chapter 16, this volume). Other late Villanyan, MN-17 local faunas come from fissure fillings in Upper Jurassic limestones at Moggaster Höhle, Schernfeld, Deinsdorf, and Schambach in the Franconian Alb (Figure 17.1) (Fejfar and Heinrich, 1989).

Present consensus identifies the Eburonian cold-climate phase as the equivalent of the earliest Pleistocene, as defined in the Vrica boundary. The late Villanyan local faunas of Gundersheim and the Bären Höhle, being of the same age as the Tegelen local fauna and its accompanying Tiglian warm flora underlying the Eburonian levels, would therefore represent the last warm-climate interval before the beginning of the Quaternary.

Biharian. Whereas the Villanyan local faunas are dominated by the arvicolid *Mimomys*, in the Biharian the first species of the advanced arvicolid *Microtus* begin to appear in northern European mammal communities, together with related forms. In western Germany the most significant sites of this age are the fissure fillings of Weissburg 7 in Jurassic limestone of the Swabian Alb (Figure 17.1) and Neuleiningen 15 (Fejfar and Heinrich, 1989; listed as Neuleiningen 11) in Aquitanian limestones near Worms, Rheinpfalz (Table 17.1).

Whereas the MN-17 fissure fillings with their warm-climate faunas represent the latest Pliocene (Villanyan) levels, these MQ-1 cold-climate faunas clearly represent the early Biharian, and their correlation with the Eburonian (Fejfar and Heinrich, 1989, p. 103) identifies them with the earliest Pleistocene. The Eburonian of the lower Rhine Basin is represented in the deposits of the "Hauptterrase 1" or "main terrace 1" (Figure 17.2), so it can be assumed that the entire sequence below the Hauptterrase 1 from Tonhorizon ("clay horizon") D down to Tonhorizon A1 are of pre-Biharian or Villanyan, late Pliocene age. The relationship to the Olduvai paleomagnetic event is obvious (Figure 17.2).

Among later Biharian sites is Hohen-Sülzen, near Worms (Rheinhessen, Table 17.1), where the gastropod marls contain mammalian fossils, *inter alia* at least 15 microtines, including three relict species of *Mimomys* (von Koenigswald and Tobien, 1990). Beds E and F in the loess section of Kärlich in the Nieuwied basin near Koblenz, between Mainz and Cologne (Figure 17.1), lie above the Matuyama–Brunhes paleomagnetic reversal, at 0.78 Ma (Preface, this volume). The fauna of these levels indicates a late Biharian age (von Koenigswald and Tobien, 1990). The younger faunas (i.e., Toringian of Fejfar and Heinrich, 1989) are beyond the scope of this chapter.

*Flora**

Pollen and spores are of particular importance for stratigraphical classification of the Pliocene and early Pleistocene, and thus as well for determination of the boundary between the Tertiary and Quaternary. Microflora can be found in almost all organogenic deposits, and therefore many localities are known; macroflora, on the other hand, are more seldom found. Zagwijn (1960, 1963, 1974) divided the quite considerable Upper Miocene, Pliocene, and early Pleistocene sequence in The Netherlands into seven named units based on the changes in the palynoflora: from bottom to top, Susterian; Brunssumian (with subunits A, B, C); Reuverian (with subunits A, B, C); Pretiglian; Tiglian (with subunits A, B, C); Eburonian; and Waalian. Zagwijn's classification serves today as the basis for all pollen-analysis research into the question of the Tertiary–Quaternary boundary in central Europe.

As for the fossil leaf and fruit floras in western Germany, there are few sites. In the Reuverian there are Massenhausen near Munich (Jung, 1963), Frankfurt am Main (Mädler, 1939), and Willershausen (Straus, 1967). For Tiglian macroflora, the best-known site is Schwanheim, near Frankfurt (Baas, 1932). The best assemblages are those from the clays of Reuver and Tegel themselves, situated in The Netherlands (Zagwijn, Chapter 16, this volume).

In regard to the microflora, for several decades the question of the Tertiary–Quaternary boundary has been linked with debates about the significance of major climate changes evidenced in the paleofloras of the lands bordering the North Sea basin. Terrestrial deposits of the Reuverian Pliocene are always easily identifiable, with a dominance of distinctively Tertiary floral elements such as *Sciadopitys*, cf. *Sequoia*, Taxodiacea, *Tsuga, Nyssa, Liquidambar, Carya, Pterocarya, Castanea, Eucommia*, Symplocaceae (*Symplocoipollenites rotundus* and *S. vestibulum*), and the form-species *Cupuliferoidaepollenites quisqualis, C. fallax, Tricolporopollenites exactus*, and *Araliaceoipollenites edmundi*. By contrast, the warm periods of the succeeding paleofloral stages (from Tiglian to Waalian) are characterized by microflora in which only a few typical Tertiary elements are to be found, such as *Tsuga, Carya, Pterocarya, Ostrya, Castanea, Philodendron*, and *Eucommia*. The decimation of the flora during the Pretiglian cold period, as Zagwijn (1960) showed, involved not only the disappearance of trees adapted to warm-climate conditions but also a marked thinning of forest density. Because of this marked change, the base of the Pretiglian was

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*This contribution reflects the state of knowledge up until 1982. The author regrets that illness has prevented a thorough revision to include more recent data.

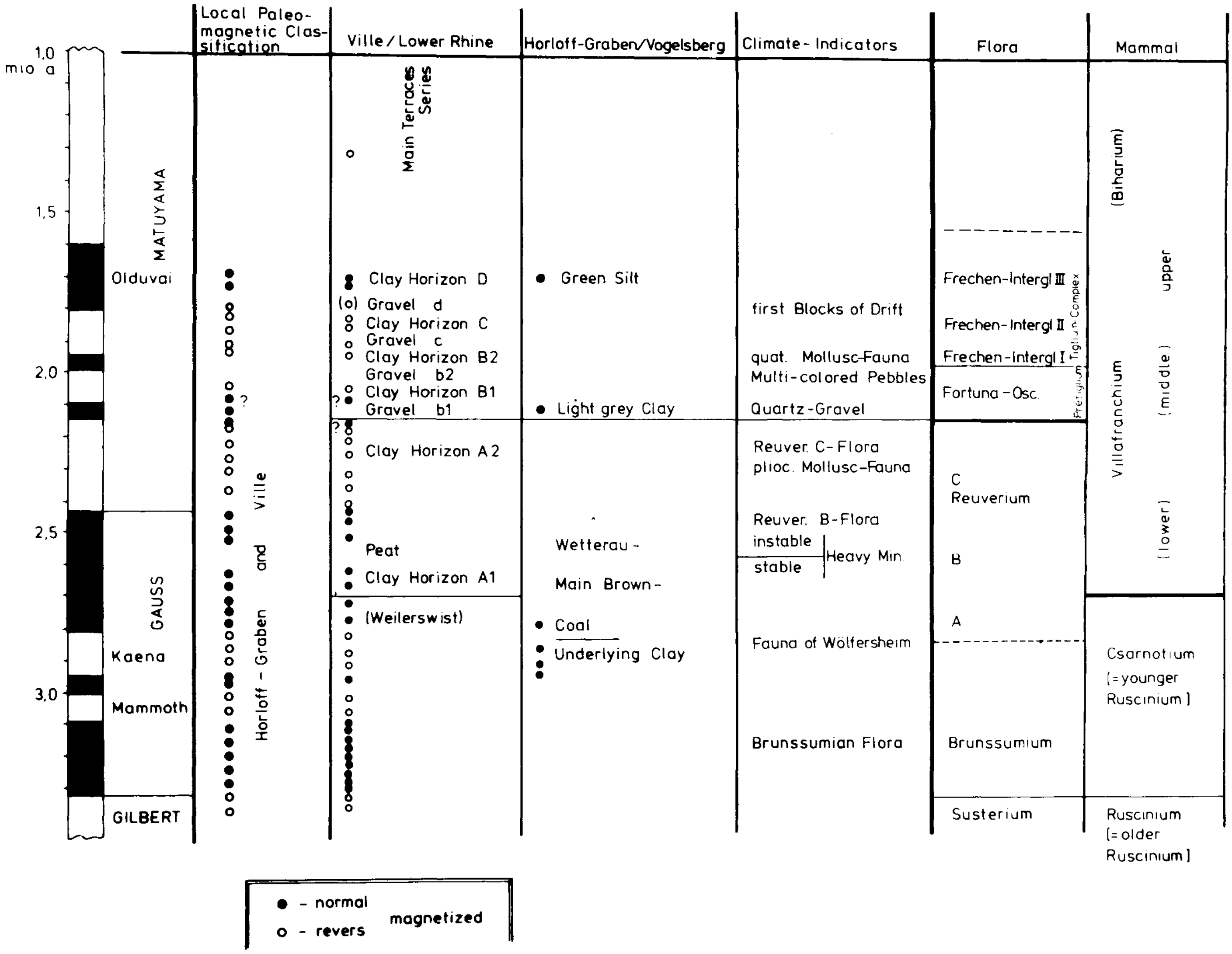

Figure 17.2. Stratigraphy in the terrestrial Tertiary–Quaternary boundary interval of western Germany

long used in this area as the equivalent of the Tertiary–Quaternary boundary (Zagwijn, 1974).

Important sections with microflora of Plio–Pleistocene age are found principally in the coastal regions, such as East Friesland (Meyer, 1981) and Schleswig-Holstein (Menke, 1975), and in structural depressions such as the lower Rhine Basin, the Leinetal Graben (Benda and Lüttig, 1968), the Rhein-Hesse depression (including the Horloff Graben, the Seligenstädter depression, and the Rhine/Main area), and the Upper Rhine Graben.

Coastal-plain palynofloras. The most important and best-studied sequence is at Lieth near Hamburg, in Schleswig-Holstein, where Menke (1975) described a sequence of paleofloras indicating five warm-climate periods separated by layers indicating cold-climate periods. From bottom to top, the warm-climate periods are named Nordende, Ellerhoop, Tornesch, Uetersen, and Pinneberg. Correlation of the Tornesch and the Waalian paleoflora seems very likely, and the early warm-climate periods correspond to the Tiglian complex. In The Netherlands and in the lower Rhine Basin, there are as yet no known parallels to Menke's youngest two warm-climate periods. Core-drilling at Oldenswort 9 showed Lieth-series sediments overlying the Reuverian and older Pliocene sections (Menke, 1975).

Lower Rhine palynofloras. It has been known for some time that in the lower Rhine area there is a very thick and important sequence of Pliocene and Pleistocene deposits (Van der Vlerk and Florschütz, 1953). The problem of the Tertiary–Quaternary transition in the vegetational history has been discussed by Urban (1978), who investigated profiles in the open-pit lignite mines on the Ville at Fortuna, Garsdorf, and Frechen, as well as in clay pits in the Brüggen area near the German/Dutch border, and by von der Brelie (1981), who analyzed samples from boreholes drilled in the lower Rhine Basin. Urban discovered that the Frechen I interglacial, characterized by a *Fagus–Tsuga* association, is equivalent to Tiglian A, and in the Peter van Eyck clay pit near Brüggen she found a previously unknown interglacial which she correlated with the Tiglian C. A Waalian correlative may also be present in the latter section. On the basis of the borehole evidence, clear parallels can be found between the Tiglian and younger microfloras in the German and Dutch parts of the lower Rhine Basin. The parallels in the Reuverian microfloras are also in good agreement (von der Brelie, Hager, and Kothen, 1981).

Paleofloras in other structural depressions. The Tertiary and Quaternary sediments in the Horloff Graben are of great importance (Boenigk et al., 1977). At Wölfersheim, a well-known middle Pliocene (Csarnotian) mammalian fauna has been collected from beneath the lignites, as discussed earlier. The lignites, formerly thought to be of Brunssumian age, must now be assigned a stratigraphic position of Reuverian A. In the lower Rhine area the pollen analytical investigations of von der Brelie et al. (1981) have shown that high proportions of *Sequoia*-like pollen, previously thought to be an indicator of Brunssumian floras, may also occur in the Reuverian A horizon. The lignites of the Wohnbach open pit therefore can be dated as Reuverian A and B, which is in better agreement with the mammalian biochronology.

The Schwanheim site in the Rhine/Main area is well known (Baas, 1932). Evidence for at least two warm periods of Tiglian aspect can be found there (von der Brelie, 1974; Semmel, 1974), and clays and paleosols with similar pollen assemblages can be found in many places intercalated with fluvial deposits in the northern and middle parts of the Upper Rhine Graben (von der Brelie, in Bartz, 1976, 1982). The relics of some Pliocene pollen forms allow these palynofloras to be distinguished from later deposits, but attempts to define a more exact biostratigraphy have been unsuccessful because of the discontinuity of exposures. Even the well-known clay deposits of Jockgrim and Rheinzabern, although generally thought to belong to the early Quaternary in the northern European climatic sequence (i.e., post-Pretiglian), have not been satisfactorily dated (Peters, 1965). Recent palynological investigations, together with regional geological considerations (Bartz, 1982), indicate that Jockgrim is almost certainly younger than the Tiglian.

The deposits from Uhlenberg, near Augsburg in southern Germany, remain to be mentioned (Scheuenpflug, 1979; Schedler, 1979). A Tiglian age is indicated by the occurrences of *Tsuga, Carya, Pterocarya*, and *Castanea*.

Summary. Knowledge of the late Pliocene and early Pleistocene plaeofloras in the southern part of western Germany remains fragmentary, and most sites cannot be exactly dated according to available geological or palynological data (Frenzel, 1973). Until now, it is only in Schleswig-Holstein and the lower Rhine area that adequate information exists to identify the Tertiary–Quaternary boundary, whether at the base of the Pretiglian or at a higher level, such as the base of Eburonian cold-climate phase (Zagwijn, Chapter 16, this volume). In the Upper Rhine Graben (Bartz, 1976, 1982) and in the Horloff Graben (Gruschkau, 1962; Boenigk et al., 1977) the deposits of that interval are generally very poorly exposed, and because of slow and erratic tectonic subsidence rates they have a complicated sedimentation geometry; the stratigraphy, therefore, is almost impossible to correlate. It must also be taken into consideration that the deposits usually are somewhat condensed and normally exhibit relatively thin interglacial vegetation phases.

Continental molluscan faunas

The upper Reuverian molluscan fauna in the A horizons of the Ville shows a distinct similarity to the molluscan assemblage of Hauterives, a later Ruscinian (medial Pliocene) site in southern France (Strauch and Schlickum, in Boenigk et al., 1974; Mein, 1989). In the past, Reuverian flora were consistently classified as Upper Pliocene, but the present mid-Pliocene attribution is more in accord with the molluscan evidence. The mollusks in the B2 Clay horizon from the Ville demonstrate an astonishing faunal change, similar to the paleofloral change at that level, as noted earlier. According to Lozek (in Boenigk et al., 1972) the fauna exhibits distinct warm-climate elements, but without any of the earlier warm-climate taxa of the Reuverian. This may be seen as an indication of an unprecedented severity of climate change and a consequent high rate of extinction in northern European molluscan faunas during the intervening cold-climate episodes represented by gravels b1 and b2; as noted earlier, pollen analysis equates these gravels with the Pretiglian, and the B2 Clay with the Tiglian.

Paleomagnetism

The samples analyzed for paleomagnetism (Figure 17.2) from the upper Brunssumian up to and including the Reuverian C horizons show normal magnetization, attributed to the Gauss normal chron. The polarity reversal in the uppermost Reuverian C (i.e., in the A2 Clay of the Ville, in the lower Rhine Basin) is therefore the beginning of the Matuyama reversed-polarity chron. Considering the paleomagnetic sequence in the overlying Pretiglian and Tiglian in The Netherlands (Zagwijn, Chapter 16, this volume), it may be that the normal polarities of the Olduvai subchron in the upper Tiglian are those recorded in the D Clay of the lower Rhine.

Marine younger Neogene

Pre-Pleistocene sedimentation. The main part of the Neogene (Table 17.3) can be divided into three sedimentation cycles in this region: (1) an early Miocene cycle (23–17 Ma), with subsidence in the Vierlandian (*tmiv*) and a regression, with

Table 17.3. *Stratigraphic correlation of the Neogene and early Quaternary in the North Sea margins*

Epoch	**Climate Stage**	**N'WEST GERMANY, JUTLAND**		**BELGIUM**	**NETHER-LANDS**	**EAST ANGLIA**	**WESTERN FRANCE**
PLEIST.	**Waalian**	▨		▨	Waalian	Beestonian	▨
	Eburonian	▨		▨	Eburonian	▨	▨
PLIOCENE	**Tiglian**	D Clay C Clay B2 Clay	Oldesloe Formation *Bredstedt / Garding / Reuver / Lieth*	▨	Tiglian	Baventian Thurnian Ludhamian	▨
	Praetiglian	(Lieth Series)		▨	Mol B (Praetigilian)	▨	▨
	Reuverian	Scaldisian		Merxemian	Mol C (u. Reuver)	Orwellian Waltonian	▨
				Scaldisian	Mol D 1	▨ Gedgravian	(Marnes à Nassa)
	Brunsummian	Morsumian		Kattendijkian	Mol D 2	▨	Redon. 3 - 4
		Syltian IV Syltian III		▨	Mol E (Delden) Mol F 1	▨	Redonian 1 - 2
MIOCENE	**Susterian**	Syltian II Syltian I		Diestian s.s.		▨	▨
		Gramian II Gramian I		Deurnian	Mol F 2 (Zenderen)	Lenhamian	▨
		Langenfeld. IV		▨	Mol F 3 - 5 (Eibergen)	▨	▨
	(Middle)	Langenfeldian III Langenfeldian II Langenfeldian I		▨		▨	▨
		Reinbeckian		▨	Mol G (Stemerdink)	▨	Pontilévian Savignéan
	(Early)	Oxlundian Hemmoorian Behrendorfian		Anversian Houthalian Edegemian	Mol H (Miste)	▨	Burdigalian
		Vierlandian		▨	▨	▨	Aquitanian
OLIGOCENE		Neochattian		▨	▨	▨	▨

fluviatile sands from the east, in the Hemmoorian (*tmih*); (2) a later Miocene Elbe cycle (17–5 Ma), with Reinbekian subsidence (*tmir*) reaching a maximum at the *tmir*/*tmil* boundary (Levensau and Lüneburg subzones, Tostedt Member and Eibergen Member), and with regression after the late Langenfeldian (*tmila*).

In regard to the third cycle, the paleogeography and biofacies distributions of the succeeding Gramian and Syltian are discussed in Hinsch (1990). During the Syltian, marine strata of the younger Syltian and Morsumian were restricted to the west (e.g., the Morsum Cliff on Sylt, and Wursterheide, near Cuxhaven) by prograding fluviatile sedimentation of the Oldesloe Formation. The surface exposures at Morsum Cliff (Hinsch, 1984, 1985), which have been deformed by glacial tectonics, have been correlated to undisturbed well sections on Sylt by Hinsch (1990). There and at Wursterheide (Hinsch, 1989) the Syltian–Morsumian boundary is observed in marine strata, in which the Syltian molluscan fauna is very rich (Hinsch, 1977) and the

Morsumian is rather poor. Well-preserved calcareous molluscan fossils of both Syltian and Morsumian age were noted (Hinsch, 1991) in the offshore well R-1 (55°00'N, 6°49'E). The Morsumian fauna of R-1, with *Spisula arcuata, Cingula inusitata, Telasco syltensis, Uzita serrata, Fusiturris helena, Mangelia nysti*, and *Philbertia perpulchra*, can be correlated to the Kattendijkian.

Later Pliocene (Scaldisian) faunas have been found in offshore wells: 89/5 (54°07'N, 2°19'E), with the *Acila–Astarte* association overlain by beds with the *Cerastroderma hostiei* biofacies, and H15/2, west of Helgoland (54°10'N, 7°E), which records the first appearance of *Dosinia imbricata*.

An Icenian, or middle Tiglian, molluscan fauna of the Bramertonian biofacies, with *Acila cobboldiae*, was also found in the 89/5 borehole. The Bridlingtonian biofacies, with *Tridonta borealis withami*, also probably of Icenian age, was reported by Hinsch (1991) from well 89/3 (54°27'N, 5°47'E).

The base of the Quaternary, or *tpl/qp* boundary, was assumed by Lagaaij (1952) and Glibert and de Heinzelin (1955), following Wirtz (1949), to be located above the Morsumian at the base of the Merxemian-Waltonian. This is equivalent to a level within the Reuverian. Zagwijn (1974), considering paleoclimatic evidence, proposed that the boundary be placed instead between the Reuverian and Pretiglian, or equivalent to the top of the Merxemian-Waltonian. Finally, the definition in the Vrica stratotype in Calabria would result in a boundary between Tiglian and Eburonian pollen-floral stages (Zagwijn, Chapter 16, this volume). Neither of the latter two boundary definitions can be represented by any known marine faunas from onshore exposures in northwestern Germany.

On the other hand, in the western part of the southern North Sea, the Pliocene and most Pleistocene interglacials are represented by marine horizons with abundant molluscan fauna. With regard to the approved new definition of the Pliocene–Pleistocene boundary, the horizon which corresponds to the Tiglian–Eburonian boundary can be recognized by the disappearance of *Acila cobboldiae* in the North Sea record. Reworked shells from the Tiglian, as well as underlying Pliocene strata, can, however, be found in the transgressive interglacials, as, for example, in basal Eemian sediments in borehole 89/2 (54°N, 5°E), as well as in borehole 89/4.

Conclusions

Floral and mammalian biostratigraphic data from the Ville and from Vogelsberg related to the Tertiary–Quaternary boundary are synthesized in Figure 17.2. The Ruscinian–Villanyan boundary is correlated with the boundary between Reuverian A and B. According to the integration of the floral sequence into the paleomagnetic scale, the Reuverian–Pretiglian boundary, with the beginning of severe cold-climate conditions seen in many stratigraphic sequences of western Germany, would be approximately 2.5 Ma, close to the Gauss–Matuyama boundary. The next major refrigeration, at the Tiglian–Eburonian boundary, dating to the top of the Olduvai normal-polarity subchron, is represented in the mammalian biostratigraphy at a level between the Gundersheim 2 Villanyan level and the transitional Villanyan–Biharian assemblages found in the Neuleiningen, Schernfeld, and Deinsdorf local faunas. In marine molluscan faunas, that level can be equated with the local extinction of *Acila cobboldiae* in offshore boreholes. On the other hand, in terms of the molluscan paleontology in onshore marine deposits, as well as for palynology and paleomagnetic data, the Tiglian–Eburonian boundary, which presently equates to the Pliocene–Pleistocene boundary, is not well represented in the geologic record of this area.

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18 The Pliocene–Pleistocene boundary in eastern Germany

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Introduction and geological background

In the late Neogene, the sea progressively retreated from wide areas of the northwestern lowlands of the North German–Polish coastal basin. Large river systems that developed in the exposed areas drained generally toward the west, laying down great amounts of fluviatile-limnic sediments. The lithology of pebbles contained in these sediments indicates that the major source area was in what is now the eastern Baltic Sea, between central Sweden and the Baltic countries. On the other hand, the presence of lydite (radiolarian chert) demonstrates that some material was also derived from the south, primarily the Bohemian Massif (Hucke, 1928; Ahrens et al., 1968; Duphorn et al., 1973; Ahrens and Lotsch, 1976). In the southern parts of Thuringia, in the area of the Thüringer Wald, there was limited late Neogene sedimentation in subsident areas (Figure 18.1).

Northern lowlands and northeastern region

In the northern lowlands, Lower Pliocene terrestrial deposits appear to have been extensive, but only in southwestern Mecklenburg have erosion relicts been preserved. As far as we know, deposits of Plio–Pleistocene transitional series can be found only in the southwestern parts of this area (Ahrens et al., 1968).

The best-known locality of this type is Rüterberg, where the lower horizons of the sequence (Bergton-Diatomit-Silbersand complex) are followed discordantly by the Loosener Kiese (Gehl, 1958; von Bülow, 1964), a unit characterized by a mixture of "southern" pebbles (Kieselschiefer, etc.) and silicified rocks and intensely weathered granite from the north. The Bergton-Diatomit-Silbersand complex was formerly correlated by Krutzsch (1959) to the Reuverian or a still earlier stage on the basis of pollen, but more recently this unit has been correlated with the Upper Miocene (Lotsch, 1981; Krutzsch, 1988). The Loosener Kiese deposits have been correlated with the kaolin sands of the Isle of Sylt by Gehl (1958), who correlated both units to the Upper Pliocene, whereas von Bülow (1964) suggested an early Pleistocene age (cf. von der Brelie et al., Chapter 17, this volume). The deposits of the Loosener Kiese provide a fossil flora (H. Mai, in Ahrens, et al., 1968) showing some affinities with the Nordhausen association. It was recently referred to a time interval characterized as "post-Reuverian" but not typical Tiglian (TC5) by Krutzsch (1988).

Pliocene gravels and climate events

In eastern Mecklenburg and in northeastern Brandenburg numerous deposits of quartz sands, with pebbles of silicified rocks of Baltic origin, were originally referred to the Pliocene (Hucke, 1928) or even the Pleistocene (Stolley, 1900) because the pebble association was similar to the kaolin sands of the Isle of Sylt. In contrast, Berger (1941), Quitzow (1953), and Ahrens and Lotsch (1976) suggested a Miocene age for these deposits because of the mode of deposition.

Recently, deposits of quartz sand showing the typical "Pliocene" pebble association have been found, unambiguously dated to the Miocene, in large lignite pits much farther south in the area of Cottbus, Finsterwalde, and Lübben. From the lower part of the quartz-sand series (0.5 m above the second Lausitz lignite seam) in the lignite open pit at Seese, there is a diverse fossil flora which belongs to the Miocene floral zone VIII (Mai, 1967). The quartz-sand series at Petersdorf and Finkenheerd is also middle Miocene age (Ahrens and Lotsch, 1976). The use of the Baltic pebble association as a criterion of Pliocene age clearly is not justified (Krause, 1933; Berger, 1941; Quitzow, 1953).

Quartz-lydite gravels, reflecting a southern origin, are quite widely distributed in the Lausitz area, and to a lesser extent in northeastern Brandenburg. Mielecke (1929, 1934) interpreted them as a southern Pliocene facies correlated with the northern, supposedly Pliocene quartz-sand facies described earlier. To account for those deposits, Genieser (1955) suggested a Pliocene "Senftenberger Elbelauf" (a channel of the Elbe River containing the Senftenberg Elbe gravels) and a succeeding pre-glacial "Bautzener Elbelauf."

According to Ahrens et al. (1968) and Ahrens and Lotsch (1976), characteristic Senftenberger Elbelauf deposits with quartz-lydite gravels can be dated not only to the Upper Pliocene or Lower Pleistocene but also as far down as the first Lausitz

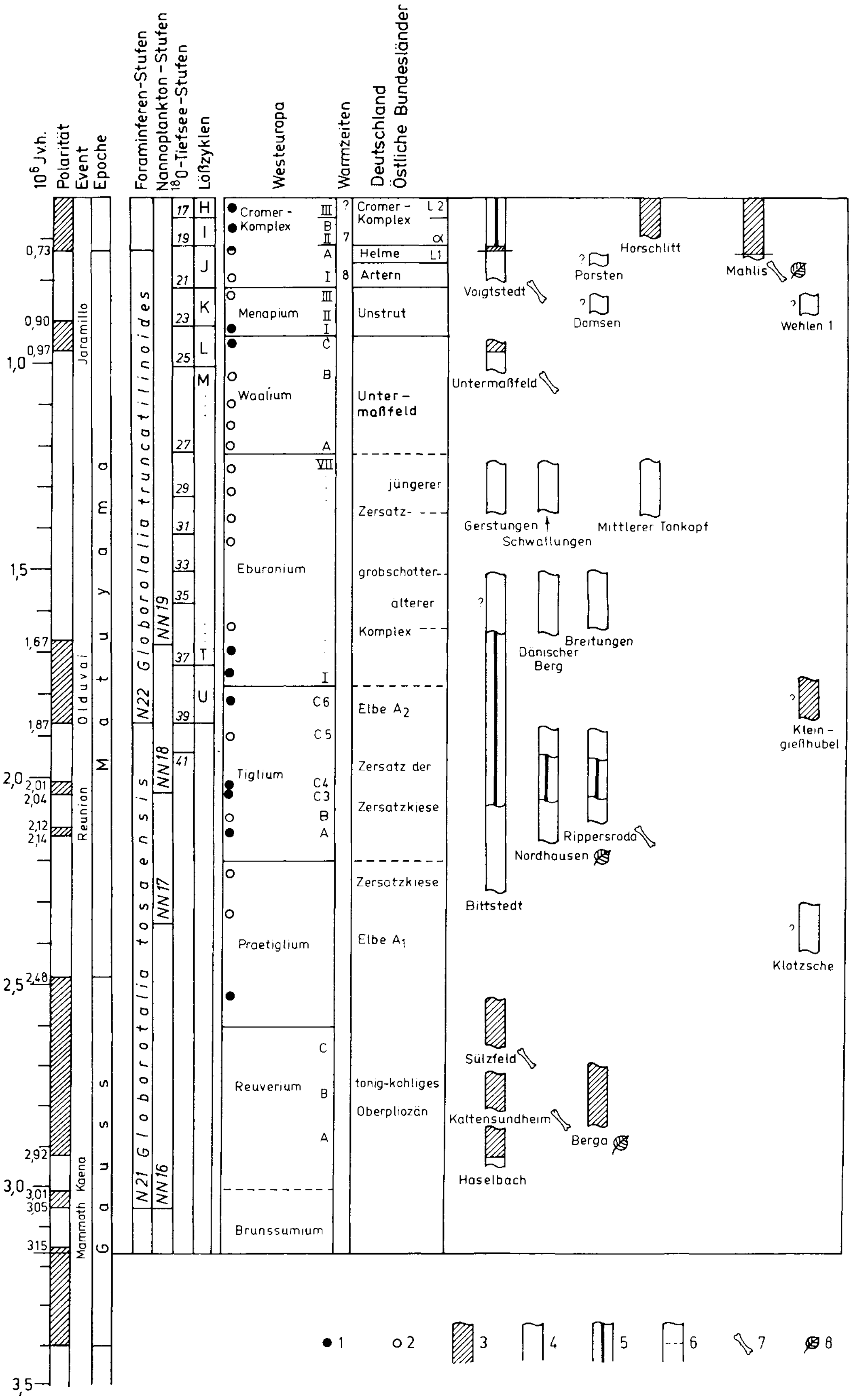

Figure 18.1. Pliocene and Pleistocene localities in eastern Germany. Paleomagnetic interpretation by F. Wiegank (1982): 1, normal polarity; 2, reverse polarity; 3, position of the section with normal polarity; 4, position of the section with reverse polarity; 5, hiatus; 6, paleomagnetic anomalies; 7, fossil fauna; 8, fossil flora. The Zersatz-Grobschotter-Komplex comprises highly weathered coarse gravels and boulders deposited in cold-climate conditions, the Zersatzkiese consists of highly weathered gravels, and the Tonig-kohliges Oberpliozän is composed of clay and lignite deposits of the Upper Pliocene. The terms “älterer” and “jüngerer” refer to older and younger, respectively.

lignite layer in the Rauno sequence of that area. As noted earlier, the Lausitz lignites are of Miocene age, and in fact the lower horizons of the quartz-lydite gravels contain microflora in clay and lignite beds at Klettwitz, Rauno, and Welzow that correspond to floral zone XIII of the Upper Miocene (Mai, 1967; Ahrens and Lotsch, 1976). Nearly 98% of the Senftenberger Elbe gravels are composed of quartz and other weathering-resistant components, and the remaining feldspathic cobbles are heavily decomposed. The stable heavy minerals also predominate over the unstable ones (Genieser and Diener, 1958; Wolf, 1980).

In the area of Ottendorf-Okrilla, where such gravels are locally up to 40 m thick, a peculiar characteristic is the occurrence of oversized "drift boulders," which are unknown in early and middle Tertiary deposits of the southern edge of the northern lowlands. The drift boulders, moreover, increase in number from bottom to top. Numerous possible "cryogene structures" have been reported, such as the wedge-like structures interpreted by Genieser (1955) and Präger (1976) as fossil ice wedges. This is difficult to reconcile with the warm–temperate *Taxodium–Liquidambar–Ulmus* macrofloral association from silty layers in the same gravel body. *Parrotia fagifolia* and *P. pristina* are also common (Jähnichen, 1968; Walther, in Kube, 1979).

According to the fossil flora and the terrace stratigraphy, an early Pliocene age for the Senftenberger Elbe gravels is the most probable, although a middle Miocene to late Miocene age cannot be excluded (Ahrens et al., 1968). The cryo-structures may reflect temporary climatic cooling. Wolf (1980) recently distinguished two members within the Senftenberger Elbe gravels: a "240-m" member and an "A1" member. According to paleomagnetic investigations (Wiegank, 1982), the A1 gravel body has primary reversed polarity, in agreement with a correlation (Wolf, 1980) to the Pretiglian.

Lower Pleistocene gravels

The next in the sequence of Elbe gravels, the deposits of the Bautzener and Schildauer Elbe channels, follow valley stream systems that are clearly seen in the hilly areas. They are very readily distinguishable from the Senftenberger gravel type in having only 50–80% quartz and often 20–30% poorly resistant feldspathic material. Similarly, the unstable heavy-mineral component predominates over the resistant.

Cryoturbation structures and ice-wedge pseudomorphs produced by intense frost activity are recognized in these gravels from several different localities (Schubert, 1980). The gravels of the Bautzener and Schildauer Elbe channels (A2 and E valley levels) are deposits of a glacially influenced fluviatile cycle (Wolf, 1980). Fine-grained sediments of the A2 sequence at Kleingiesshübel are normally polarized, and Wiegank (1982) postulated that sedimentation occurred during either the Réunion subchron or the Olduvai subchron of the lower Matuyama.

In the area west of the Elbe River, four individual gravel terraces are older than the Elsterian glaciation. Those terraces are clearly distinguishable, because of their heavy minerals and gravel composition, from the quartz gravels mentioned earlier. In that group, with the exception of the youngest terrace, which has been correlated with the early Elsterian, the older gravel bodies formerly were interpreted as either pre-glacial or Pliocene in age. In all those gravel bodies, permafrost patterns, including ice wedges, are developed (Eissmann, 1975, 1981). Today we correlate those three earlier accumulations with the Brüggen, main Eburonian, and Menapian cold-climate phases. The Brüggen has been identified with the Pretiglian by some researchers, and by others with the early Eburonian. The post-Brüggen limnic sequence of Zeuchfeld-Borntal, however, with *Fagotia acicularis, Valvata naticina, Lithoglyphus pyramidatus*, and *Discus perspectivus* (Mania, 1973), is correlated with the late Pliocene Tiglian climate stage of the lower Rhine Basin.

Thuringian basin

In the southwestern parts of eastern Germany, between the Harz and the Thüringer Wald mountains, very localized late Neogene sedimentation is known. In contrast to the northern lowlands, all recorded deposits of the southern area are of late Pliocene–early Pleistocene age. Despite the considerable distances between the sites, these uppermost Pliocene and lower Pleistocene sequences show a general correspondence of characters because they were accumulated under similar conditions, albeit in isolated closed depressions.

The lignite layers of Gerstungen (Mai and Walther, 1988), Kranichfeld (Mai, 1965; Mai and Walther, 1988), Oberzella, and Berga (Krutzsch and Majewski, 1965; Mai and Walther, 1988) are characterized by fossil floras correlated with the Dutch Reuverian or an even older stage (Brunssumian?). The key taxa here are mainly exotic elements like *Nyssa, Sciadopitys, Platycaria, Liquidambar, Sequoia, Taxodium*, and *Symplocos*. The basin clays of Berga, containing a "Reuver flora," are normally polarized and have been correlated by Wiegank (1982) with the upper Gauss chron. The microflora of Gerstungen, with *Tsuga, Ilex, Carya, Pterocarya, Sciadopitys*, and *Nyssa* and other palynomorphs, is characterized by a still greater number of exotic elements, again suggesting correlation with the Dutch Reuverian or, according to Mai and Walther (1988), possibly with the Brunssumian, because of a greater number of ancient ("Miocene") elements in that macroflora. A similar but much richer microflora is known from the borehole of Görsbach, near Kelbra, showing high facies differences in comparison with the nearby "Reuver" (or older?) flora of Berga (Krutzsch and Majewski, 1965). In the lignite horizons of Kranichfeld (Mai, 1965; Mai and Walther, 1988) a fossil macroflora has been described, with about 47% of its taxa being exotic. The last appearances of some of these taxa are in deposits of Reuverian age.

The Mastodon-Schotter (gravels with mastodont remains) of Sülzfeld, with *Mammut borsoni, Anancus arvernensis, Tapirus arvernensis*, and *Dicerorhinus* sp., are normally polarized and have been correlated by Wiegank (1982) with the upper Gauss chron. Based on pollen analysis of samples from the *Mammut borsoni* horizon, from the doline filing of Kaltensundheim near Meiningen, Ukrainzeva (in Kahlke and Ukrainzeva, 1986) suggested a correlation of that horizon with the Dutch Pretiglian, but Krutzsch (1988) postulated a much earlier correlation (lower

Villafranchian climatic minimum). According to Mai and Walther (1988), the fossil macroflora of Kaltensundheim suggests the "first real intra-late-Pliocene phase of cooling and impoverishment of the flora, because this flora with 65 species is not poor in species, but poor in exotic elements." The mammalian assemblage from the deposit suggests a later Pliocene (early Villafranchian) age, on the basis of *Hypolagus* sp., *Mammut borsoni*, and a primitive cervid (Böhme, 1963). The horizon has normal remanent paleomagnetism (Wiegank, 1982) and has been correlated with the upper Gauss chron.

In the later group, the fossil macroflora of Nordhausen (Mai, 1965; Mai and Walther, 1988) comprises, in additon to 39 species found in the recent European flora, stratigraphically important exotic taxa such as *Salvinia cerebrata, Carpolithus (Epipremnum) ornatus*, and *Decodon bashkiricus*, as well as *Spirematospermum wetzleri*, previously found only in Pliocene deposits. In the pollen spectra of Nordhausen (Erd and Majewski, in Ahrens et al., 1968), *Pinus, Alnus, Tsuga, Picea*, Graminaceae, and *Osmunda* are found regularly. Other taxa, like *Abies, Quercus*, and *Tilia*, rarely exceed 1%. Apart from the relatively high *Tsuga* pollen levels, the other exotic taxa recorded are *Sciadopitys* and *Eucommia*. According to pollen analyses, Erd (in Ahrens et al., 1968) correlated the fossiliferous deposits of Nordhausen with the Tiglian A, whereas Mai and Walther (1988) suggested an "uppermost Pliocene" level according to the macrofloral record. The fossiliferous layers show predominantly reversed remanent polarity and have been correlated with the lower Matuyama (pre-Olduvai) interval by Wiegank (1982).

North of the Thüringer Wald, mainly in the area of Ohrdruf on the terrain known as Ohrdrufer Muschelkalkplatte, a formation of highly weathered gravels is known as the Zersatz-Kies complex (Ziegenhardt, 1965). The most important locality for those highly weathered gravel deposits north of the Thüringer Wald is the Rippersroda II horizon (Mai, Majewski, and Unger, 1963; Schramm, 1964; Unger, 1964, 1974). In general, the Zersatz-Kies complex discordantly overlies the Triassic limestone (Muschelkalk) of this area. Southeast of Rippersroda, however, the heavily weathered gravels discordantly cover the limnic filling of a doline, which includes the Rippersroda I lignite. The fossil flora from the Rippersroda I lignite layers, with 33% exotic species such as *Tsuga, Carya*, and *Sciadopitys* (Mai et al., 1963; Mai and Walther, 1988), is associated with the Rippersroda II fauna of fossil mammals from the overlying Zersatz-Kies complex (Dietrich, 1953; Kahlke, 1968), which includes the following species: *Mimomys* sp., *Trogontherium minor, Anancus arvernensis, Dicerorhinus etruscus, Eucladoceros ernestii*, Cervidae gen. et sp. indet. (small cervid), and *Leptobos* sp.

Following intensive erosion and deep weathering of the Zersatz-Kies complex in the area north of the Thüringer Wald, the Zersatz-Grobschotter complex (highly weathered coarse gravels and boulders) was deposited (Ziegenhardt, 1965), with evidence of having been accumulated during cold conditions. Similar coarse gravels of the Werra River area were also accumulated during cold conditions, as proved by Siegel (1959). The local extreme thicknesses of the layers, as well as the mode of deposition, have been explained as the result of syngenetic and postgenetic subrosion of the deeper horizons (Siegel, 1959), and Ellenberg (1969) suggested that these Werra gravels were the equivalents of the Zersatz-Grobschotter complex north of the Thüringer Wald. The Basiskiese ("basal gravels") of Voigtstedt, near Sangerhausen, are thought to have accumulated at about the same time (Cepek, 1968).

The weathered coarse gravels of the Zersatz-Grobschotter complex north of the Thüringer Wald and their equivalents of the Werra River system (Dänischer Berg, Breitungen, Gerstungen, Schwallungen, Mittlerer Tonkopf) have been correlated according to lithostratigraphic and magnetostratigraphic criteria with the Matuyama epoch between the Olduvai and the Jaramillo subchrons by Wiegank (1982).

Pliocene–Pleistocene boundary horizons of eastern Germany

The proposal made by the INQUA Subcommission 1-d and the IGCP-41 working group at the 1982 INQUA meeting in Moscow, to define the Neogene–Quaternary boundary-stratotype in the Vrica section, has been approved by international bodies (Foreword, this volume). The boundary is placed at the base of the shale layer overlying sapropelic layer *e*, within the uppermost part of the Olduvai event, according to present information (Pasini and Colalongo, Chapter 2, this volume). The few deposits correlative to this level in eastern Germany indicate that the boundary should be drawn between (a) the level of Nordhausen and Rippersroda II (correlated to the late Pliocene according to fossil macroflora and fossil mammals, respectively) and (b) the Zersatz-Grobschotter complex of the Thüringer Wald and its equivalents in the Werra River system (lowermost Pleistocene). In the area of the Elbe River the A2 valley fill ("A2-Talboden") is correlated with the uppermost Pliocene, and the E valley fill ("E-Talboden") corresponds to the lowermost Pleistocene.

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19 The Plio–Pleistocene of Hungary

ANDREW A. RÓNAI

Introduction

Recent studies and investigations have agreed that paleomagnetic polarity changes are the most expedient criteria for establishing the main divisions of Quaternary stratigraphy and for tracing the "N/Q" limit between Neogene and Quaternary (actually, the Pliocene–Pleistocene boundary). In Hungary, there are two viewpoints as to where to mark the N/Q boundary: at the Gauss–Matuyama paleomagnetic boundary (Rónai, 1984), dated to 2.5 Ma, or at the top of the Olduvai event (Pasini and Colalongo, Chapter 2, this volume), dated to 1.8 Ma. Each of these reversals is close to the initiation of a period of global cooling.

If we accept the paleomagnetic polarity changes as signposts of stratigraphic boundaries, it is still necessary to reach consensus on which of the cooling events should be recognized as the lower boundary of the Quaternary. Both reversals have been recognized in Hungary, but most Hungarian scientists have long favored the view that the beginning of the Quaternary should be determined by a major stratigraphic and paleontological change, which we now know to have been almost coincident with the Gauss–Matuyama paleomagnetic boundary – see Rónai (1984) and the references cited therein.

There are several reasons to prefer this date, but the main one is that it corresponds to the most impressive change in the Carpathian Basin during the past 5–10 million years, namely, the epirogenetic uplift of that great territory in the middle of the European continent, and the regression of the Pannonian Lake, comparable to regression in the Dacic Basin (Ghenea, Chapter 20, this volume). This event changed the geomorphology of the entire region and seems also to have been coincident with the starting point of a new tectonic cycle.

Geological investigations in the Great Hungarian Plain

Because of growing interest in problems of agrogeology, engineering geology, hydrogeology, and environmental protection, a very detailed and thorough program of mapping the younger deposits of the Pannonian Basin was started by the Quaternary Department of the Hungarian Geological Institute in 1964 (Rónai, 1972; Franyó, 1977). The research was amplified to include Quaternary tectonic movements, with the ultimate goal of demonstrating the late Neogene and Quaternary geological history of the Great Hungarian Plain.

The project was organized around the mapping of three depth horizons (Franyó, 1977). The first horizon consisted of the superficial and near-surface sediments, as revealed by a network of shallow boreholes, up to 10 m in depth, drilled on a regular 1.5-km grid that covered most of the plain. Field data were mapped at a scale of 1:50,000 and published in atlases at a scale of 1:200,000, in which agrogeologic, hydrogeologic, and engineering geologic data and test results were also plotted. As many as 18–20 geologic map sheets were prepared for each mapped quadrangle.

The second depth horizon was studied in medium-depth (100–1,500 m) boreholes in two linear arrays at right angles. The holes were drilled to investigate the thickness and stratigraphy of the Quaternary down to the Neogene–Quaternary boundary. In that part of the project the Pliocene and Pleistocene aquifers were identified and tested, and the core samples were analyzed for paleotectonic information. The boreholes were then equipped with artesian piezometers to monitor selected aquifers; three to four wells were completed at each location in order to make simultaneous observations of ground water at different depths.

Finally, the third depth horizon of the investigation consisted of pre-Pliocene deposits involved in the floor of the modern basin. At its inception in 1964, the project was scheduled to take 21 years; after systematic progress, the fieldwork and laboratory analyses were completed by 1984, and the final atlases were in press by 1985.

History of the Pannonian Basin

The geological evolution of the modern Pannonian Basin (Figure 19.1) began in the middle Miocene, with contemporaneous uplift of the Carpathian mountain arc and subsidence of the enclosed lowlands. The newly defined basin was occupied by an arm of the Paratethys continental sea, connected to the proto-Mediterranean Tethys Ocean. In that constantly subsiding regime, some 2,000–3,000 m of shallow-marine shelf sediments

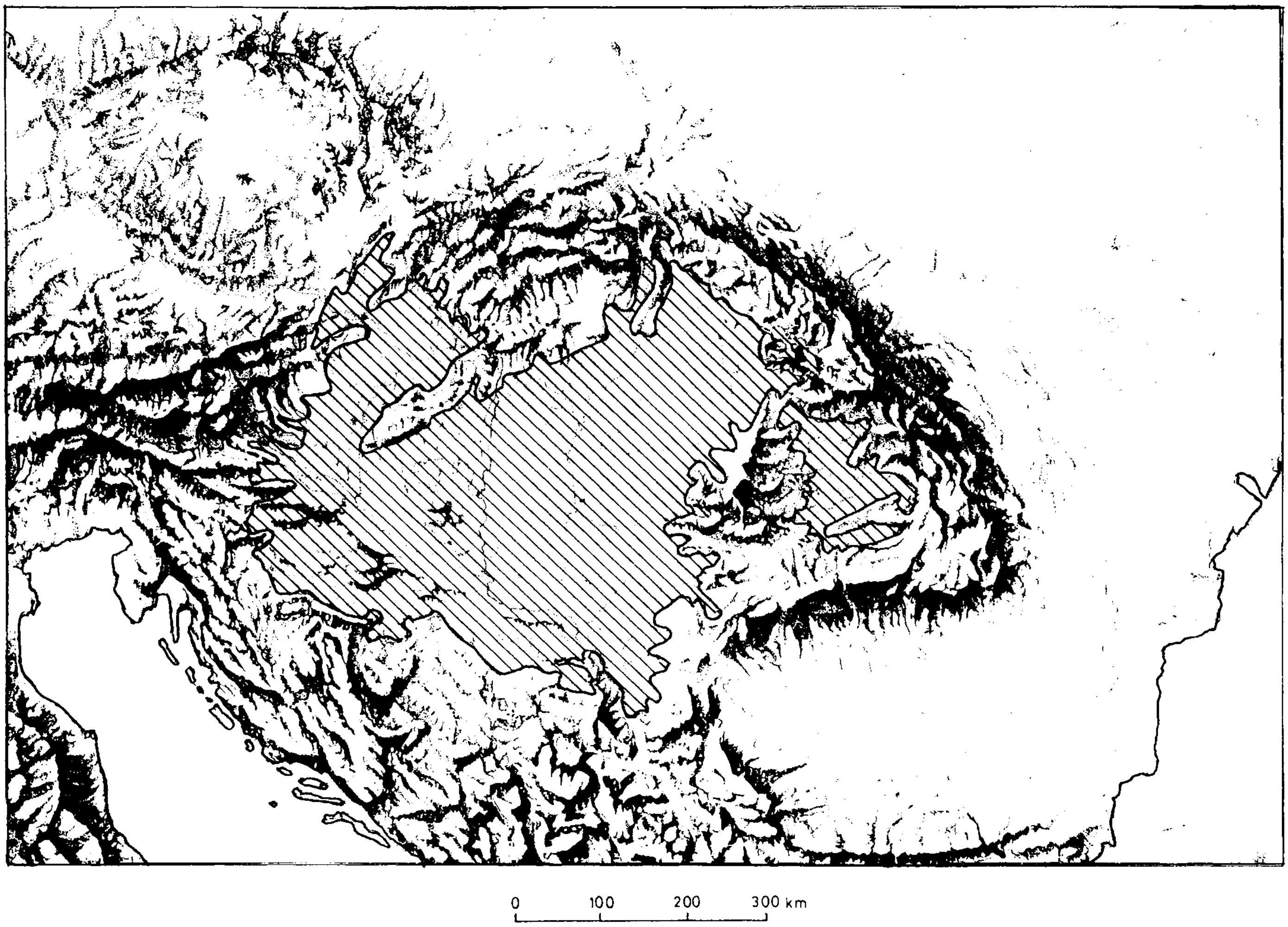

Figure 19.1. The Hungarian Great Basin, with the area of the Pannonian lake in Pliocene time indicated by the hatched area.

of the Lower Pannonian Stage were deposited, consisting of mixed conglomerates, sandstones, marls, and clays.

Near the end of the Lower Pannonian time, the connection to the Tethys became more restricted, and all of the Paratethys, from central Asia to southern Germany, became brackish-marine. In the Pannonian Basin, isolation became complete, and the Lower Pannonian sea was transformed to the Upper Pannonian fresh-water lake. During that phase, some 1,000–2,000 m of lacustrine sediments were deposited in the basin. In the lower levels, those strata consist of rapidly alternating sand, fine sand, silt, and clay; the higher levels are dominantly variegated clays, representing the Levantine substage (Nikiforova, Chapter 21, this volume).

In economic terms, the lower part of the Upper Pannonian sequence contains the known oil and gas reserves of the Great Plain, as well as large volumes of saline, locally thermal water. In the upper part of that same sequence, brown coal and lignite deposits occur in the marginal facies near the edges of the plain.

At about 2.5 Ma, subsidence slowed and was overtaken by deposition as the lake began to fill in from the margins, forming a broad land surface traversed by a stream network flowing from the surrounding highlands toward a central sink (Figure 19.1). In different parts of that central area, some 600–700 m of fluviatile sediments were deposited. Before that time, the Pannonian Basin had subsided at approximately the same rates over very large areas, as shown by the fact that the limnic layers are horizontal and that their stratigraphy correlates in detail throughout major portions of the basin. The beginning of fluvial sedimentation, on the other hand, was marked by a change in tectonism and the formation of deep local depressions of smaller extent. From the beginning those depressions were filled with fluvial sediments, consisting of coarse sands and gravels in some and fine-grained silts and clays in others. The lake bottom was tectonically dismembered, and local areas subsided at varying rates (Figure 19.2).

In detailed paleontological studies devoted to identification of the Pliocene–Pleistocene boundary, a persistent problem was the biotic sterility of thick sedimentary sequences in the upper levels

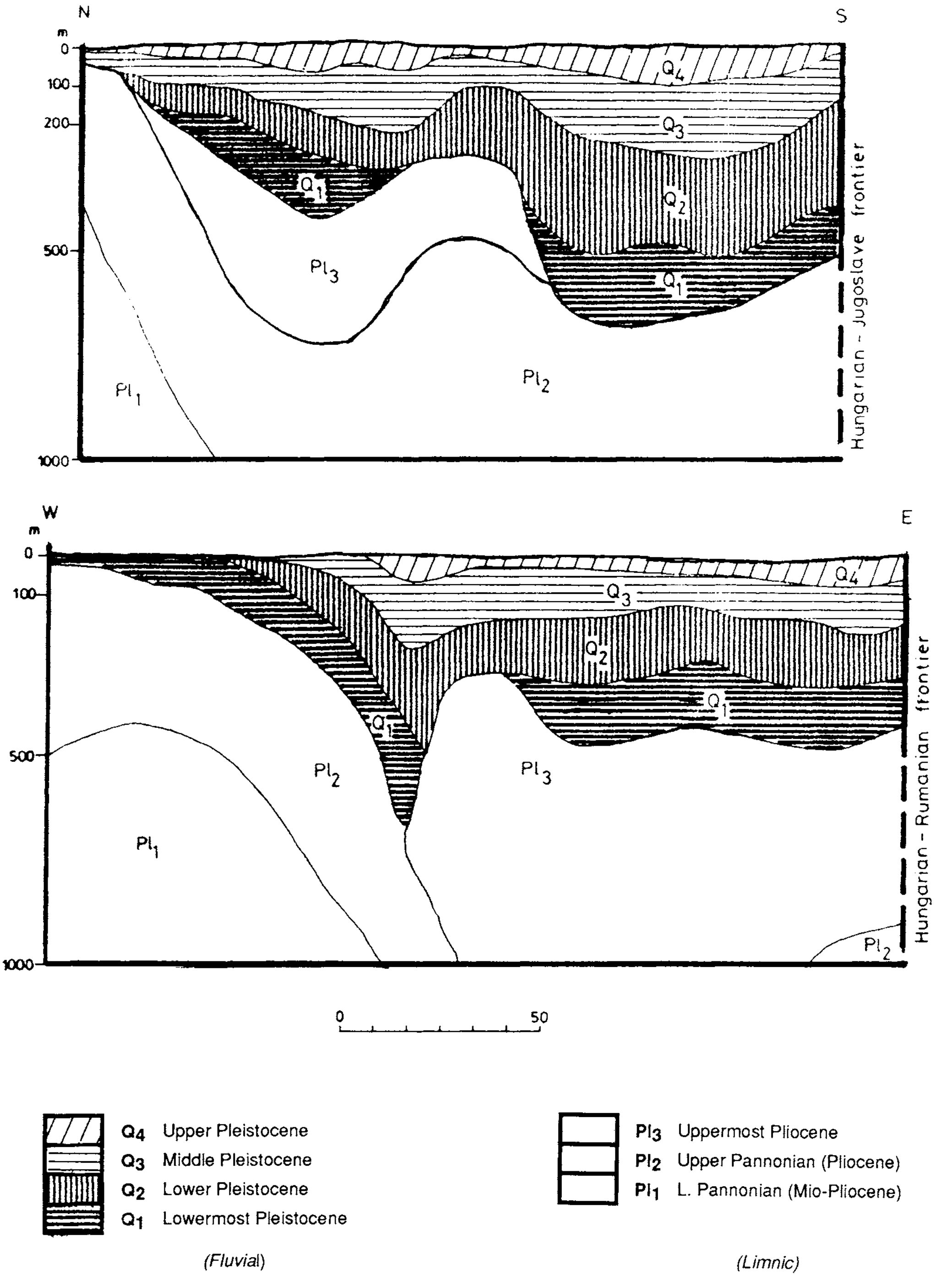

Figure 19.2. Quaternary sedimentation in the Hungarian Great Basin. Limnic sediments of the Lower, Middle and Upper Pliocene (Pl_1–Pl_3) and fluvial sediments of the lowest, lower, middle, and upper parts of the Pleistocene (Q_1-Q_4) are shown in N–S and E–W sections. According to paleomagnetic analysis, the "lowest Pleistocene" (Q_1) is older than the Vrica boundary.

of the Upper Pannonian lake deposits, which prevented any indication of climate changes. On the other hand, paleomagnetic analyses of the core samples from two boreholes in the Körös Basin represented a significant scientific achievement – see Cooke (1981) and the Appendix to this chapter. The Körös Basin is one of the deep local depressions where sedimentation was essentially continuous throughout the Quaternary, so that a high-resolution picture of paleomagnetic reversals can be traced back to 5 Ma. In those cores, the change from limnic sedimentation to fluviatile was almost precisely synchronous with the Gauss–Matuyama reversal, falling at 2.4 Ma.

The tectonically determined shoaling of the Pannonian lake appears to have been a period of distinct climate change as well. The disappearance of the Pannonian lake was aided by an extremely arid climate, as indicated by the sterile layers at the top of the lacustrine sequence, in which no trace of vegetation or animal life can be found, as contrasted with the rich mollusk and ostracode faunas and the abundant pollen in the strata below and above the sterile layers.

The paleontological evidence shows that at the beginning of the fluviatile deposition, the climate became more humid, although remaining quite warm; precipitation rates rose, and vegetation again covered the terrain, becoming increasingly opulent up to the level of the Olduvai subchronozone. At that time, the climate began to alternate between cold and warm, with strong fluctuations in precipitation, but none of those changes had an effect on vegetation comparable to the transition in the earliest Gauss (Rónai, 1970) (Figure 19.3). Paleontologically, it has not been possible to differentiate any major biotic changes that would allow identification of internal, sub-epoch boundaries of the Pleistocene, such as the four major glacial intervals observed and demonstrated in the Alps. Only two major biostratigraphic units can be distinguished in the post-Gauss interval, with the boundary marked by the transition to more variable climates at the top of the Olduvai subchronozone.

The great change in geomorphology, tectonism, and climate at the Gauss–Matuyama boundary supports the Hungarian opinion that the disappearance of the Pannonian lake should mark the end of the Pliocene and that the beginning of the fluvial accumulation should mark the start of the Quaternary, coincident with the Gauss–Matuyama reversal. In this schema, the climatic and faunal changes at the top of the Olduvai subchronozone divide the older Pleistocene from the younger, colder part, similar to a long-held opinion about the divisions of the Quaternary in the former Soviet Union (Nikiforova, Chapter 21, this volume).

Sedimentation cycles and climate change. The beginning of the fluvial sedimentation at the Gauss–Matuyama boundary can be detected in most boreholes on the Pannonian plain by a shift to cyclical variations in granulometry (Rónai, 1984). Whereas the lake sediments below that boundary are characterized by rapid and sharply bounded oscillations between sandy and clayey layers, in the fluvial sequence one can detect *sediment cycles* in which the granulometry changes gradually from coarse (gravel or sand) to fine (silt or clay) and again gradually back to coarse. Although cyclic sedimentation may have resulted from irregular, step-wise shifts in the rates of subsidence in the basin, other evidence from the magnetostratigraphic calibration (discussed later) suggests that the granulometric variations reflect alternations in precipitation and climate during the Quaternary that influenced the quantity and coarseness of the river deposits (Figure 19.4).

Palynological studies of the interval from the present back to the Gauss–Matuyama boundary in the Great Hungarian Plain have provided remarkably detailed data showing 17 complete cold/warm climatic cycles, or 35 fluctuations in total (Rónai, 1970, 1984). The succession of climates in the Pannonian Basin demonstrates a steady shift from southern to northern characteristics (Rónai, 1972). In the lower Matuyama interval, the climate of the Pannonian Basin was generally like the present climate along the North African coastline, and in the upper Matuyama it became more like that of Italy or Greece. For the Brunhes, we find a cool, continental climate, which changed finally to a climate like that of Scandinavia nowadays (Rónai, 1970; Kretzoi and Pécsi, 1979). Overall, erosion was generally slow during the warmer climates and faster during the cooler, so that the lower part of the sequence has a higher proportion of silty and clayey layers in the core profiles, and the sediments deposited during the later period are more sandy and gravelly.

In the 800-k.y. interval of relatively warm and quite humid climate in the first part of the Matuyama, we find seven major climate cycles, according to variations in the indicators of temperature and humidity. From the Olduvai to the beginning of the Brunhes, six additional climate cycles can be distinguished according to pollen statistics, followed by four cycles in the Brunhes. As for the latter 10 cycles, the average climate was cool to cold; four cycles were relatively humid, and six were arid.

The Körös Basin boreholes. Two of the boreholes in the Körös Basin, Dévaványa (1,116 m) and Vésztö (1,200 m), are noteworthy for their paleomagnetic data (Cooke, Hall, and Rónai, 1980; Appendix, this chapter). The Körös Basin is the second deepest Plio–Pleistocene depression in the Great Plain, with Paleozoic basement at depths between 3,000 and 4,000 m overlain by the full Lower Pannonian, Upper Pannonian, and Quaternary sequence. The Quaternary (i.e., fluviatile) sequence, 200–500 m thick, consists mostly of silts and clays, with occasional intercalations of thin sand beds. The paleomagnetic stratigraphy in the Dévaványa and Vésztö boreholes indicates that the Körös sedimentary series represents the whole fluviatile Quaternary, without any appreciable hiatus (Cooke et al., 1980). In these boreholes, the measurements indicated all of the known polarity reversals of the international paleomagnetic scale, and the distances between the paleomagnetic polarity changes measured on the cores were closely proportionate to the dated time intervals between the polarity reversals.

Core samples were taken at 1-m intervals, with 20-cm-offset backups in some sections. Samples were sent to the paleomagnetic laboratory at Dalhousie University, Halifax (Cooke et al., 1980). The results of the analyses, correlated to the analyses of sedimentology, paleontology, and palynology in the cores, were initially presented at the INQUA Congress in Birmingham in 1977, at the Paleomagnetic Symposium in Budapest-Szeged in 1979 (Cooke et al., 1980), and at the International Field Conference of the International Geological Correlation Program (Projects 41 and 128) in Tucson in 1981.

At first, these data were used to designate major subdivisions of the Quaternary, based on the historic concept that the Quaternary was equivalent to the full sequence of fluviatile deposition in the Pannonian Basin (Rónai, 1968; Kretzoi and Pécsi, 1979). The beginning of that phase, and thus the beginning of the Pleistocene, was dated at about 2.4 Ma in the old calibration (Preface, this volume) at the Gauss–Matuyama boundary. The Olduvai normal-polarity event was used as an

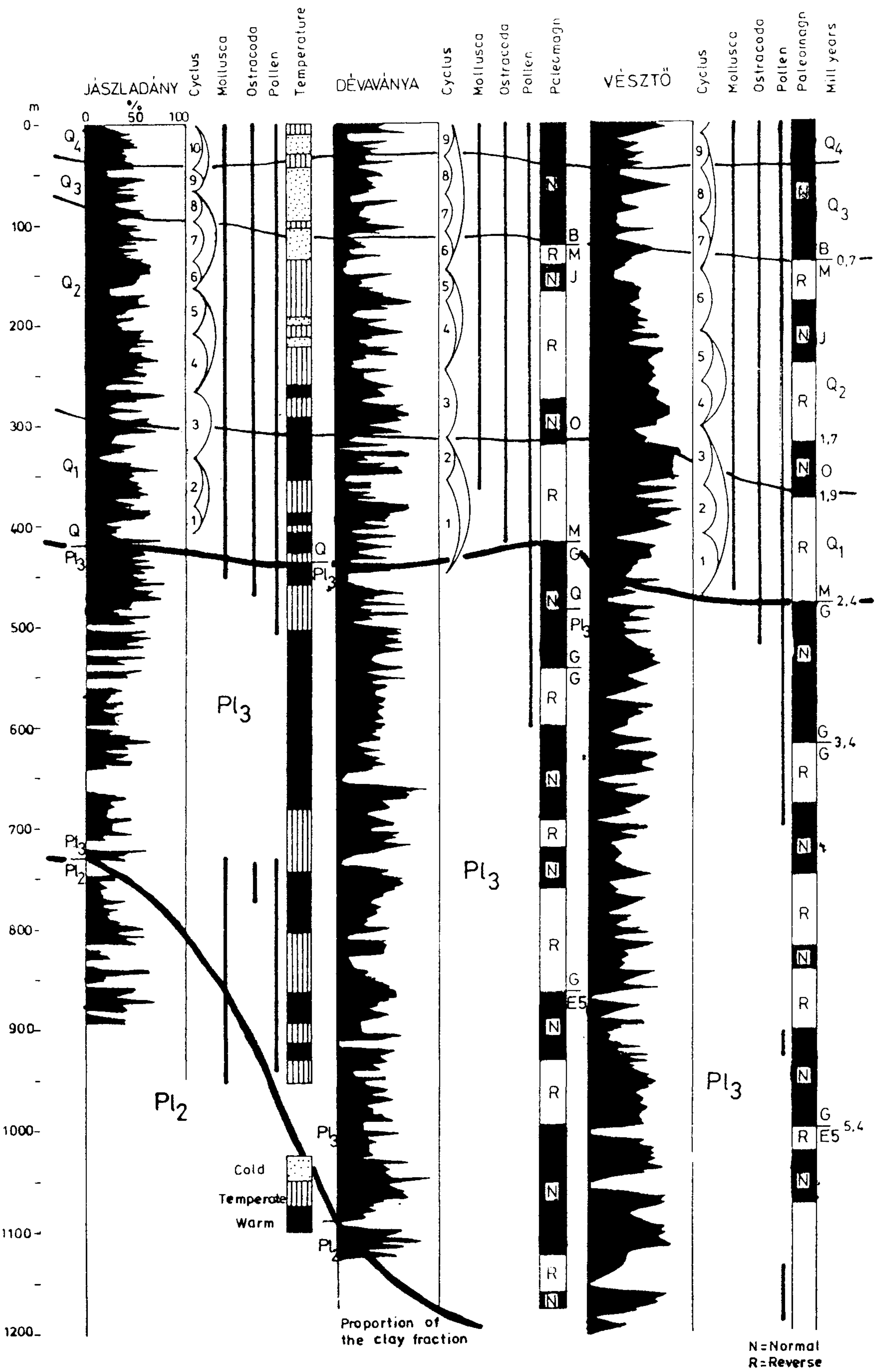

Figure 19.3. Plio–Pleistocene stratigraphy in three key boreholes in the Hungarian Great Basin. The clay fraction is shown by the black profile, increasing to the right. Magnetostratigraphy in the Dévaványa and Vésztö boreholes is compared to paleoclimatic data from the Jászladány borehole.

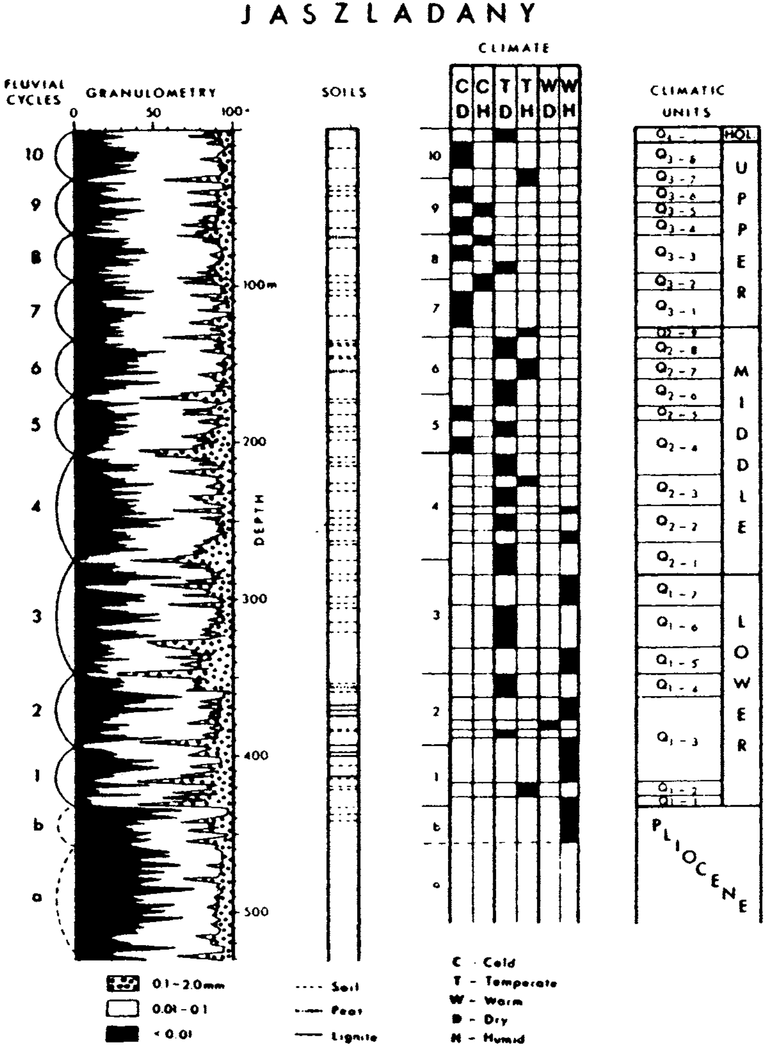

Figure 19.4. Summary of data in the upper part of the Jászladány borehole. The temperate-dry interval of cycle 3 is approximately equivalent to the Olduvai subchron. (From Cooke, 1981, fig. 2, with permission.)

index for dividing between the "Oldest" and the "Old" (or Lower) Pleistocene, and the Matuyama–Brunhes paleomagnetic reversal was used to divide the Lower from the Middle Pleistocene (Rónai, 1984). Between the Middle and Upper Pleistocene there are no paleomagnetic events to serve as boundary markers. As noted earlier, this schema is inconsistent with the currently adopted criterion of the Vrica bed *e*, which locates the base of the Pleistocene to be equivalent to the cold-climate maximum at the end of the Olduvai event. As noted earlier, this level has been the boundary between Oldest Pleistocene and Old Pleistocene in the conventional Hungarian usage.

Sedimentation rates determined from the paleomagnetic data indicate that subsidence was almost steady at 0.2 mm/year during the fluviatile interval in the Körös Basin, despite the changes in sediment types. That evidence confirmed that the variations in granulometry probably were almost completely independent of tectonism and therefore should be considered as closely reflecting the changes in environmental conditions.

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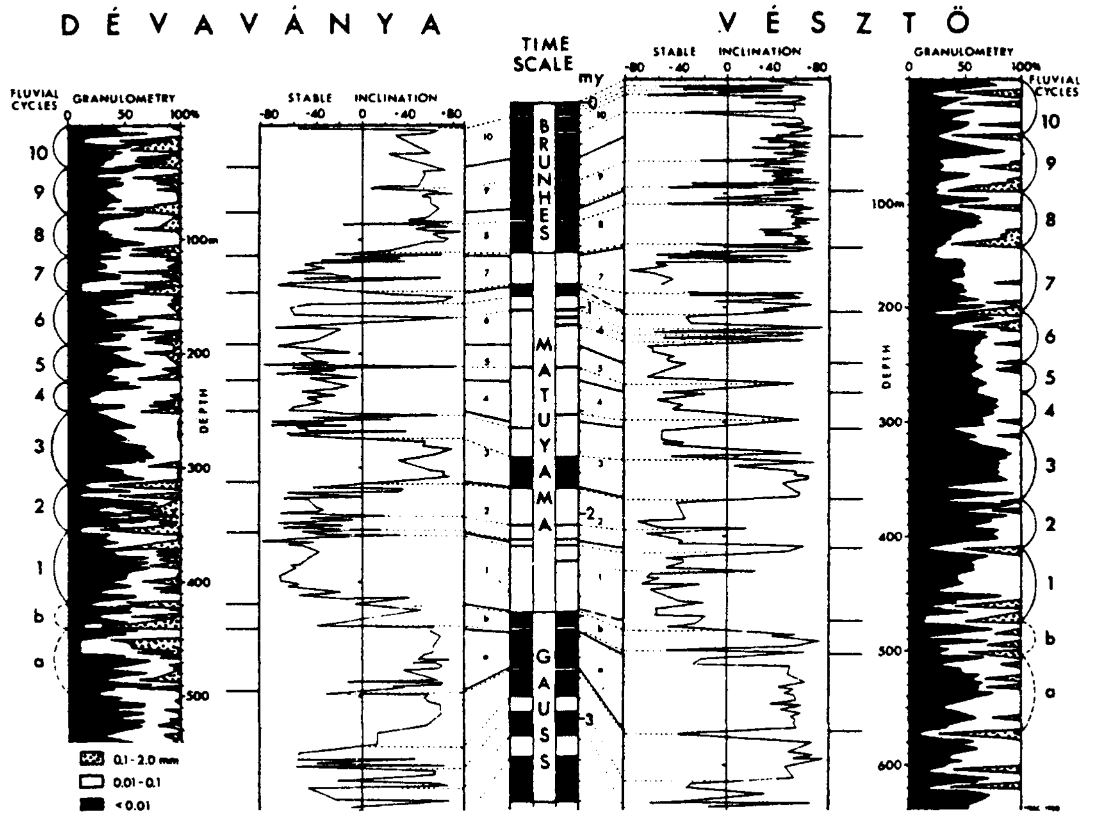

Figure 19.5. Correlation of fluvial cycles and magnetostratigraphy in the Dévaványa and Vésztö boreholes. (From Cooke, 1981, fig. 3, with permission.)

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Appendix: Quaternary record from deep boreholes in the Great Hungarian Plain*

H. BASIL S. COOKE

The Carpathian Basin has been subsiding continuously since the mid-Tertiary, so that the Great Hungarian Plain is underlain by several thousand meters of Quaternary, Pliocene, and Miocene sediments. The Pliocene beds were laid down in a great "Pannonian" lake and are overlain by 200–600 m of fluviatile sediments, long regarded by Hungarian geologists as representing the Quaternary. Ten sedimentological cycles have been recognized in these "Quaternary" deposits, and a borehole at Jászladány in 1964/65 yielded a good pollen record showing evidence for at least 35 climatic fluctuations.

The earlier climate was warm, changing to temperate in the middle "Quaternary" and to cold in the later "Quaternary," but good dating criteria have been lacking. Two recent boreholes at Dévaványa and Vésztö have provided a high-resolution paleomagnetic record in which magnetostratigraphic boundaries can be drawn and used to provide time controls. The first two fluvial cycles prove to be older than the Olduvai event and in terms of present usage would fall in the Upper Pliocene. Both holes have provided fair mollusk and ostracode faunas and moderate pollen records that make possible broad correlations with Jászladány.

The warm climate dominated until approximately 1.8 Ma, as in the marine record. Further studies are needed before a detailed evaluation of the climatic changes is possible, but the prospects are excellent.

The Jászladány borehole

In 1964/65 one of the early deep wells was drilled at Jászladány, in the northern part of the basin (Figure 19.3) and reached 950 m. It penetrated the entire "Quaternary" sequence, 420 m thick, consisting of fine-grained sand, silt, and clay, completely devoid of gravel or even coarse sand. The underlying Pannonian beds comprised variegated clays to a depth of 730 m, below which there were alternations of sand, silt, and clay. The sediments were relatively rich in paleontological material, including mollusks, ostracodes, and even some vertebrates, as well as pollen and spores, so that the locality has proved to be useful as a reference standard.

In his analysis of the results, Rónai (1969) drew attention to the existence of ten major cycles of deposition within the "Quaternary" sequence. Each cycle begins with coarse sedi-

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*Adapted from: Cooke, H. B. S. 1981. Age control of Quaternary sedimentary/climatic record from deep boreholes in the Great Hungarian Plain. In *Quaternary Paleoclimate*, ed. W. C. Mahoney, pp. 1–12. Norwich: Geo Abstracts Ltd. The figure numbers and reference citation style have been changed to conform to this work–Editor.

ments; the material then becomes progressively finer, although with fluctuations, until almost half of the sediment is fine clay. Thereafter it coarsens gradually, again with fluctuations, until the commencement of the next cycle. This is readily seen in Figure 19.4, which shows the variations in grain size in the Jászladány core and the nomenclature of major cycles. There are also some overall trends in the deposition which have led to recognition of three lithological divisions: a Lower Phase embracing cycles 1, 2, and 3, a Middle Phase comprising cyles 4, 5, 6, and 7, and an Upper Phase comprising cycles 8, 9, and 10. The boundaries between them are placed at 275 m and at 95 m.

These depositional cycles probably were related to tectonic factors that had a general impact on the basin as a whole. The delicate balance between subsidence and sediment supply had led to occasional subaerial exposure and the formation of minor soil horizons, shown in the second column of Figure 19.4, but they do not represent more than localized pauses of short duration in what is essentially an unbroken succession. There are also three periods of peat formation displayed in this hole, and in the vicinity of the 400-m level lignitic horizons occur. The present-day sedimentation in the vicinity of Jászladány is so similar to that shown by the borehole that it is reasonable to consider that the depositional environment has not changed significantly. Geodetic observations have demonstrated that subsidence of 5 mm occurred in a decade, and on that basis Rónai (1969) estimated that deposition of the entire 420-m fluvial succession occupied at least 1.3 to 1.4 m.y., a figure compatible with assignment of the whole unit to the Quaternary.

Dévaványa and Vésztö boreholes

Figure 19.5 summarizes the sedimentological and paleomagnetic data for the upper half of the Dévaványa and Vésztö boreholes and attempts to correlate them with the "standard" paleomagnetic time scale. The boundary between the Brunhes normal and Matuyama reverse epochs and that between the Matuyama and the Gauss are readily located in the plot of stable inclinations, and the Olduvai event is very clearly defined in both holes. The Jaramillo event is also recognizable, although there is a magnetically disturbed or oscillatory zone below it that is not usually seen in the marine records or in the lava sequences. Dotted lines have been drawn from these known age levels to the appropriate depth positions in the respective cores, and they serve to provide some means for estimating average sedimentation rates between them. Very fortunately, the paleomagnetic record shows a good deal of fine structure, including a number of short-period excursions not normally resolved. Several of these can be matched in the two cores and provide additional intermediate controls on the intercorrelation.

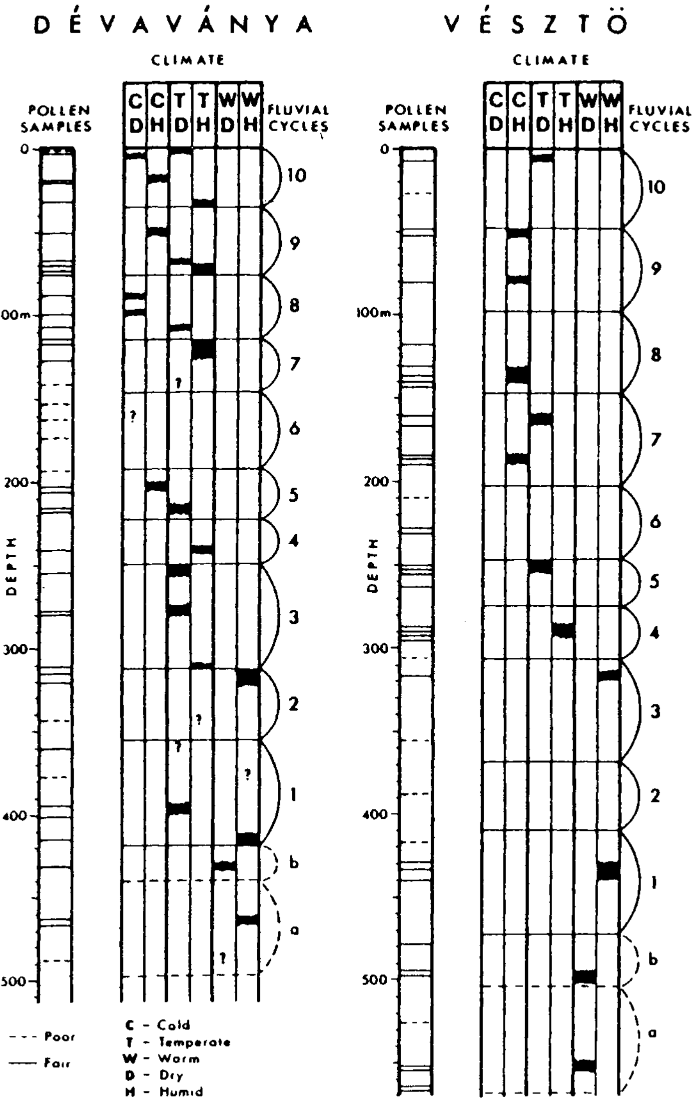

Figure 19.6. Paleoclimatic inferences from pollen samples in the Dévaványa and Vésztö boreholes. (From Cooke, 1981, fig. 4, with permission.)

Interpretation

Turning now to the sedimentological record, fluvial cycles can be distinguished in each of the holes, and the boundaries between them can be transferred to the time scale by drawing links parallel to the dotted lines of the paleomagnetic control points. The findings are in close accord for all the fluvial-cycle boundaries except that between cycle 6 and cycle 7, where there is a slight mismatch, albeit within the time span of the Jaramillo event. But this means that it is possible to estimate with reasonable accuracy the most probable ages of the boundaries between the fluvial cycles.

If the present concept of the Neogene/Quaternary boundary as lying just above the Olduvai event is confirmed (Haq, Berggren, and Van Couvering, 1977), then fluvial cycles 1 and 2 will belong to the Upper Pliocene, rather than to the Quaternary, as has been assumed hitherto.

Unfortunately, both Dévaványa and Vésztö have furnished rather scanty pollen records, with counts in the order of tens of grains per gram of sediment, and only a few with hundreds of grains per gram. Nevertheless, it is possible to make reasonable inferences about the climates in parts of the cores, and these are shown in Figure 19.6. Fluvial cycles 1 to 3 show generally warm

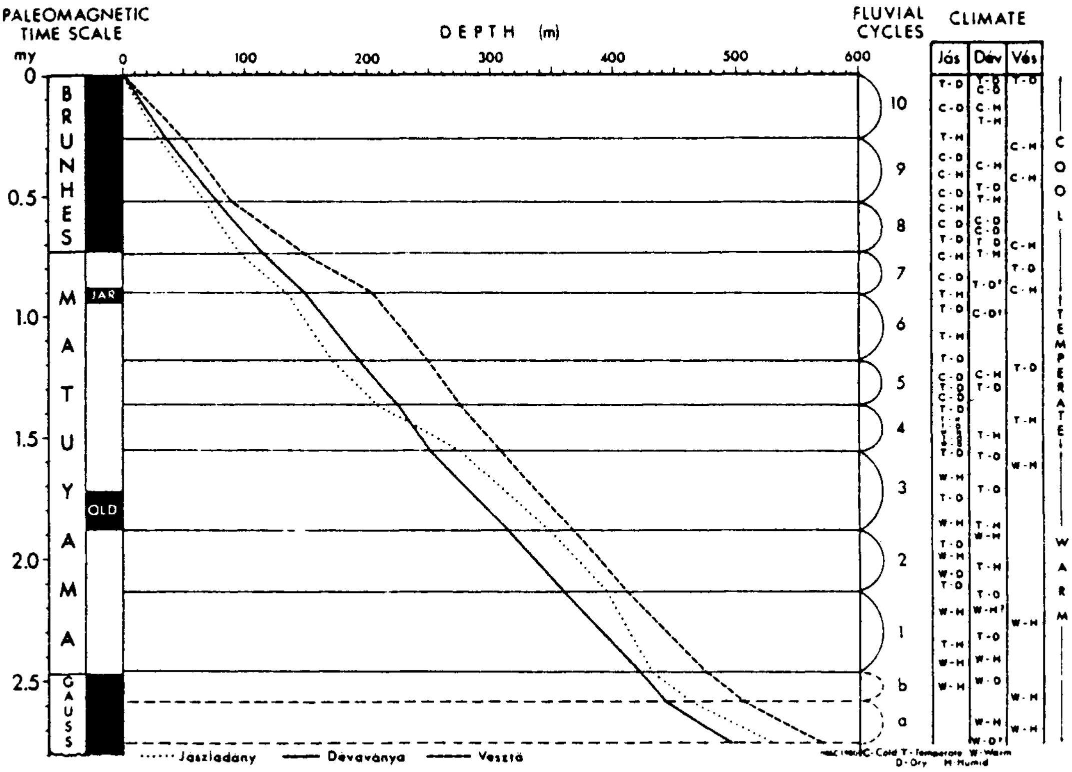

Figure 19.7. Depth/age plots for fluvial cycles in Jászladány, Dévaványa, and Vésztö boreholes, and paleoclimatic inferences derived from pollen analysis. (From Cooke, 1981, fig. 5, with permission.)

conditions, cycles 4 and 5 are temperate, and cycles 7, 8, 9, and 10 indicate a cool environment; data for the general character of the climatic changes leave little doubt that they parallel those of the Jászladány core and strengthen the belief that the fluvial cycles can be matched over a large part of the basin.

Figure 19.7 has been drawn on the assumption that the fluvial cycles match and that the age boundaries are reasonable approximations. For each of the three boreholes under consideration, the depths to the fluvial-cycle boundaries have been plotted against the estimated ages derived from the paleomagnetic data. The resultant curves show the apparent relative changes in sedimentation rates for each cycle and are in good agreement. It would appear that there was an above-average rate of sedimentation at Jászladány in cycle 4 and a below-average rate in cycle 1. There are also rather sharp changes in the sedimentation rates in all three holes below the b/a-cycle boundary [= Gauss–Matuyama boundary–*editor*], also noted in the earlier interpretation of the whole paleomagnetic record (Cooke et al., 1980). It must be pointed out that the relative linearity of the curves for Dévaványa and Vésztö may in part be an artifact of the procedure by which the cycle boundaries were derived by interpolation from the paleomagnetic control points; however, the boundary values would change very little if an attempt were made to "straighten" the Jászladány curve. It is also the case that soil horizons are much less apparent in the Dévaványa and Vésztö cores, so that the Jászladány sedimentation rate may indeed have been a little more variable. It is also clearly impossible, without more control points, to take any account of the variations in actual sedimentation rates that probably accompanied the changes in grain size within the cycles. Accordingly, it is considered that the age estimates for the boundaries of the fluvial cycles are not likely to be seriously in error.

On the right side of Figure 19.7, the climatic inferences for each of the three boreholes have been located in relation to their positions in the fluvial cycles. While they do not match in detail, they confirm Rónai's (1969) limits for the three major climatic divisions. Thus the warm climate dominated until approximately 1.5 Ma [= 1.7 Ma in the orbitally tuned time scale–*editor*], and the onset of consistently cool conditions began at about 0.9 Ma [= 1.0 Ma], just above the time of the Jaramillo event, although there is evidence for a cold spell at about 1.25 Ma [= 1.33 Ma]. This is in remarkably good agreement with the broad aspects of faunal and isotopic changes in the oceans, and it is worth quoting from the abstract to an important paper by Ruddiman (1971) on cores from the equatorial Atlantic:

> When applied to two cores containing 1.8 m.y. of equatorial sedimentary history, this analysis pinpoints two prominent, large-scale climatic shifts: (1) at 1.3 Ma BP, the mean climatic situation deteriorated, and short but severe cold pulses began to punctuate the previous moderate warmth of the late Matuyama; (2) following 900,000 Ka, the duration of cold intervals increased. Prior to the Jaramillo, no cold pulse exceeded 30,000 yrs; three post-Jaramillo cold intervals ranged in

duration from about 50,000 to 150,000 yrs. The shortest and most recent of these correlates with the Wisconsin glaciation.

Indeed, it is tempting to try to integrate the climatic inferences from the three Hungarian boreholes and then to compare them with Ruddiman's core, when a plausible correlation can be made. However, in point of fact the Hungarian observations are on too coarse a scale for such a comparison to be justifiable at the present time. It is clear that if a new core could be obtained from the pollen-rich Jászladány area, a closely spaced sampling program might yield a record as important as the remarkable one from the Grande Pile peat bog in France (Woillard, 1975, 1978). It has also been demonstrated very clearly that the traditional "four glacial" concept of the European Pleistocene is a gross oversimplification, and the story told by the loess stratigraphy of Central Europe is much more complicated, with 17 interglacial cycles since the Olduvai event (Kukla, 1975; Fink and Kukla, 1977). Doubtlessly, closer study of the Hungarian profiles will eventually resolve some of these problems.

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20 The Pliocene–Pleistocene boundary in Romania

CONSTANTIN GHENEA

Introduction

In recent years, detailed geological studies of the Pliocene–Lower Pleistocene interval in Romania, including modern biostratigraphic analysis, have been combined with paleomagnetic data. At present, the chronology of the interval between 3.8 and 1.0 Ma in the Dacic Basin is well known, correlations being possible not only with the Euxinic Basin (Ukraine) but also with western Europe.

The Dacic (or Dacian) Basin is the name given to a vast sedimentary basin bounded by the southern Carpathians, the Balkans, and the Danube (Figure 20.1). In the middle Pliocene, about 3.8 Ma, the salinity of the waters filling that basin was reduced, a trend that became more pronounced in the late Pliocene. Strata with fresh-water fauna were deposited, and at certain levels beds rich in fossil mammal faunas were deposited. This is the Romanian Stage, and data obtained in the past few years are basic to correct interpretation of the stratigraphy of this interval.

In most of the Dacic Basin, deposition of continental molasse, known as the Cîndesti Formation, began in the late Pliocene and continued into the early Pleistocene. This formation occurs throughout the basin, except for the western part, where lacustrine conditions prevailed. Recently, the stratigraphic limits of the Cîndesti deposits were correlated to the paleomagnetic scale, but no stratigraphic subdivisions have been identified within this complex.

Above the Cîndesti Formation, in the central and southern parts of the Dacic Basin, deposits of a mixed, fluviolacustrine regime developed in the early and middle Pleistocene; paleontological evidence at various levels permits their correlation with the formations of the Euxinic Basin (Table 20.1).

Chronostratigraphy of the 3.8–1.0-Ma interval

The major part of the Plio–Pleistocene interval is occupied by the Romanian Stage, from about 3.8 Ma to 1.8 Ma. This term was first designated by Krejci Graf and redefined by Mihaila and Andreescu (Andreescu, 1981) for the period of freshening water in the Dacic Basin. The brackish-water mollusk faunas disappeared and were replaced by fresh-water lacustrine faunas.

The following biostragraphy is based on fresh-water mollusk faunas studied by Andreescu (1981). These have been correlated with paleomagnetic data by Ghenea et al. (1982; Alekseeva et al., 1981), with the paleomagnetic boundary ages revised in this chapter according to the orbitally tuned calibration of Cande and Kent (1992).

Romanian Stage

Lower Romanian. In the Carpathian Bend, near Beceni, an alternation of sands, clays, and marls with lignite intercalations about 250 m thick conformably overlies the deposits of the Upper Dacian Stage (Andreescu, 1981). The fossil content is mainly smooth unionids of the *slanicensis, sturdzae*, and *brandzae* groups, as well as *Viviparus bifarcinatus*. The Lower Romanian has the same lithologic and paleontological features in the central part of the Dacic Basin and in its western extremity, where lignitic strata are also present.

On the Moldavian Plateau, a fluviolacustrine sequence equivalent to the Lower Romanian contains a mammalian fauna characterized by the following taxa: mastodons *Anancus arvernensis* and *Zygolophodon* (= *Mammut*) *borsoni*; the earliest arvicolid, *Promimomys moldavicus*; horses of the genus *Equus* together with the genus *Hipparion*; and *Tapirus arvernensis* and *Sus provincialis*. A mollusk fauna with *Psilunio sibinensis* occurs in the same formations.

In the Brasov intermontane basin, a lacustrine sequence consisting of marls, clays, and sands, with lignite intercalations, is also considered to be equivalent to the Lower Romanian. At Capeni and Virghis, local mammal faunas include *Anancus arvernensis, Zygolophodon* (= *Mammut*) *borsoni, Dicerorhinus* cf. *D. leptorhinus, Equus* (*Macrohippus*) *sylvanum, Sus provincialis, Protarctos boeckhi, Parailurus anglicus, Prospalax priscus, Mesopithecus monspessulanus*, and others (Samson and Radulescu, 1973).

The mollusk and mammal faunas found in the Lower Romanian can be correlated with the lower part of the Moldavian complex in the former Soviet Union. This is in agreement with recent paleomagnetic analyses, which have

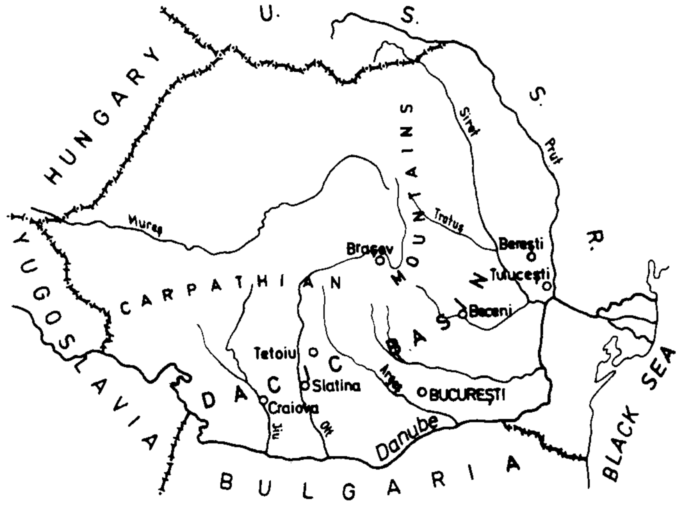


Figure 20.1. Localities mentioned in the text.

established that the formations of the Lower Romanian were deposited between 3.8 Ma and 3.4 Ma (Table 20.1).

Middle Romanian. The Middle Romanian is marked by a major change in the fresh-water mollusks: the appearance and further development of the Levantine thermophilous fauna (Tshepalyga, 1972), characterized by numerous genera and subgenera of sculptured unionids. At Craiova in the western Dacic Basin, sands and clays with sculptured unionids form the stratotype of the Levantine Stage. Current knowledge allows us to consider that this stage, as previously used in the relevant Romanian literature, is the equivalent of the entire Middle Romanian.

The typical Middle Romanian in the western part of the Dacic Basin, in the Jiu Valley, contains an extremely varied and rich mollusk fauna consisting of *Rugunio lenticularis, Potamides porumbarui, P. herjeui, Cuneopsidea recurvus, C. vukotinovici, C. sculpta, C. iconominanus, C. beyrichi, C. doljensis, Rytia bielzii, R. brandzae, R. slavonica, Wenziella clivosa, W. subclivosa, W. cymatoidea, W. gorjensis, Rugunio condai, R. turburensis, R. mojsvari, R. pilari, Pristinunio pristinus, P. davilai, Cyclopotomida munieri, C. pannonica, Psilunio craiovensis, P. altercarinatus, Ebersininaia stefanescui, Viviparus bifarcinatus, V. stricturatus, V. rudis, V. turgidus, V. strossmayerianus, V. craiovensis*, and *V. mammatus*, among others. The biozonation of the Middle Romanian is based on the unionid fauna (Andreescu, 1981), and the definition of zones and subzones permits good correlations with the equivalent formations in the Ukraine (Tshepalyga, 1972).

In the western extremity of the Dacic Basin (the Jiu-Danube interfluve), deposition of coal began with the Dacian Stage of the middle Pliocene, continued through the whole Romanian, and ended in the early Pleistocene. Sporopollen sequences within the coal sequence show two characteristic reference horizons. The first, at the level of stratum VI (the Dacian–Lower Romanian boundary), is marked by substantial reductions in the *Liquidambar* pollen and the Myricaceae–Cyrillaceae association. The second, between strata VII and VIII, is marked by the maximum development of the Myricaceae–Cyrillaceae association. The palynomorph suite in the upper part of the Romanian is characterized by the dominance of conifers such as *Picea, Cedrus, Pinus, Tsuga, Abies*, and *Larix*, indicating cold climate (Roman, in Marinescu, Ghenea, and Papaianopol, 1981), or, more probably, higher elevations in the source areas from which the sediments were derived, as the geological evidence suggests.

In the Middle Romanian, fossil mammal remains have been found at several locations in the Dacic Basin, including the deposits of Cernatesti and Tulucesti. In addition to the two mastodons *Anancus arvernensis* and *Zygolophodon borsoni*, the faunas include some typical middle Villafranchian species, such as *Dicerorhinus etruscus, Equus stenonis, Cervus issiodorensis*, and the earliest elephant, *"Archidiskodon rumanus,"* which, properly speaking, is a species of *Mammuthus* (Aguirre et al., Chapter 9, this volume). In the Brasov intermontane basin, remains of *Dicerorhinus jeanvireti* and *D. etruscus* were mentioned by Samson and Radulescu (1973), within the "Iaras Sand Formation," which is assigned to the Middle Romanian.

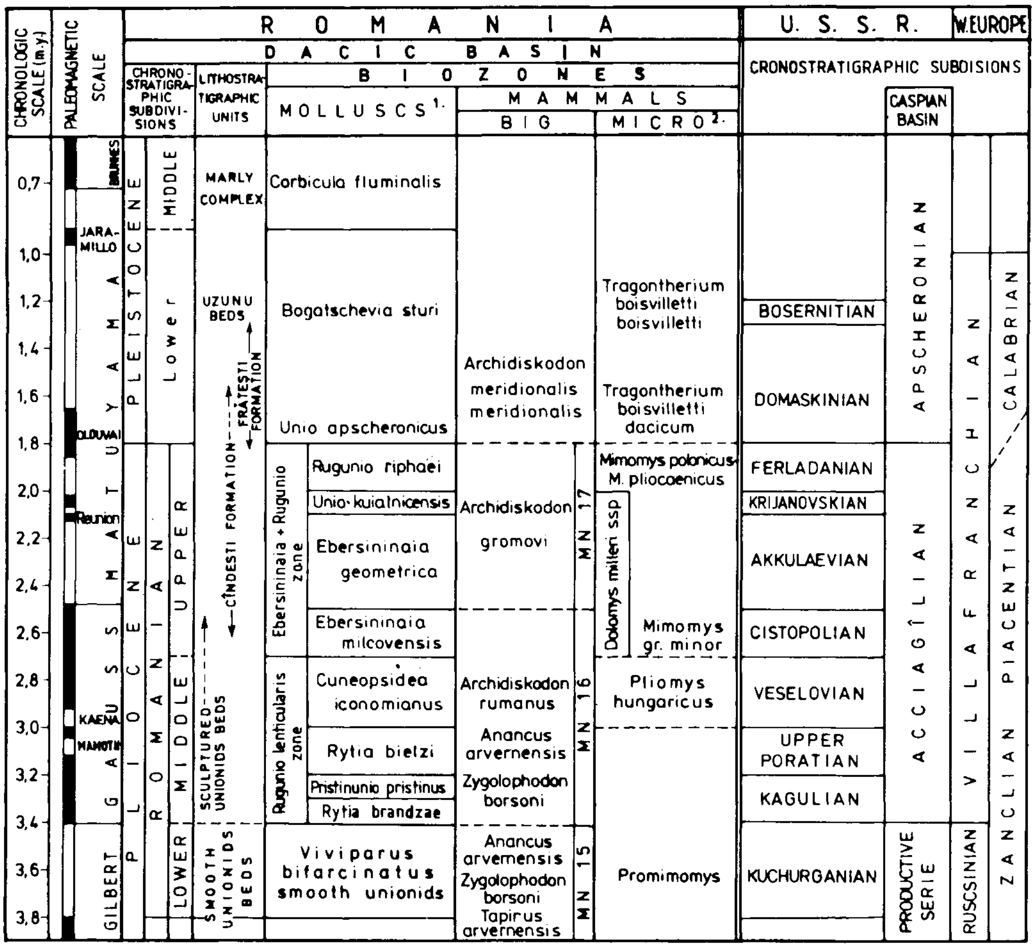
Upper Romanian. In the western part of the Dacic Basin, lacustrine conditions are still represented in the Upper Romanian strata, consisting of a clayey-sandy sequence with a fresh-water mollusk fauna. The latter is marked by the disappearance of most of the typical "Levantine" genera and species of the Middle Romanian and by the appearance of new genera, such as *Bogatschevia* and new species such as *Ebersininaia milcovensis, E. geometrica, E. struevi, Unio kujalnicensis*, and *Rugunio riphai*.

In the Olt Valley, at Slatina, the sequence of clays, clayey sands, and gravels representing the Middle and Upper Romanian is highly fossiliferous. Paleontological evidence (mollusks, micromammals) and paleomagnetic correlations provide a detailed stratigraphy across the Pliocene–Pleistocene boundary (Andreescu et al., 1981). The Upper Romanian is represented by strata with *Ebersininaia milcovensis, E. geometrica*, and *Rugunio riphai*. The boundary between the strata with *Rugunio riphai* and those with *Unio apscheronicus* seems to correspond to the base of the Olduvai subchronozone.

The Upper Romanian strata of the Olt Valley also contain a rich fauna of fossil mammals. The large mammals include an archaic elephant, which, according to Radulescu and Samson (in Andreescu et al., 1981), is similar to *Mammuthus* (= *"Archidiskodon") gromovi* and is characteristic of the Kotlovina level in the former Soviet Union.

Small mammals are present at several levels in the Olt Valley, at Slatina and Cherlesti (Feru, Radulescu, and Samson, 1978). The lower level (Slatina 1) is characterized by *Desmana kormosi, Apodemus* sp. 1, *Dolomys milleri* subsp. 1, and *Mimomys minor* subsp. 1; *D. milleri* dominates this association. The upper level at Slatina (Slatina 2) contains *Desmana nehringi, Talpa fossilis, Beremendia fossidens*, Leporidae cf. *Hypolagus brachygnathus*,

Table 20.1. *Upper Pliocene and Lower Pleistocene chronostratigraphy in Romania*



1. For Romanian stage-Andreescu 1981
2. Samson and Rădulescu 1981

Apodemus sp., *Dolomys milleri milleri*, and *Mimomys minor* subsp. 2. The reduction of *D. milleri* and the prevalence of *Mimomys* are characteristic.

At Cherlesti, 13 km north of Slatina, the Upper Pliocene (between 2.0 and 1.8 Ma) includes a faunal association of *Desmana nehringi, Allactaga ucrainica, Dolomys milleri milleri, Mimomys* ex gr. *M. polonicus–pliocaenicus*, and *Mimomys minor* subsp. 2 (Radulescu and Samson, in Andreescu *et al.*, 1981).

Cîndesti Formation. At the beginning of late Romanian time, the continental molasse known as the Cîndesti Formation, made up of alternating gravels, conglomerates, sands, and clays, developed over a vast area in the Dacic Basin. In many places, its thickness reaches hundreds of meters, with a maximum of 1,000 m in the Carpathian Bend zone.

The extent and thickness of the Cîndesti Formation in the Carpathian foreland are the results of massive uplift of the Carpathians and the resulting intensification of erosion and deposition. The stratigraphic position of the Cîndesti Formation is difficult to establish because paleontological evidence is extremely scarce (mainly, isolated remains of *Mammuthus meridionalis*). Nevertheless, the stratigraphic limits have been estimated through paleomagnetic profiles in sections where sedimentation was continuous during the Romanian–early Pleistocene interval. In the Carpathian Bend zone, coarse-grained intercalations have been assigned to the "Cîndesti Formation" as far back as the interval between 2.7 Ma and 2.5 Ma (Alekseeva et al., 1981).

Stratigraphic relations in the Carpathian region indicate that the Cîndesti type of deposition continued up into the early Pleistocene over a wide area. In that respect, the previous location of the Pliocene–Pleistocene boundary within the Cîndesti Formation has no important geological or climatic significance. The boundary near the top of the Olduvai event, at about 1.80 Ma (Pasini and Colalongo, Chapter 2, this volume), however, seems to have been associated with evidence of important changes.

Correlation of Middle Romanian and Upper Romanian

The biostratigraphic divisions of the Romanian based on the fossil mollusk and mammal faunas and their calibration with the paleomagnetic scale now in use allow detailed correlations with the Upper Pliocene and the Lower Pleistocene of the Euxinic and Caspian basins. Thus, the Lower Romanian–Middle Romanian boundary is almost the same age as the division between the Gilbert and Gauss magnetic epochs, at about 3.6 Ma, providing a criterion that can be used for the Dacic Basin. The Middle Romanian, which is correlated by its mollusk faunas with the Kagulian, upper Poratian, and Veselovian horizons in the former Soviet Union, also corresponds in its mammal fauna to the Middle Villafranchian (zone MN-16) of the Mediterranean basin. The Upper Romanian corresponds to the upper part of the Akchagylian and to the Chistopolian, Akkulaevian, Krizhanovskan, and Ferlandanian horizons in the Ukraine; its base is considered to be at about 2.6 Ma, near the Gauss–Matuyama boundary. The upper boundary of the Romanian, represented by the boundary between the strata with *Rugunio riphai* and the overlying strata with *Unio apscheronicus*, coincides with the base of the Olduvai event of the Matuyama epoch.

Regional correlation of the Romanian Lower Pleistocene

In the western part of the Dacic Basin, the lacustrine conditions, evidenced by strata with unionids and fossil mammals, seem to have ended at the *Unio apscheronicus* level, close to 1.9 Ma. In the central and southern parts of the Dacic Basin, as already stated, that is equivalent to the lower alluvial deposits of the Cîndesti Formation, which continue into the Pleistocene. In the eastern part of the Dacic Basin, drillings in the Bucuresti area have shown the presence, under younger formations, of 100–200 m of lacustrine beds similar to those of the western Dacic Basin, at depths ranging between 100 and 300 m. Those accumulations were given the name of "Fratesti Beds" by Liteanu in 1953 (Marinescu et al., 1981). Regarding the paleontological content, the Fratesti interval is characterized by rich mammal faunas which allow correlations to the European scale. The best-known deposit is situated at Tetoiu, better known in the literature as Bugiulesti. There, the mammal association is marked by the presence of a single elephant, *Mammuthus meridionalis*, together with *Paradolichopithecus geticus, Equus (Allohippus) stenonis, Dicerorhinus etruscus, Ursus etruscus, Nyctereutes, Trogontherium boisvilleti, Megalovis, Leptobos, Libralces gallicus*, and *Canis etruscus* (Samson and Radulescu, 1973).

This ensemble might seem to be derived from two levels. An earlier warm-climate part, with the last *Nyctereutes*, could be correlated to the older part of the Odessan complex of the Euxino-Caspian realm and with the younger part of the middle Villafranchian, for example the Senèze fauna and its correlatives in western Europe, which under current definition are placed in the latest Pliocene (Aguirre et al., Chapter 9, this volume; Azzaroli et al., Chapter 11, this volume; Nikiforova, Chapter 21, this volume). A younger level, with a cold-climate fauna indicated by *Canis etruscus*, would be equivalent to the Olivola phase in the transition to Upper Villafranchian (Azzaroli et al., Chapter 11, this volume).

In some areas of the southwestern Dacic Basin, the upper part of the "Fratesti Formation" consists of sandy deposits with clay-marly interbeds, the so-called Uzunu Beds, showing a mollusk fauna including *Bogatschevia sturii, Corbicula fluminalis, Pisidium amnicum, Viviparus craiovensis, V. diluvianus, V. geticus, V. romaloi, Litoglyphus naticoides, Planorbis planorbis*, and others. The presence of the fauna with *Bogatschevia sturii* allows the correlation of the Uzunu Beds to the Boshernitskan horizon of the Pleistocene in the former Soviet Union, dated to about 1.3–1.2 Ma.

Problems of the Pliocene–Pleistocene boundary in Romania

In the past, the main problem in establishing the Pliocene–Pleistocene boundary in the Dacic Basin of Romania concerned disagreements about correlating Villafranchian mammal faunas to the Calabrian marine deposits, considered to represent the Pleistocene base. With respect to subdivisions of the Villafranchian, which are well represented in Romania, the Pliocene–Pleistocene boundary has been recognized variously at about 3.8 Ma, with the appearance of the first Villafranchian elements, at about 2.6 Ma, with the appearance of the earliest Middle Villafranchian faunas, or at about 1.8 Ma, with the appearance of mammal assemblages comparable with the Senèze and Olivola faunal phases, termed "upper Middle Villafranchian" by Azzaroli et al. (Chapter 11, this volume), but historically considered as Upper Villafranchian in the Romanian literature.

Applying the recommendations of the working group for IGCP-41 to Romania, the acceptability of these various proposals can be summed up as follows:

1. The first Lower Villafranchian fauna (upper Ruscinian–Csarnotian transition) is slightly older than the Gilbert–Gauss boundary. That paleomagnetic reversal, at 3.6 Ma, corresponds to the boundary between the Lower and Middle Romanian and was associated with an important change in the fresh-water mollusk fauna to include a multitude of sculptured unionids representing a markedly thermophilous fauna, and obviously demonstrating a climatic optimum. It is worth noting that the mammal faunas at Beresti-Malusteni and Capeni-Virghis, in which the appearances of the "Villafranchian" elements are considered to mark an important change, are situated within the Lower Romanian, below the Gilbert–Gauss boundary.
2. The first Middle Villafranchian fauna is approximately coeval with the Gauss–Matuyama boundary, at 2.6 Ma, and thus with the boundary between the Middle

Romanian and Upper Romanian. It was marked by cold climate throughout the Paratethys, manifested by the disappearance of the thermophilous mollusk fauna with sculptured unionids. Within the interval between 2.7 Ma and 2.5 Ma, important orogenic movements in the Carpathians brought about the deposition of the Cîndesti continental molasse over a vast area of the Carpathian foreland. It may also be supposed that the intensification of erosion and the deposition of large amounts of materials indicate a great increase in runoff as a result of higher precipitation in the cold climate.

3. Although originally considered to be older, the first cold-climate mammal faunas of the Upper Villafranchian (as this term is understood in Romania) and the top of the Upper Romanian are now placed at the top of the Olduvai subchronozone, approximately 1.77 Ma. At present that is the official criterion adopted in Romania for the base of the Pleistocene, although recent studies suggest that the Pliocene–Pleistocene boundary-stratotype at Vrica, Italy, is in fact slightly older, with an age of about 1.80 Ma within the upper Olduvai (Pasini and Colalongo, Chapter 2, this volume). The concept appears to be the most conventional, since it is not based on a climatic or biostratigraphic criterion, but on correlation to a fixed point, the Vrica stratotype.

In most of the Dacic Basin, the designated Upper Romanian–Pleistocene boundary falls within the typically undifferentiated and generally unfossiliferous conglomerates and sands of the Cîndesti Formation, with the result that in those deposits geological mapping of the boundary is an insoluble problem. On the other hand, in western Romania, and again in the "Fratesti Beds" of the Bucuresti area, biostratigraphic sequences of Upper Villafranchian mammals and paleomagnetic data allow correlation to the presently defined Pliocene–Pleistocene boundary.

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21 The Pliocene and Pleistocene of the European part of the Commonwealth of Independent States

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Introduction

Within western Russia and the adjoining states, the Upper Pliocene and Quaternary (Anthropogene) sequences are abundantly documented in terms of fossil mammals, marine and fresh-water mollusks, foraminifera, ostracodes, macro- and microflora, and remains of human occupation, as well as extensive data on neotectonics, lithology, and paleomagnetism. For many years, however, the stratigraphers in that vast region did not share a unified point of view on the location of the lower boundary of the Quaternary. Although the majority considered that the lower boundary of the Quaternary should be correlated to the base of the Bakuan Stage, some favored the base of the Akchagylian Stage, whereas others preferred the base of the Apsheronian.

At a joint meeting of the IGCP-41 working group and INQUA Subcommission 1-d at the XI INQUA Congress in Moscow in 1984, the proposal was adopted to place the Pliocene–Pleistocene boundary, and thus the Neogene–Quaternary (N/Q) boundary, in a physical reference point, or boundary-stratotype, at Vrica, Calabria, located at the base of the claystone layer conformably overlying sapropelic marker *e* in component-section B of Selli et al. (1977) (Aguirre and Pasini, 1985). According to Tauxe et al. (1983), with slight modification by Zijderveld et al. (1991), that level is close to the top of the Olduvai normal-polarity subchron. Following that concept, the N/Q boundary should be located somewhere in the lower part of the Apsheronian beds of the Caspian Basin and their stratigraphic equivalents. That boundary level has now been officially accepted by stratigraphic workers in the former USSR.

The Eopleistocene. As is well known, the "Eopleistocene," as used throughout the former USSR and in most of eastern Europe, corresponds to the Lower Pleistocene in western European and North American stratigraphy. According to the correlation of the Vrica definition, the lower part of the Apsheronian regiostage contains the base of the Eopleistocene and the boundary of the Quaternary. That boundary concept was used in preparing a detailed chronostratigraphic scheme for late Cenozoic deposits in European Russia and adjacent states (Nikiforova et al., 1976, 1980; Nikiforova and Alexandrova, 1991). This scheme can be successfully applied to most, if not all, of northern Eurasia, although with different degrees of precision.

A correlation of the Vrica boundary-stratotype to a level within the lower half of the Apsheronian, rather than at its very base, is supported by paleomagnetic analysis. Gurarij, Pevzner, and Trubikhin (1973) correlated the normally magnetized beds within the lower parts of the Apsheronian sequences in Azerbaijan and Turkmenia with the Olduvai paleomagnetic event, which is identified with the stratigraphic interval just below the proposed boundary-stratotype at Vrica (Pasini and Colalongo, Chapter 2, this volume). Similar correlations have been made by Pevzner and Tshepalyga (1970) through paleomagnetic analysis of matrix adhering to Apsheronian mammal fossils collected from the terraces of large rivers in Moldavia and in southern Ukraine (the Dniester, Danube, and Prut rivers), as well as from material from other regions. Additional information comes from the work of V. M. Trubikhin (Nikiforova, 1987).

The southern part of the region is quite favorable for recognizing the lower boundary of the Quaternary. Although direct comparisons of marine deposits in the Caspian and Black Sea basins with those in the Mediterranean are hampered by the dissimilarities of the marine invertebrate faunas, correlations can be made with the help of the mammal and fresh-water mollusk faunas, augmented by radiometric dating, stable-isotope curves, and particularly paleomagnetic sequencing.

Pre-Quaternary stratigraphy

The Permanent Quaternary Commission of the former USSR Interdepartmental Stratigraphic Committee suggested new terminology for the subdivision of chronozones, which in the Soviet scheme have the scale of epochs. In English, these units, in order of descending hierarchy, are divisions (i.e., sub-chronozones); links; over-steps (i.e., super-steps); and steps. The latter two are the equivalents of over-horizons and horizons in regional biostratigraphic schemas. In this terminology, the Upper Pliocene is considered to be a division (i.e., the "upper division of the Pliocene"). The lower division of the Pliocene includes the

Cimmerian, or at least its upper part, the Kamyshburunsky, and the Panticapeisky horizons (Semenenko and Pevzner, 1979).

The Upper Pliocene division is composed of two links. The lower link, equivalent to the lower Akchagylian, corresponds to continental facies with the Moldavian fossil mammal complex. In those deposits, the key mammal taxa are *Anancus arvernensis, Dicerorhinus jeanvireti*, archaic elephants of the genus *Archidiskodon* (or *Mammuthus* according to some specialists; see Aguirre et al., Chapter 9, this volume), species of the microtine rodents *Dolomys* and *Pliomys*, and in the uppermost levels the first *Mimomys* (e.g., *M. polonicus*) (L. P. Alexandrova, in Nikiforova, 1987; Chaline, Chapter 14, this volume). The upper link includes the middle and upper Akchagylian, which correspond to continental facies with Khaprovian mammal assemblages. The key forms of the Khaprovian are *Archidiskodon gromovi*, diverse species of *Mimomys*, including *M. pliocaenicus*, and in upper Khaprovian levels the first *Villanyia*, as *V. lagurodontoides*. The boundary between the Moldavian and Khaprovian is close to the Gauss–Matuyama boundary, at about 2.6 Ma.

The Moldavian mammal assemblages are similar to those of the late Ruscinian (Csarnotian) and Lower Villafranchian faunas of southwestern Europe and to the Reuverian of the northern European scale. In western Siberia, the Betekian complex appears to correspond to the uppermost phase of the Moldavian complex (Veselovsky horizon), but older phases are missing. Deposits with Khaprovian mammals are mostly correlative to the upper part of the Lower Villafranchian unit in western Europe (characterized by the fauna of Montopoli) and possibly, in the uppermost Khaprovian, to part of the Middle Villafranchian. It is probable, however, that the FAD (first-appearance datum) of *Archidiskodon meridionalis*, seen in the Psekups fauna, is younger than the Khaprovian, although older than the Odessan complex. The first appearance of this proboscidean is a marker for the Middle Villafranchian in western Europe, as at Saint-Vallier, Senèze (lower level), and Tegelen (Azzaroli et al., 1988; Aguirre et al., Chapter 9, this volume; Azzaroli et al., Chapter 11, this volume). In western Siberia the Podpusk-Lebyashinsky faunal complex is roughly equivalent to the Middle Villafranchian (Alekseev, Chapter 22, this volume).

Eopleistocene and Pleistocene stratigraphy

In the Eopleistocene, two links are distinguished. The lower link embraces the lower Apsheronian and most of the middle Apsheronian, corresponding approximately to the continental facies with the Odessan complex of mammals, in which *Archidiskodon meridionalis meridionalis* is a key element. The presence of the first rootless-toothed voles, *Allophaiomys* and *Prolagurus*, indicates that the Odessan complex is equivalent to the early Upper Villafranchian. In particular, we may correlate the fauna of the Domashkinian horizon at the base of the Odessan complex to the earliest level of the Upper Villafranchian, represented by the Olivola fauna, to the lower Eburonian climatic stage of The Netherlands, and to the Kizikhan complex of western Siberia, by the presence of *Allophaiomys pliocaenicus* and *Prolagurus (Lagurodon) arankae* (Azzaroli et al., 1988; Chaline, Chapter 14, this volume). The beginning of the Eburonian is located in The Netherlands just at the top of the Olduvai event (Zagwijn, 1974, 1985; Zagwijn, Chapter 16, this volume). According to this interpretation, leading from biostratigraphy to magnetostratigraphy, the Domashkinian cool-climate deposits thus closely coincide with the strata at the base of the Quaternary in the marine sequence at Vrica.

The upper link of the Eopleistocene is composed of the uppermost middle Apsheronian and the upper Apsheronian, correlative to the Tamanian complex of mammals in continental facies of European Russia and the adjacent states. That interval in western Europe is represented by the transitional or Epivillafranchian sequences, and in western Siberia by the Razdolinian complex. In those deposits the key taxa are *Archidiskodon meridionalis tamanensis* and the related western subspecies *cromerensis* and *vestinus* (Aguirre et al., Chapter 9, this volume; Azzaroli et al., Chapter 11, this volume), and *Microtus (Pitymys)* appears in the small-mammal faunas. The boundary between these two links coincides approximately with the base of the Jaramillo subchron.

The boundary at the top of the Eopleistocene, formerly considered to be the base of the Pleistocene and the beginning of the Quaternary by many workers in the former USSR, lies at the base of the Tyurkanskaya suite, at the beginning of the Bakuan regiostage. In continental facies, that is correlative to the deposits with the earliest elements of the Tiraspol faunistic complex (i.e., faunas with *Microtus raticepoides*), equivalent to the latest Biharian and earliest Cromerian of western Europe (L. P. Alexandrova, in Nikiforova, 1987). Strata of this age have negative remanent polarity at Solilhac and Tiraspol and are thus older than the Brunhes–Matuyama. At an earlier level, the "Betfia" phase of early Biharian age marks the end of the Villafranchian of western Europe; it is characterized by cold-adapted assemblages which appear to correspond to a glacial-climate (Menapian) phase in the uppermost Matuyama, above the Jaramillo.

The Russian Pleistocene (equivalent to the Middle and Upper Pleistocene in western Europe and North America) consists of three links. The lower link, as found throughout the region west of the Urals and in the Caspian and Black Sea basins, is characterized by the main part of the Tiraspol fossil mammal complex. In the main Tiraspol fauna there are at least two subdivisions in northern Eurasia, with the lower part equal to the Bakuan regional stage and the "true" Cromerian or Günz-Mindel interglacial interval of western Europe, and the upper, or Oksky, part of a cold-climate faunal interval corresponding to the Elsterian or Mindel glacial-climate period of western Europe, the beginning of which is dated at about 0.54 Ma. Thus, in the scheme of the European part of the former USSR, at least eight horizons (or steps) are distinguishable in the Lower Pleistocene.

The Anthropogene question

The earliest evidence of humans, or hominids, is of special significance because Soviet scientists have long emphasized a connection between the Quaternary sub-period and human activity. On the African continent, the oldest pebble-tool cultures date to more than 2.3 Ma, and the earliest acknowledged remains of true *Homo*, presently classified as *H. rudolfensis*, are even older (Aguirre, Chapter 10, this volume). Outside of Africa, however (i.e., north of the Levant), the appearance of humans in the form of *Homo erectus* is evidenced by tools or fossil remains from sites at Sangiran in Java, Gongwangling and Nihewan in China, Dmanisi in Georgia, and possibly Orce and Cueva Victoria in Spain. All of those southern Eurasian sites are older than the Jaramillo subchron, and the oldest may be almost as old as the first evolutionary appearance of the *H. erectus* clade in Africa at about 1.6 Ma (Aguirre, Chapter 10, this volume). Although the rarity of such sites leaves a small margin of uncertainty, we may be justified to conclude that *Homo* did not become established in (southern) Eurasia until after the beginning of the Quaternary, as presently defined, and before the Quaternary boundary previously adopted in the former USSR at the base of the Bakuan regiostage in the uppermost Matuyama, above the Jaramillo.

Climate changes

Although correlating the deposits of the Pliocene and Lower Quaternary according to continental glacial episodes is a debatable means of dating, climatic fluctuations during that interval are recognized by all scientists.

In marine deposits, the late Pliocene (Piacenzian) of southern Europe is characterized by warm-climate invertebrates, although with some colder fluctuations near the Gauss–Matuyama boundary. In terms of the late Pliocene mammalian faunas, the first deterioration of the climate was seen in the early Villafranchian, and climatic oscillations tended to become more significant during the middle Villafranchian. An appreciable cooling in the beginning of the middle Villafranchian was marked by a major erosional phase, the Aquatraversan, in central Italy (Bigazzi, Bonadonna, and Iaccarino, 1973) and by sharp changes in vegetation distinguished by the regional extinction of *Taxodium, Sequoia*, and *Sciadopitys*, marking the base of the Tiberian paleoflora (Lona and Bertoldi, 1973). In The Netherlands, the Pretiglian deposits with "cold" flora, dated paleomagnetically to just above the Gauss–Matuyama boundary, correspond to that time (Zagwijn, 1985; Zagwijn, Chapter 16, this volume). In the Soviet schema, the late Pliocene cold period corresponded to a cold phase in the middle Akchagylian, represented by the Chistopol horizon of the Kinel Formation of the Middle Volga and Kama regions. The deposits in that horizon are characterized by taiga flora (Nikiforova et al., 1976).

A lesser cooling seems to be reflected in the specific features of the Tiglian B horizon in The Netherlands and the lower Kuyalnik beds from the Odessa region (Kryzhanovka horizon) containing a cold-adapted mollusk fauna.

The severe climatic deterioration associated with the classic concept of the Calabrian is correlative to the "cold" fauna of Olivola (Azzaroli et al., Chapter 11, this volume) and the "cold" (upper) faunal levels at Senèze (Bout, 1970) at the beginning of the Upper Villafranchian and also to "cold" flora of the Rhine Valley Eburonian, both correlated to the top of the Olduvai subchron (Zagwijn, 1974, 1985). As noted earlier, in our region that cooling corresponds to the very beginning of the Apsheronian (Domashkinian horizon), represented by the Paludina beds with a boreal molluscan fauna overlying the marine upper Akchagylian in the Lower Volga region. The lower Apsheronian of the Duzdag has also a similarly boreal fauna (Nikiforova et al., 1976).

In conclusion, the base of the Upper Villafranchian in Italy, the base of the Apsheronian in the western part of the former USSR, and the base of the Eburonian cold-climate stage of the North Sea Basin correlate in terms of paleoclimate, paleontology, and paleomagnetism to the base of cold-climate marine deposits at Vrica, Italy, that conform to the traditional concept of the Calabrian Stage designated by the London Commission (IGC 1948) to define the beginning of the Pleistocene. This level, which is close to the top of the Olduvai normal-polarity chron, represents the beginning of the Quaternary wherever it can be identified, and the cooling at that time can be traced on a global scale.

Following the cold-climate Domashkinian (lowermost Apsheronian) horizon, the beginning of the middle Apsheronian is characterized by impoverished spore-pollen spectra in which the broad-leafed species decline and disappear. The pollen indicate a warming in the upper part of the middle Apsheronian which correlates with the warm Baventian stage of The Netherlands. The cooling of the Morozovka horizon of the upper Apsheronian (marked by the complete disappearance of pollen from broad-leaf trees and the predominance of pollen from steppe vegetation) corresponds to the Linge and Dorst glacial episodes.

Invertebrate biostratigraphy

Climatic fluctuations are also reflected in the molluscan faunas in continental deposits in the southern European parts of the former USSR, which correlate with the Akchagylian and Apsheronian (Table 21.1). According to Tshepalyga (1972), the two major stages, called by that author the Levantine and the Pleistocene, can be distinguished according to the evolution of molluscan faunas in the East European province. Characteristic of the Levantine stage are the appearances of new genera and subgenera. The most significant changes in the evolution of fresh-water mollusks are seen in the Kishlitzian (Kimmerian) and upper Poratian horizons, including the appearance and development of sculptured forms (thermophilic fauna) during the early and middle Akchagylian (early Villafranchian). By the midpoint of the middle Akchagylian (beginning of the middle Villafranchian) the fauna had already changed. The Levantine elements had disappeared, and the typical boreal fauna without "warm" elements had come into being, as seen in the Chistopol horizon. By the end of the middle Akchagylian transgression climax, the thermophilic molluscan fauna had returned, only to

Table 21.1. *Mollusk zonation of the Plio–Pleistocene sequence in southwestern Russia and adjoining states*

Epoch	**Stage**	**Complex**	**Biozone**	**Age (Ma)**
LOWER PLEISTOCENE	**Bakuan**	Kolkotovian	*Pseudunio moldavica*	
		Platovian	*Viviparus pseudoachatinoides*	
		Mikhailovian	*Crassiana crassoides*	0.8
EOPLEISTOCENE	**Apsheronian**	Morozovian	*Potomida litoralis*	
		Kosnitsian	*Pseudosturia candata*	
		Nesmeyanian	*Bogatschevia scutum*	
		Boshernitsian	*Bogatschevia sturi*	
		Domashkinian	*Unio apsheronicus*	1.6
UPPER PLIOCENE	**Akchagylian**	Polivadinian	*Bogatschevia tamanensis*	
		Kryzhanovian	*Unio kujalinicensis*	
		Akkulaevian	*Erbersininaia*	2.4
		Simbuchinian	*Potomida bashkirica*	
		Poratian	*Rugunio lenticularis*	3.3
MIDDLE PLIOCENE	**Kimmerian**	Kimmerian	*Unio sturdzae*	

Source: Adapted from Tshepalyga (1972).

be replaced once more in the beginning of the late Akchagylian by cold-adapted, non-sculptured forms. At the end of the late Akchagylian, the Levantine species of Unionidae and Viviparidae, particularly *Potomida tamanensis*, reappeared once more, but with less emphasized sculpture.

The beginning of the Apsheronian (late Villafranchian) was characterized by the complete absence of Levantine elements (Domashkinian fauna), reflecting the sharp cooling associated with the beginning of the Quaternary. The subsequent outburst of speciation and sculptured forms (Boshernitskan fauna with *Bogatschevia sturi*) in the early Apsheronian is evidence for a new amelioration of the climate. The cooling trend of the middle Apsheronian (Kosnitsa fauna, with *Pseudosturia caudata*) was succeeded by a short-term warming, which in turn was followed by cooling and the development of the Morozovian fauna, during which the last Levantine elements became extinct (Tshepalyga, 1972). The succeeding warm period at the end of the Apsheronian saw the development of the Kolkotovian assemblage, most species of which are living today.

Geological history

The climatic fluctuations discussed earlier are clearly reflected in the superficial deposits widely distributed in the southern European parts of the former USSR. The oscillations of climatic conditions between dry-and-cold and wet-and-warm during the late Pliocene and early Quaternary are marked by alternating loamy, frequently loess-like deposits and fossil-soil horizons (Nikiforova et al., 1976).

According to students of molluscan biostratigraphy (Alizade et al., 1972; Tshepalyga, 1972), significant changes in molluscan faunas and in geotectonic regimes occurred in both the Black Sea and the Caspian Sea regions at closely correlative levels of the late Pliocene and early Quaternary:

1. Akchagylian-Kuyalnikan (predominance of the Pliocene fauna)
2. Apsheronian-Gurian (mostly Pliocene fauna; appearance of earliest Quaternary mollusk lineages)

3. Bakuan-Chaudan (predominance of Quaternary mollusks; rare Pliocene relicts)

Alizade et al. (1972) considered that the last stage should be equivalent to the base of the Quaternary. It can be seen, however, that the Quaternary fauna began to appear in the second stage, which coincides with the customary (if not stratotypical) Calabrian, or Selinuntian, of Italy, defined by the Vrica level at which the lower boundary of the Quaternary is drawn.

The character of the Apsheronian fauna suggests that the salinity of the Apsheronian basin was similar to that of the present-day Caspian, although some parts of the basin may have had lower salinity because of major rivers discharging into the sea. The flora of the Apsheronian is indistinguishable from the modern flora (Bogachev, 1936). The climate of that time also seems to have differed only slightly from today's climate.

Intense volcanic activity was characteristic of the Apsheronian, as indicated by abundant interlayers of volcanic ash brought far to the east of the eruption centers by winds and sea currents. In the beginning of the Apsheronian, there were also significant movements of the earth's crust. The Apsheronian sea had lower salinity than the Akchagylian, despite the reduction in its surface area as a result of the tectonic movements; the depth of the basin was also increased (Andrusov, 1923). As an example, in the Apsheron peninsula, relatively shallow-water Akchagylian deposits are succeeded by deeper-water Apsheronian sediments, demonstrating subsidence of the seafloor at that time. Some subsidence was also characteristic of the Manysh area. Maximum tectonism was recorded at the beginning of the Apsheronian, in both the Caspian and the Black Sea regions.

Thus, the transition from the Akchagylian to Apsheronian was accompanied by major tectonism that caused the Caspian Basin to become smaller and deeper. Mass appearances of *Limnaea* and, more frequently, *Dreissena* mollusks seem to have resulted from freshening of the water at the same time. In the middle Apsheronian, the Caspian-type fauna penetrated the Euxinian region through the Manysh strait, and a similar but impoverished fauna inhabited the Black Sea basin in the late Apsheronian (Koleshnikov, 1940).

Renewed tectonism at the end of the Apsheronian changed the configuration of the Caspian Basin, downwarping the northern part. That opened a new connection between the Caspian and Euxinian basins through the Manysh strait and resulted in immigration of characteristically Bakuan Cardiidae from the *Didacna crassa* group.

The Gurian beds in the Black Sea basin are the equivalents of the Apsheronian of the Caspian region. The Gurian flora of Georgia differs sharply from the preceding Pliocene flora, in that evergreens were replaced by deciduous species that are still common in Transcaucasia. The Gurian basin appears to have been located mainly inside the present Black Sea basin; for that reason, Gurian beds are relatively scarce, occurring mainly in western Transcaucasia and in the northwestern part of the Kerch (Chersonese) Peninsula.

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22 The N/Q boundary in Asian Russia and Tadjikistan

MIKHAIL N. ALEKSEEV

Introduction

On the Siberian Platform (i.e., the region extending from the Urals eastward to the Pacific Ocean), biostratigraphic, radiometric, and paleomagnetic scales have been applied to Pliocene and Pleistocene deposits to correlate them not only with the stratotype sequence of Vrica but also with key sequences of the European part of the former USSR and in Europe.

Stratigraphic summary

Nine areas are particularly helpful in studying the Plio–Pleistocene succession of this region. These are western Siberia, Transbaikalia, Olkhon Island in Lake Baikal, the upper basin of the Lena River, central Yakutia, the Kolyma Basin, Kamchatka, and the Primorie/Priamurie region (i.e., the Pacific coast provinces), all in eastern Russia, and the Tadjik Depression of Tadjikistan (Figures 22.1 and 22.2). There are difficulties, however, including the unequal extent of exploration, inadequate exposures, and the vast area of territory in Siberia affected by permafrost. As a result, few sequences are adequately known.

Western Siberia

The Plio–Pleistocene deposits of southwestern Siberia, around the headwaters of the Ob River, can be considered of paramount importance, as they include faunistic complexes that support extensive biostratigraphic correlations. The Kochkov Formation, which covers the Upper Pliocene and Eopleistocene, from about 2.4 Ma to 0.6 Ma, contains a series of sediments with four faunal complexes (Figure 22.2). The layers of Betekian age at the base contain remains of *Hipparion* sp., *Paracamelus gigas, Paracamelus praebactrianus, Trogontherium minor, Promimomys gracilis, Mimomys polonicus, Mimomys hintoni, Villanyia petenyii, Villanyia steklovi*, and so on (Zazhigin, 1980).

The Betekian, according to the paleobotanical and geological data of Volkova and Baranova (1980), was characterized by a fairly warm climate and forest vegetation that spread far beyond northwestern Siberia. In the final stages of the Betekian (2.4 Ma) the evidence indicates considerable cooling; southwestern Siberia developed taiga vegetation, and the north developed forest-tundra and tundra associations.

The overlying Podpusk-Lebyaginsk layers show wide adaptive radiations of *Equus* species, the dispersal of *Archidiskodon* (or *Mammuthus* according to some specialists) (Aguirre et al., Chapter 9, this volume) and *Eucladoceros*, reductions and eventual disappearances of thermophilous molluscan faunas in this region, and the initial appearances of recent species. Major changes in the entire environment are registered at the transition from the Podpusk-Lebyaginsk to the overlying Kizikhinsk layers, that is, at the boundary of the Pliocene and Eopleistocene (between 2.0 Ma and 1.7 Ma). That time was marked by intense deterioration of the climate, as indicated by the appearance of tundra and forest-tundra assemblages on the drainage divides and fir-tree forests in the valleys of southwestern Siberia. Rootless-toothed voles such as *Allophaiomys, Prolagurus*, and *Eolagurus* appeared in the rodent fauna for the first time. Thus, the boundary between the Podpusk-Lebyaginsk and Kizikhinsk layers reflects the usual small-mammal biostratigraphic changes at the boundary between the Neogene and Quaternary (Aguirre et al., Chapter 9, this volume; Chaline, Chapter 14, this volume).

Evidence of some warming is found in the upper part (late Eopleistocene) of the Kochkov Formation, and that was again followed by renewed cooling, with the maximum climatic deterioration seen in the Shaitan horizon, in an interval coinciding approximately with the Mindel glacial period in Europe.

Tadjik Depression

In Tadjikistan, strata relating to the Upper Pliocene and Lower Pleistocene in the Tadjik Depression and in the eastern Pamirs provide evidence of three major geological boundaries, dated to about 3.6, 2.2, and 0.7–0.8 Ma by paleomagnetic analysis. The level corresponding to the Olduvai paleomagnetic interval (1.95–1.77 Ma) is located in the lower part of the middle unit, the Kayrubak suite (Dodonov, 1980); however, a more distinct biostratigraphic boundary is somewhat earlier, at about 2.2 Ma,

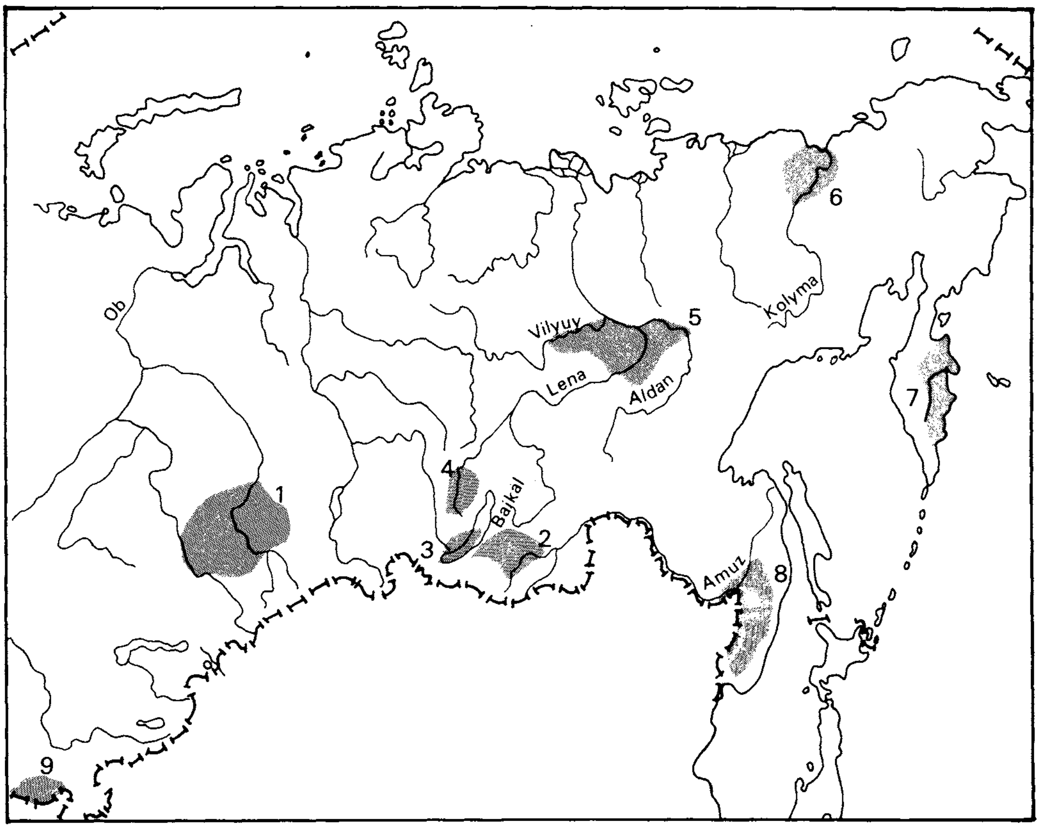

Figure 22.1. Key areas of Neogene and Quaternary deposits in the Siberian Platform region: 1, Western Siberia; 2, Transbaikalia; 3, Olkhon Island in Lake Baikal; 4, upper Lena River drainage; 5, central Yakutia; 6, Kolyma Basin; 7, Kamchatka; 8, Priamurie/ Primorie; 9, Tadjik depression.

at the transition between the Kayrubak and the underlying Kuruksay suite.

Transbaikalia

In southeastern Siberia and Transbaikalia, Vangengeim (1977) and Zazhigin (1980) distinguished the Tchikoian faunal complex, which correlates with the Betekian and the lower parts of the Podpusk-Lebyaginsk complexes of western Siberia. Abundant remains of large mammals and rodents have been found in the key section at Beregovaya in a layer of reddish, clayey sands. The fauna includes *Hipparion* sp. ex gr. *H. houfenense* Teilhard & Young, *Hipparion tchikoicum* Ivanijev, *Dicerorhinus* sp., *Gazella* cf. *G. sinensis* Teilhard & Piveteau, *Antilospira* sp., *Paleotragus* sp., *Canis* cf. *C. chihlienensis* var. *minor* Teilhard & Piveteau, *Nyctereutes* cf. *N. sinensis* (Schlosser), *Euryboas* cf. *E. lunensis* Camp, *Felis (Lynx) shansius* Teilhard, *Acinonyx* sp., *Mimomys* cf. *M. reidi* Hinton, *Mimomys minor* Fejfar, *Villanyia* sp., *Sinocastor* sp., and *Prosiphneus* sp. ex gr. *P. praetingi* (Vangengeim, 1977). That complex can be dated to the beginning of the Middle Villafranchian (Lower Eopleistocene and beginning of the Middle Eopleistocene). The red formation in the lower part of the Tologoi section, which also belongs to the Tchikoian horizon, is normally magnetized and belongs to the Gauss paleomagnetic epoch, according to Gnibidenko, Erbayeva, and Popelova (1976). The boundary between the Pliocene and Eopleistocene in Transbaikalia is drawn at the base of the deposits with a Dodogolian mammalian fauna, characterized by *Citellus* sp., *Villanyia laguriformis*, another *Villanyia* sp., and a *Prosiphneus* sp., collected from the red clays and loams exposed in the lower part of the Dodogol sequence.

The stratigraphically overlying sediments of slope-proluvial origin contain fossil mammals of the Itantsa complex. They include *Equus* ex gr. *E. sanmeniensis* Teilhard & Piveteau, *Villanyia* sp., *Mimomys* sp., *Prosiphneus* sp. ex gr. *P. youngi*, and others. The Itantsa complex is correlated with the Tamanian mammalian complex of eastern Europe (Vangengeim, 1977), where Tamanian-age deposits are considered to be the lowermost part of the Quaternary sequence.

Lake Baikal

Four mammalian and five molluscan complexes have been identified in the Pliocene–Lower Pleistocene deposits of Olkhon Island in Lake Baikal (Logachev, 1982). Environmentally induced changes in the composition of mammalian and molluscan faunas were closely parallel and clearly reflect the environmental and magnetic changes of the proposed boundary between the Neogene and the Quaternary.

At the base, the contact between the colored sediments of the Olkhon Island suite and the underlying clays of the Kharanskian suite has been dated to 3.0–3.5 Ma, according to paleomagnetic measurements. The Sasinsk deposits at the base of the Olkhon Island suite contain the Saraian mammalian faunal complex,

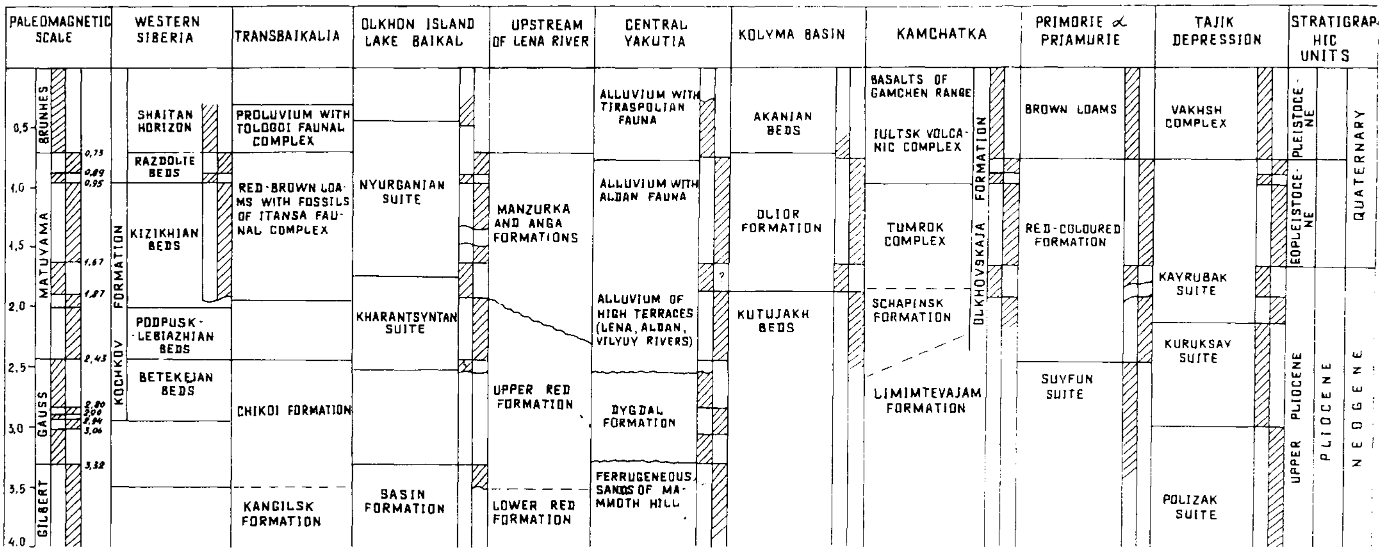

Figure 22.2. Correlation of Pliocene and Lower Pleistocene key sequences in Asian Russia and Tadjikistan.

with elements close to the Lower Pliocene fauna of the Pavlodar suite and the "Pontian" of China. At the same time, the presence of *Microtocoptes* and *Microtodon* in the Saraian fauna makes it similar to the Novostanichian fauna, dated to the Middle Pliocene in western Siberia. Thus, the age of the Saraian fauna in the Baikal area is estimated to be Lower–Middle Pliocene. The Saraian climate was warm and dry, probably steppes with spots of bushes and forest vegetation.

The paleomagnetic level corresponding to the beginning of the Olduvai event is found above that horizon, in the upper part of the Kharanskian suite. It is associated with the first appearances of the late rodent species of the *Villanyia–Mimomys* assemblage. Resting on eroded Kharanskian strata, beds containing the *Eolagurus–Lagurodon* fauna, with single species of *Allophaiomys* and *Mimomys*, characterize the Njurganian suite, which embraces the Eopleistocene and Pleistocene. The 0.78-Ma level, at the transition from the Matuyama to the Brunhes, is recorded only by polarity changes, without any marked faunal change.

Lena River basin

In central Yakutia, the sequences in the Lower Aldan depression, the valleys of the middle Lena River, and the lower course of the Vilyuy River provide significant data for regional correlation.

In the upper part of the Mammoth Hill section on the Aldan River, an alluvial layer termed the "ferruginous sands" is assigned to the Pliocene, based on palynology in overlying and underlying deposits (Baranova et al., 1976). Paleomagnetic studies (Minjuk, 1982) indicate that the ferruginous sands were deposited in the Gilbert epoch. In the palynoflora, the main Angiospermae pollen are *Betula* and *Alnus*, while the Gymnospermae consist mainly of *Pinus* and *Picea*. That layer is widely exposed in the Quaternary terrace deposits of the Aldan River, and a layer with similar palynologic and paleomagnetic character can be traced in the subsurface levels of the Lower Aldan depression.

Sands of the Dygdal suite that have the normal polarity of the Gauss epoch occur in the downwarped central part of the Lower Aldan depression.

In the Lena River valley, upstream of Yakutsk, normally magnetized alluvial deposits occur in elevated terraces, between 100 and 120 m above base level. Eopleistocene deposits, with remains of *Alces latifrons, Bison* aff. *B. schoetensacki, Equus* sp. ex gr. *E. sanmeniensis*, and *Trogontherium* cf. *T. cuvieri*, were found resting on the bedrock of the Aldan River terraces. The reversed polarity of those beds corresponds to the Matuyama epoch. Although the section is of insignificant thickness, the boundary between the Neogene and Quaternary, as related to the Olduvai event, must be drawn within that interval. More precise identification of the boundary position is a matter for the future.

Sequences in the Aldan and Vilyuy rivers also contain the boundary between the Matuyama and Brunhes epochs. In the Vilyuy basin there are alluvial deposits in the 50–60-m terrace that contain a fauna close to that of the Tiraspolian gravel, and in which the basal Brunhes paleomagnetic reversal has been identified. The same relationships between paleomagnetic events and the distribution of Tiraspolian fossils have been established in the Kolkotova Balka sequence in Moldavia.

Kolyma Basin

In northeastern Russia, in the Beringian province east of the Yakut basin, 50–60% of the early Pliocene floras show features that connect them with the seeds, leaves, and pollen of living deciduous flora (Volkova and Baranova, 1980). In the interval between 3.5 Ma and 3.0 Ma, when there was a decrease in land area and the Bering land bridge was inundated, the region's vegetation was transformed into something similar to forest tundra.

In the upper layers of the Kutujakh (or Kututyak) Formation, E. I. Virina (in Volkova and Baranova, 1980) identified an interval of normal polarity in the zone of reversed magnetization of the Matuyama epoch that is considered to represent the Olduvai event. The lower Kutujakh Formation yielded rare remains of a fossil mammal fauna similar to that of the Khaprovian Podpusk-Lebyaginsk assemblage, together with the earliest distinct signs of cryoturbated soil of the Hypoarctic zone with perannual permafrost. The pre-Quaternary intensification of Arctic basin glaciation, at about 2.5 Ma, probably can be referred to this period.

In the Kolyma lowland, the type section of the Olior (Olyor) suite documents the upper Matuyama, including the Jaramillo subchron, and the lowest part of the Brunhes (Sher, 1987). The boundary between Eopleistocene and "true" or glacial Pleistocene could be drawn, in the type Olior, between the layers with an early Olior small-mammal fauna, correlated to the Tamanian complex, and the layers with a late Olior small-mammal fauna, correlated to the Tiraspolian complex.

Kamchatka

Within the Kamchatka-Chukotka region, age calibrations of magnetostratigraphic reversals indicate that the Gilbert–Gauss, Gauss–Matuyama, and Matuyama–Brunhes chron boundaries can be recognized, as well as the Olduvai subchron. The earliest reversal found in the sequences on the Pacific coast of Asia, and the least pronounced, occurs in the volcano-sedimentary sequences of the Kamchatka region in the upper part of the Schapinsk suite (Pevzner, 1972), which according to paleomagnetic evidence dates within the Gilbert and Gauss epochs. According to Shantser (in Alekseev et al., 1979), a volcanogenic layer within the suite has a K/Ar age of approximately 3.8 Ma.

The Gauss–Matuyama boundary and a normally magnetized zone corresponding to the Olduvai event are recognized in the marine sequence of Kamchatka. In the lower portion of the Olkhovian sequence, the boundary between Neogene and Quaternary at the top of the Olduvai subchronozone is very well characterized in terms of marine micropaleontology, with clear changes in the subarctic North Pacific diatom floras and foraminiferal faunas (Gladenkov, 1994; Gladenkov et al., in press). In the continental sequence, the Olduvai paleomagnetic zone is recorded at the boundary between the Schapinsk suite and the Tumrosk volcanogenic complex.

The stratigraphic level corresponding to the transition from the reversed polarity of the Matuyama epoch to the normal polarity of the Brunhes epoch is recorded in the lower part of the shield and stratovolcanic lava sequence overlying the Tumrosk volcanogenic complex. It may also be recorded in the sequences of alluvial and lacustrine deposits of the upper Olkhovian suite of Kamchatka. In the Karagian marine deposits, the Matuyama–Brunhes boundary lies in the lower part of the arctic–boreal molluscan complex.

Priamurie/Primorie

In the Pacific margin or far-eastern region of Russia, levels with the age of the Olduvai and Matuyama–Brunhes paleomagnetic reversals are clearly recognized. The oldest Upper Cenozoic stratigraphic level in the region is found in the Evoron-Chukchagyr depression, where the Pliocene is recognized by a reduction of the Turgai paleofloral elements in the pollen complex and an increase of boreal elements. In the Primorie province, that level is established at the top of the lower subsuite of the Suifun suite. The upper, normally magnetized part of the Suifun suite is referred to the Gauss paleomagnetic epoch.

In the Priamurie and Primorie territories, the magnetostratigraphic horizon corresponding to the Olduvai event is in the base or lower part of the reversely magnetized, red-colored layer that overlies the Suifun suite. The red layer corresponds in general to the Matuyama paleomagnetic chron, and in the area of Spassk-Dalny the first considerable cooling is recorded approximately at the level of the Olduvai event. A more pronounced cooling is recorded close to the transition between reversely magnetized red-colored strata and overlying normally magnetized brown-colored beds, marking the transition from the Matuyama to the Brunhes epoch.

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23 The Pliocene–Pleistocene boundary in the Indian subcontinent

M. V. A. SASTRY

Introduction

In 1973, the Geological Survey of India proposed a project on the Tertiary–Quaternary boundary to cover the global perspective of the boundary between the Pliocene and Pleistocene. This proposal was merged with the "Neogene/Quaternary Boundary" project initiated at about the same time by the Geological Institute of the USSR Academy of Sciences, thus giving rise to IGCP Project 41 (Nikiforova and Alekseev, Chapter 1, this volume).

The Indian national working group for IGCP-41 was formed in 1974 with M. V. A. Sastry as convenor and (from March 1981) with A. Ranga Rao as co-convenor. Members were as follows: B. S. Towari, Panjab University; V. V. Sastri, Oil and Natural Gas Commission, Dehra Dun; D. Niyogi, Indian Institute of Technology, Kharagpur; S. N. Rajaguru, Deccan College, Pune; M. S. Srinivasan, Banaras Hindu University; K. N. Prasad, K. K. Verma, and A. K. Dutta, Geological Survey of India. The working group functioned under the guidance of the director general of the Geological Survey of India, who was also the chairman of the Indian National Committee for the IGCP. A major contribution was the 1979 field conference on the Neogene–Quaternary boundary as it related to the Siwalik and the Karewa deposits of India, which stimulated much new work (Sastry et al., 1981). The present report contains the results of progress since that field conference.

Neogene and Quaternary deposits are widely exposed in the foothills of the Himalaya, in the Vale of Kashmir, in the coastal regions of the Indian peninsula, and in the Andaman Islands. While the Himalayan foothills, Kashmir, and peninsular India have continental deposits, the Andaman Islands preserve a good sequence of deep-water marine facies. Isolated shallow-water marine outcrops of latest Cenozoic age are also found along the coasts of Kutch-Saurastra, Kerala, and Coromandel (Figure 23.1).

Andaman-Nicobar Islands

An almost continuous sequence of Neogene and Quaternary deep-water marine sediments is exposed in the Andaman group of islands, in the Bay of Bengal some 1,300 km away from the mainland of India. Good sequences can be found in many of the islands, but the one on the west coast of Neil Island is relatively accessible and contains a rich and well-preserved foraminiferal fauna that has been taken as a reference section for the Neogene–Quaternary boundary (Srinivasan, 1981).

The late Pliocene and early Pleistocene are identified as the Taipian and Shompenian stages, respectively, based on occurrences of planktonic foraminifera. The Taipian consists of a 15-m-thick section of dark gray, highly calcareous, silty mudstone. A rich assemblage of planktonic foraminifera in the *Globorotalia tosaensis tenuitheca* zone, equivalent to zone N.21 of Banner and Blow (1965), has been recovered from these beds.

Pleistocene strata of the Shompenian Stage overlie the Taipian, with a possible disconformity, and consist of about 45 m of moderately hard, buff-colored, lithic grainstone made up of fairly well-sorted fragments of foraminifera, corals, and algae, mixed with angular quartz grains. The microfaunal assemblage is comparatively poor, but is distinguished from the underlying Taipian by the presence of the initial phylogenetic form of *Globorotalia truncatulinoides*. This horizon is taken to delineate the Pliocene–Pleistocene boundary in the Andaman region. Figure 23.2 shows foraminiferal ranges of the sequence.

No serious attempts had been made to study the nannofossils, diatoms, and radiolaria from these horizons when this chapter was prepared, and data on palaeomagnetism and radiometric ages were not available.

The Karewa of Kashmir

North of the Siwaliks, and separated from them by the Pir Panjal Range, the Kashmir Valley contains a thick pile of sediments of lacustrine, glacial, and fluvial origin capped by loessic strata. Those sediments are known as the "Karewa" and are divisible into lower and upper formations.

The Lower Karewa consists of faulted and folded beds composed of clay, shale, sands, boulder conglomerates, and lignite beds, with plant and vertebrate fossils throughout. The Upper Karewa, on the other hand, has thick horizontal beds of

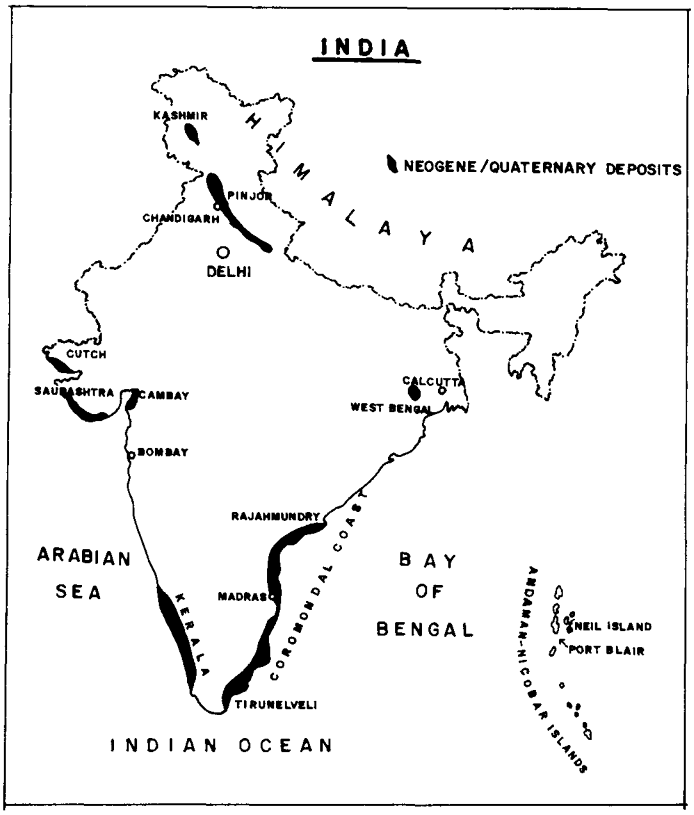

Figure 23.1. Neogene and Quaternary deposits in India. The Pliocene–Pleistocene boundary sequences discussed in the text are (a) Neil Island, in the Andaman-Nicobar archipelago, with deep-water marine sediments, (b) Kashmir, with lacustrine, glacial, and fluvial sediments, and (c) Pinjor (fresh-water molasse of the Siwaliks).

gravel, calcareous clay with marl bands, and loamy clay, with very few fossils.

Lydekker (1883) equated the Karewa with the Upper Siwalik (discussed later) and assigned it a Pliocene age. De Terra and Paterson (1939) recognized four major glacial advances in the Karewa. They correlated the occurrence of *Elephas hysudricus* in the Lower Karewa with the Pinjor of the Siwaliks and included it in the Pleistocene. Wadia (1951) argued that the basal beds of the Lower Karewa might extend down to "Pontian" (i.e., Upper Miocene) levels. On the basis of a diatom study, Roy (1975) thought that the Lower Karewa could be Mio–Pliocene, and Bhatt and Chatterji (1981) preferred to assign a Plio–Pleistocene age to the entire Karewa deposits.

The Karewa is divided into two parts, the lower termed the Hirpur Formation and the upper termed the Nagum Formation. *Elephas hysudricus* and *Equus sivalensis* occur within the Hirpur Formation, at Sombur and Shopian (Bhatt and Chatterji, 1981). Those authors recognized three zones in the Hirpur Formation and placed the Neogene–Quaternary boundary immediately above the conglomerate in their zone 2, based on the evidence of vertebrate fossils and on the premise that the Neogene–Quaternary boundary should be isochronous with the Tatrot–Pinjor boundary in the Siwaliks. On the other hand, Roy (1975) preferred to place the Pliocene–Pleistocene boundary at a higher level, between Lower Karewa (Hirpur) and Upper Karewa (Nagum) formations, based on evidence from studies of tectonism and fresh-water diatoms.

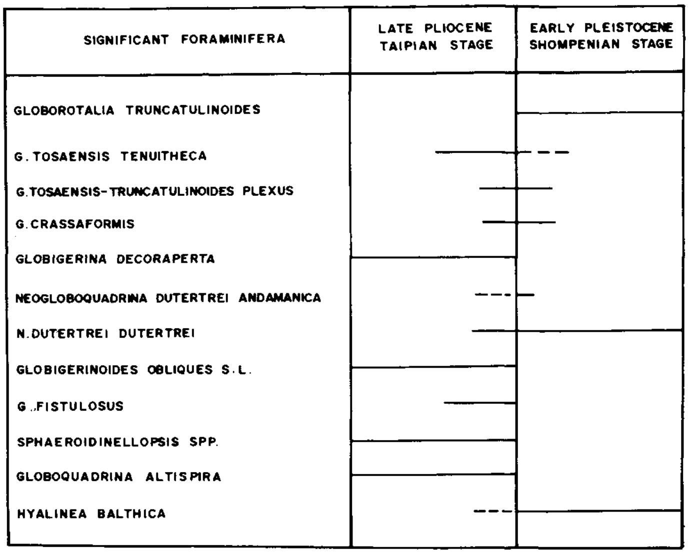

Figure 23.2. Ranges of some significant Plio–Pleistocene foraminifera in the Andaman-Nicobar Islands. (Adapted from Srinivasan, 1981.)

Paleomagnetic studies were carried out on samples from the Hirpur and Nagum formations by Agrawal et al. (1981). Three distinctive conglomerate horizons within the Hirpur Formation served as markers to control paleomagnetic sampling of intervening muds, sands, and silts (Figure 23.3). The paleomagnetic data indicate that the Nagum strata, above the Hirpur, are in the Brunhes normal-polarity chron. Reversed polarities of the Matuyama chron were detected in a 10-m-thick deposit below Conglomerate III, and normal polarities indicative of the Gauss chron were found in sand and mud deposits below and above Conglomerate II. The Gilbert reversed-polarity chron is evident above Conglomerate I. According to those authors, the Pliocene–Pleistocene boundary, taken at the top of the Olduvai subchron, can be recognized between Conglomerates III and II in the Hirpur Formation (Figure 23.3).

From the foregoing it is evident that the Pliocene–Pleistocene boundary lies within the Hirpur Formation significantly above the level where Bhatt and Chatterji (1981) put it (i.e., equivalent to Conglomerate II), and well below the Hirpur–Nagum boundary where Roy (1975) estimated it. However, the paleomagnetic sampling was widely spaced, and the studies are still preliminary, so the exact position of the Neogene–Quaternary boundary in the Karewa cannot be stated with precision.

The Siwaliks

In the foothills of the Himalaya in northwestern India and on the Potwar Plateau in Pakistan there are thick, fresh-water molasse

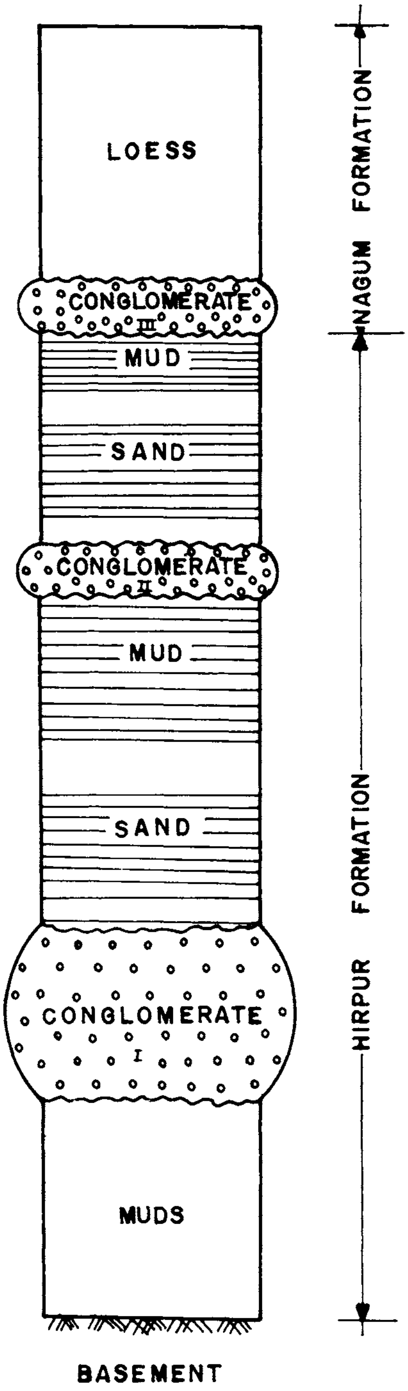

Figure 23.3. Schematic lithologic section of the Karewa. The Pliocene–Pleistocene boundary is surmised to lie between the Hirpur and Nagum formations (i.e., below the Conglomerate III horizon). (Adapted from Agrawal et al., 1981.)

deposits up to 5,000 m thick, known as the Siwaliks. The Siwalks are well known for their continuity and their abundant and diversified mammalian faunas, spanning a period from the Middle Miocene to the Lower Pleistocene. In terms of lithology, the beds show great variation laterally and become progressively coarser from the lower to the upper parts of the sequence. The following is the broad classification of the Siwalik Group (Pilgrim, 1913; Acharyya, Dutta, and Sastry, 1979):

Upper Siwalik sub-group	Boulder Conglomerate Formation Pinjor Formation Tatrot Formation
Middle Siwalik sub-group	Dhok Pathan Formation Nagri Formation
Lower Siwalik sub-group	Chinji Formation Kamlial Formation

The Potwar Plateau has well-developed Siwalik deposits, and the type sections of Kamlial, Chinji, Nagri, Dhok Pathan, and Tatrot are described from that area, whereas the Upper Siwalik beds are better developed in India. At the Pinjor type section, an excellent sequence of Tatrot, Pinjor, and Boulder Conglomerate formations is exposed. Good sections are available east of Chandigarh, Masol (Tatrot–Pinjor), and Nadah (Pinjor–Lower Boulder Conglomerate) (Figure 23.4).

Early studies

The continuity of the record of fossils and environment in the Siwaliks led earlier workers to undertake detailed studies in mammalian biostratigraphy. In addition to interpretations of stratigraphy and climatic changes based on fossils, research on the Pliocene–Pleistocene boundary was also part of these studies.

Pilgrim's 1913 classification of the Siwaliks became the basis for later workers. He compared the faunas with those of Europe, taking the Pontian assemblage to be transitional between Miocene and Pliocene. Based on the occurrence of *Elephas planifrons*, he assigned a Middle Pliocene age to the Pinjor and considered both the Upper Pinjor and the overlying Boulder Conglomerate zones to be uppermost Pliocene. Earlier, Lydekker (1883), on the evidence of *Equus namadicus* and *Elephas* cf. *E. namadicus*, had suggested that the Boulder Conglomerate zone was Lower Pleistocene.

Matthew (1929) advocated a younger age, assigning Lower Pleistocene to the Upper Siwalik and Middle Pliocene to the Dhok Pathan. He considered that the mammal fauna recorded a time break equivalent to the Upper Pliocene between the Dhok Pathan and the Tatrot. This was supported by De Terra and Paterson (1939) on the evidence of conglomerate at the base of the Tatrot, indicating a period of erosion. The apparent coincidence of those two lines of evidences gave strong support to the idea that the base of the Upper Siwaliks should be equivalent to the base of the Pleistocene. De Terra and Teilhard de Chardin (1936), Lewis (1937), Colbert (1935, 1942), Paterson (1941), and Movius (1944) emphasized the occurrence in the Tatrot of *Archidiskodon* (i.e., *Mammuthus*, according to Aguirre et al., Chapter 9, this volume) and advocated from this that the Tatrot should be assigned a place within the Pleistocene. That was in line with Haug's proposition (1911) that the Pleistocene

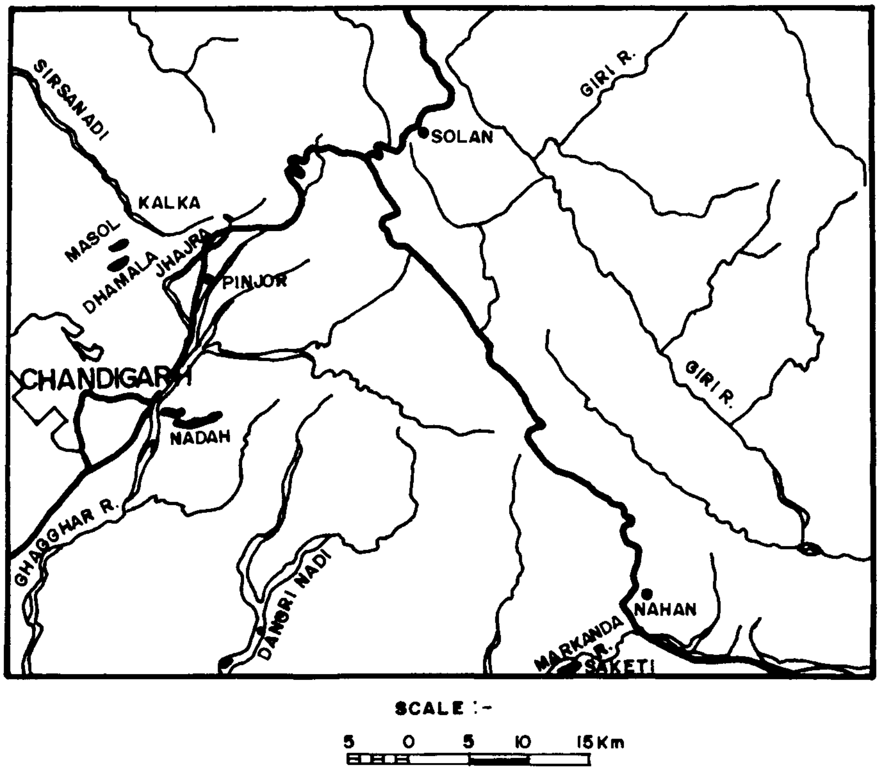

Figure 23.4. Upper Siwalik area. Pliocene and Pleistocene strata are best exposed between Pinjor and Nadah, east of Chandigarh.

was characterized by the first appearances of *Elephas* (as primitive forms that were later redesignated as *Archidiskodon*, and now *Mammuthus*), *Leptobos*, and *Equus* simultaneously on several continents. Hooijer and Colbert (1951) carried out a statistical analysis of the fauna on the premise that the Tatrot and Pinjor faunas could be correlated to a Villafranchian age between the Pliocene Dhok Pathan and Pleistocene Pinjor.

Pilgrim (1944) maintained that the Tatrot and Pinjor rocks were equivalent to Astian and Villafranchian, respectively, of Europe. He included the Pinjor beds in the Pliocene because of the fact that *Elephas* and *Bos* had definite origins in or near India, and their appearance in India should therefore have taken place in the Pliocene, much earlier than in Europe. Because of the presence of *Hipparion* in the Pinjor beds, the close relationship between Pinjor and Villafranchian faunas, and the fact that the glacial formation of the Bain Boulder Bed overlies the Pinjor, Pilgrim concluded that the Pliocene–Pleistocene boundary should also lie above the Pinjor.

Wadia (1951) favored Pilgrim's view that glaciation had begun earlier in Europe than in Asia and that basing the advent of the Pleistocene on the first glaciation was not important. However, he advocated a transition from the pre-glacial Pinjor to the glacial Boulder Conglomerate zone. Gill (1951) suggested the existence of a regional unconformity in northwestern India representing a Siwalik phase of orogeny that commenced in post-Pinjor times. He suggested that glaciation began earlier than the Siwalik orogeny and the uplift of the Pir Panjal Range.

Later studies

Sahni and Khan (1968) recorded *Leptobos* from the Tatrot and indicated that the first appearances of that genus and *Archidiskodon* in India marked the end of the Pliocene, because those genera were indigenous to India. Further, the immigrations of *Equus*, *Bubalus*, *Hipselephas*, and *Rhinoceros* indicated, according to them, the beginning of the Pleistocene in India. They equated the Tatrot with the Upper Pliocene (Astian) and the Pinjor with the Lower Pleistocene (Villafranchian). Prasad (1974) showed that of 26 vertebrate genera present in the Tatrot, 8 were holdovers from the Dhok Pathan, 9 were newcomers, and the remaining 9 persisted into the Pinjor. From the faunal data, he argued that the Tatrot might be the end of the Pliocene.

Balasundaram and Sastry (1972) contended that the Pinjor was of Pliocene age, based on tectonic, climatic, and paleontological evidence. Sastry and Dutta (1977a,b) observed that the upper part of the Pinjor is conspicuously marked by the simultaneous development of rhythmic alternations in the ratios of pebbles to sand and silt and the absence of fossils. That horizon, according to them, represents drastic changes in climate, life, and environment, related to the initiation of a final phase of Himalayan orogeny, and is the appropriate location for the Neogene–Quaternary boundary.

A systematic regional mapping of the Siwaliks, coupled with lithological and heavy-mineral analysis by the ONGC (Oil and

Natural Gas Commission), has shown that a boundary drawn on the basis of any one criterion does not precisely coincide with boundaries drawn on other criteria. Using the first appearance of *Equus* in the Pinjor as an index to the Pleistocene, Ranga Rao et al. (1981) drew the Pliocene–Pleistocene boundary between the Tatrot and the Pinjor.

As part of IGCP Project 41, an international field conference on the Neogene–Quaternary boundary in the Siwaliks was organized in 1979 (as noted earlier), in which leading workers on the subject took part and examined important sections. It was not possible, however, to arrive at definitive results regarding the placement of the boundary, and the conference group concluded that further paleomagnetic and radiometric studies were needed.

Based on paleomagnetic studies, M. N. Alekseev (personal communication) suggested that the Tatrot is likely to correspond to the Gauss normal-polarity chron, although reversed remanent magnetization in its upper part correlates to the early part of the Matuyama chron. The uppermost part of the Pinjor has been found to be normally magnetized, indicating the Olduvai subchron (Sastry and Dutta, 1977a,b). Yokoyama's (1981) paleomagnetic results (following vertebrate fossil analysis) agree that the Olduvai event is reflected by a normally polarized zone within the Pinjor.

Azzaroli and Napoleone (1982) measured a well-developed 300-m section at Nadah exposing the Pinjor and Lower Boulder Conglomerate formations. The base of the Pinjor is not seen in the Nadah section, but is presumed to be 4 m below the base of the exposed section, beneath river terraces. The sedimentation in this section suddenly became conglomeratic during the Jaramillo normal-polarity event, indicating renewed erosional activity in the Himalayan belt, coeval with significant environmental changes in Europe related to the Mindel glacial maximum. The bulk of the Siwalik fauna disappeared between the Olduvai and Jaramillo events, but those authors concluded that the top of the Olduvai event was closely equivalent to the contact of the Pinjor and the Boulder Conglomerate.

Thus, differing opinions regarding the basic criteria and the emphasis to be placed on certain groups of animals in deciphering the Neogene–Quaternary boundary have resulted in at least three different placements: (1) at the base of the Tatrot (Matthew, 1929; Colbert, 1935; Lewis, 1937; De Terra and Paterson, 1939; Hooijer and Colbert, 1951); (2) at the base of the Pinjor (Sahni and Khan, 1968; Prasad, 1974; Badam, 1979; and many others); and (3) at the top of the Pinjor (Pilgrim, 1944; Gill, 1951; Balasundaram and Sastry, 1972; Sastry and Dutta, 1977a; Azzaroli and Napoleone, 1982). It should be noted that the advocates of placements (1) and (2), as well as early advocates of (3), were employing criteria other than the recommendation of the 1948 London IGC committee, which was to refer the boundary to a physical reference point at the base of the Calabrian Stage in recognition of a change to colder climate conditions in the Italian marine sequences, at a level that subsequently was determined to be close to the top of the Olduvai event.

Recent studies in Pakistan

On the Potwar Plateau, a multidisciplinary study, including biostratigraphy, paleomagnetism, and radiometric dating, has been conducted by several teams of workers. Opdyke et al. (1979) investigated eight separate stratigraphic sections. Two prominent bentonitized tuffs were also recorded in some sections, and radiometric dates obtained by fission-track analysis on zircons from those tuffs have enabled correlations to the standard GPTS (geomagnetic polarity time scale) of the Pliocene and Pleistocene (Cande and Kent, 1995).

The important findings include the observation that the boundary between the Tatrot and the Pinjor, defined faunally as the first simultaneous occurrences of *Equus–Elephas–Bos* (*sensu* Haug, 1911) and cervids with antlers, should be dated to about 2.5 Ma, and thus is very near the Gauss–Matuyama paleomagnetic boundary (2.6 Ma). Other findings were that the Pinjor faunal zone extends to the base of the Brunhes chron, contra Azzaroli and Napoleone (1982), and that the Pliocene–Pleistocene boundary (top of the Olduvai event) is therefore located within the Pinjor faunal zone (although possibly not within the lithologic limits of the Pinjor Formation at the type Tatrot section). The fossil mammal faunas do not show any change, except for the extinction of the giraffid *Sivatherium*, at the level of that boundary in the Potwar area (Figure 23.5).

More recently, Barry, Lindsay, and Jacobs (1982), in their biostratigraphic zonation of the Siwaliks on the Potwar Plateau, proposed an *Elephas planifrons* interval-zone and recorded within it several characteristic mammalian taxa such as *Elephas planifrons*, *Stegodon* sp., *Equus sivalensis*, *Hexaprotodon sivalensis*, *Proamphibos lachrymas*, and cervids. The magnetic-polarity correlations indicate that the top of the zone is very close to the Pliocene–Pleistocene boundary level. Bentonitized tuffs in the Campbellpore area have been dated to 1.61 Ma (Johnson et al., 1982), while the paleomagnetic transition from normal to reversed polarity below the tuff is recognized to be the reversal that marks the top of the Olduvai subchron, whose inferred age is about 1.8 Ma (Cande and Kent, 1995; Pasini and Colalongo, Chapter 2, this volume).

In the trans-Indus region, the Marwat Formation is rich in the fossils of typical Pinjor assemblages and has a well-developed lithostratigraphic unit, the "Bain Boulder Bed," whose glacial origin is debated (West, 1981). The boulder bed, however, is equated with the Olduvai event and also marks an important point in the tectonofluvial history of the region.

Analysis

The Tatrot–Pinjor boundary is based essentially on faunal transitions, and the Pliocene–Pleistocene boundary has been estimated at that level by several workers. Investigations in Pakistan by Opdyke et al. (1979), Azzaroli and Napoleone (1982), Barry et al. (1982), and Johnson et al. (1982), however, have shown that the Tatrot–Pinjor boundary is correlatable with the Gauss–Matuyama paleomagnetic boundary at about 2.6 Ma.

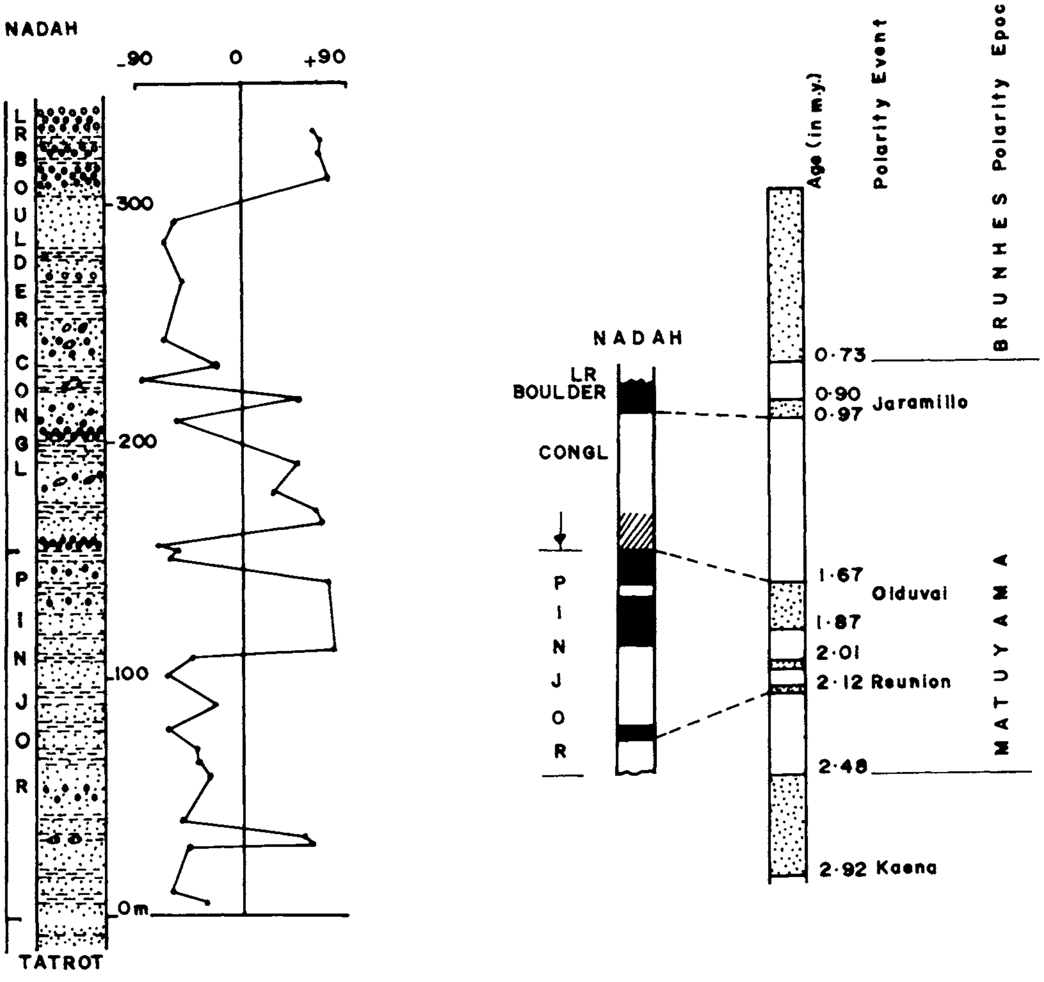

Figure 23.5. Lithology and paleomagnetic polarity in the Nadah section. (Adapted from Azzaroli and Napoleone, 1982.)

If we accept that the top of the Olduvai event is the indicator for the Pliocene–Pleistocene boundary, the Dhok Pathan–Tatrot and the Tatrot–Pinjor concepts become irrelevant.

The change to the Pinjor fauna from that of the underlying Tatrot is not significant either in variety or in abundance, but on the other hand the rich Pinjor fauna suddenly disappears at the level of the Boulder Conglomerate. The ratio of pebbles to sand and silt in the sediments and the rates of deposition continued without much change from the Tatrot to the Pinjor, but, again, not into the Boulder Conglomerate. Thus the continuity of the fauna and the similarity of depositional and climatic conditions support a close relationship between the Tatrot and the Pinjor.

At the end of the Pinjor, the change in sedimentation suggests a sudden increase in erosion due to a combination of tectonism in the adjoining mountains, and also possibly some climate change, at about 1.0 Ma (Jaramillo), if not 0.7 Ma or later (lower Brunhes). The high-energy sedimentary environment was much less favorable for the preservation of fossils, and in fact fossils are rare if not absent in the Lower Boulder Conglomerate, which may account for much of the apparent faunal change at that level. The marked changes at the Pinjor–Boulder Conglomerate contact are unrelated to the top of the Olduvai subchronozone and should not be used to mark the Pliocene–Pleistocene boundary.

In the Indian region, nevertheless, further systematic collecting, coupled with detailed paleomagnetic and possibly radiometric investigations, will be necessary before a final conclusion can be reached on the location of the Pliocene–Pleistocene boundary in the continental Siwalik deposits. It is not possible at this stage to establish a local reference or parastratotype to mark the Neogene–Quaternary boundary in those deposits.

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24 The Pliocene–Pleistocene boundary in Japan: the Osaka Group, Kinki district

MINORU ITIHARA, SHUSAKU YOSHIKAWA, and TADAO KAMEI

Introduction

The INQUA Subcommission 1-d on the Pliocene–Pleistocene boundary organized the working group on the Neogene–Quaternary boundary as the action body for Project 41 of the International Geological Correlation Program (IGCP) in 1974. As a research and contact body corresponding to the INQUA subcommission, the Japanese National Committee for Quaternary Research in the Science Council of Japan organized the Japanese National Subcommission on the Pliocene–Pleistocene Boundary, chaired by M. Itihara. The Japanese national subcommission has functioned simultaneously as the Japanese national working group for IGCP-41 for investigations of the boundary problem in Japan.

In August 1982, the Japanese national working group for IGCP-41 presented its third report at the XI INQUA Congress in Moscow (Itihara and Kuwano, 1982). That report, however, covered too many areas (Figure 24.1) to be included in this final volume. Therefore, for this purpose, it was decided to select two representative studies: this chapter, on the Pliocene–Pleistocene boundary in the Osaka Group, as a typical section of lacustrine and fluvial sediments with marine intercalations, and Chapter 25, on the Plio–Pleistocene stratigraphy in the Boso Peninsula, as a typical section of marine sediments.

Background and stratigraphic framework

Deposits of Pliocene and Pleistocene age are well exposed in Japan (Figure 24.1). In the Osaka Group in the Kinki district of central southern Honshu, strata that correlate well to the interval containing the Pliocene–Pleistocene boundary have been studied or discussed by Itihara (1961), Ishida et al. (1969), Itihara and Kamei (1970, 1982), Itihara et al. (1973, 1975, 1984), and Maenaka et al. (1977). This chapter describes the present status of the problem of the Pliocene–Pleistocene boundary in this well-described sequence.

The Plio–Pleistocene Osaka Group (Osaka Group Research Group, 1951) is composed mainly of unindurated sand, gravel, and clay of lacustrine and fluvial origin, intercalated with a number of tuff layers and 12 marine clays that are the top units of sedimentary cycles. The Osaka Group is well exposed in hilly areas around Osaka, Kyoto, and the Nara and Harima basins, in the central part of the Kinki district. The sequence, 200–1,500 m or more thick, unconformably overlies Miocene and pre-Tertiary basement rocks and is discordantly overlain by terrace deposits. The Osaka Group has been folded and faulted, with maximum fault throws estimated to be 200 m to more than 500 m. The sum of the displacements that have taken place in these basins since Pliocene time is called the Rokko movement (Ikebe, 1956).

The Osaka Group is divisible by means of key tuff layers, 3–4 m in maximum thickness, and the marine clay beds, 10 m or so in maximum thickness (Itihara, 1961; Itihara et al., 1975) (Figure 24.2). Among these key horizons with regional significance, the most informative are as follows: the Fukuda, Yellow, Pink, Azuki, and Sakura tuff layers, and the 12 marine clays identified as Ma-1, Ma0, and Ma1 to Ma10, in ascending stratigraphic order (Figure 24.2).

Biostratigraphy

Plant and animal fossils are abundant in the Osaka Group. Among them, the most important are plant megafossils from many horizons and proboscidean fossils from a rather small number of horizons.

Plant megafossils

The taxonomy of plant megafossils was studied in detail by Miki (1941a,b, 1948). In a detailed review of the stratigraphy of the plant-bearing strata, Itihara (1961) concluded that the lowermost part of the Osaka Group was deposited during a warm-climate period characterized by dominance of the *Metasequoia* flora. Subsequent climatic deterioration led to a decline of the *Metasequoia* flora, which eventually became extinct by the time of a cold-climate horizon just below the Azuki tuff layer, at about 0.9 Ma, associated with the Jaramillo subchron (Figure 24.2). Although that was followed by climatic amelioration, Itihara (1961) suggested that "in the Osaka Group, the beginning of the age of extinction of the *Metasequoia* flora, i.e. the

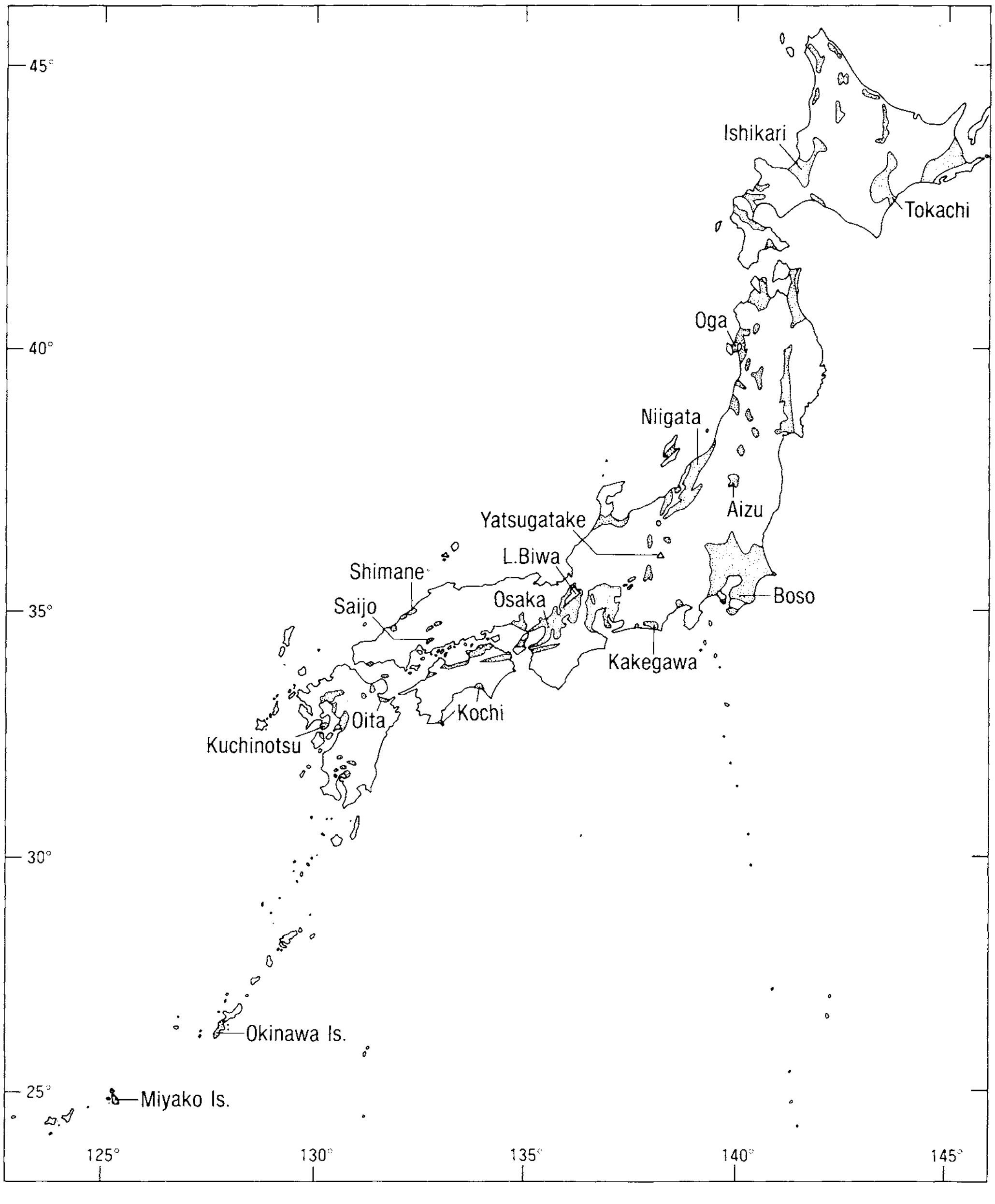

Figure 24.1. Distribution of Plio–Pleistocene sediments in Japan.

beginning of climatic deterioration, should be inferred as the Pliocene–Pleistocene boundary."

The *Metasequoia* flora in the Osaka Group is characterized by an assemblage of *Metasequoia, Glyptostrobus, Sequoia, Pinus koribai, Pinus fujii, Juglans cinerea* var. *megacinerea, Liquidambar, Ginkgo, Pseudolarix, Keteleeria, Nyssa,* and other elements. As shown in Figure 24.1, the *Metasequoia* flora still had the essential characters of the Tertiary, even though *Nyssa, Ginkgo, Pseudolarix,* and *Liquidambar* disappeared during its later stages (Itihara et al., 1975). It can be inferred that climate fluctuations were relatively feeble during the time of deposition of the lowermost part of the Osaka Group.

The beginning of significant climatic deterioration is evidenced in the Osaka Group not only by the extinction of the *Metasequoia* flora in the lower part of the Osaka Group but also by the successive appearances of *Menyanthes trifoliata* and the subalpine species *Pinus koraiensis* and *Picea maximowiczii* in the interval between a horizon below the Kamifukuda tuff layer and a horizon below the Ma1 bed (Kokawa, 1959; Itihara, 1961; Ibaragi Research Group, 1966; Yoshikawa, 1973; Itihara et al., 1975; Momohara, 1990). Thus, the first indication of that remarkable floral change, the extinction of the *Metasequoia* flora, is seen in an interval between the Fukuda and Kamifukuda tuff layers, and the last relics of the *Metasequoia* flora, such as *Metasequoia* and *Picea koribai,* disappear just below the Azuki tuff layer (i.e., just below the Ma3 marine bed). Those events are shown in Figure 24.1 as being just above the Olduvai and Jaramillo subchrons, respectively. After the extinction of the

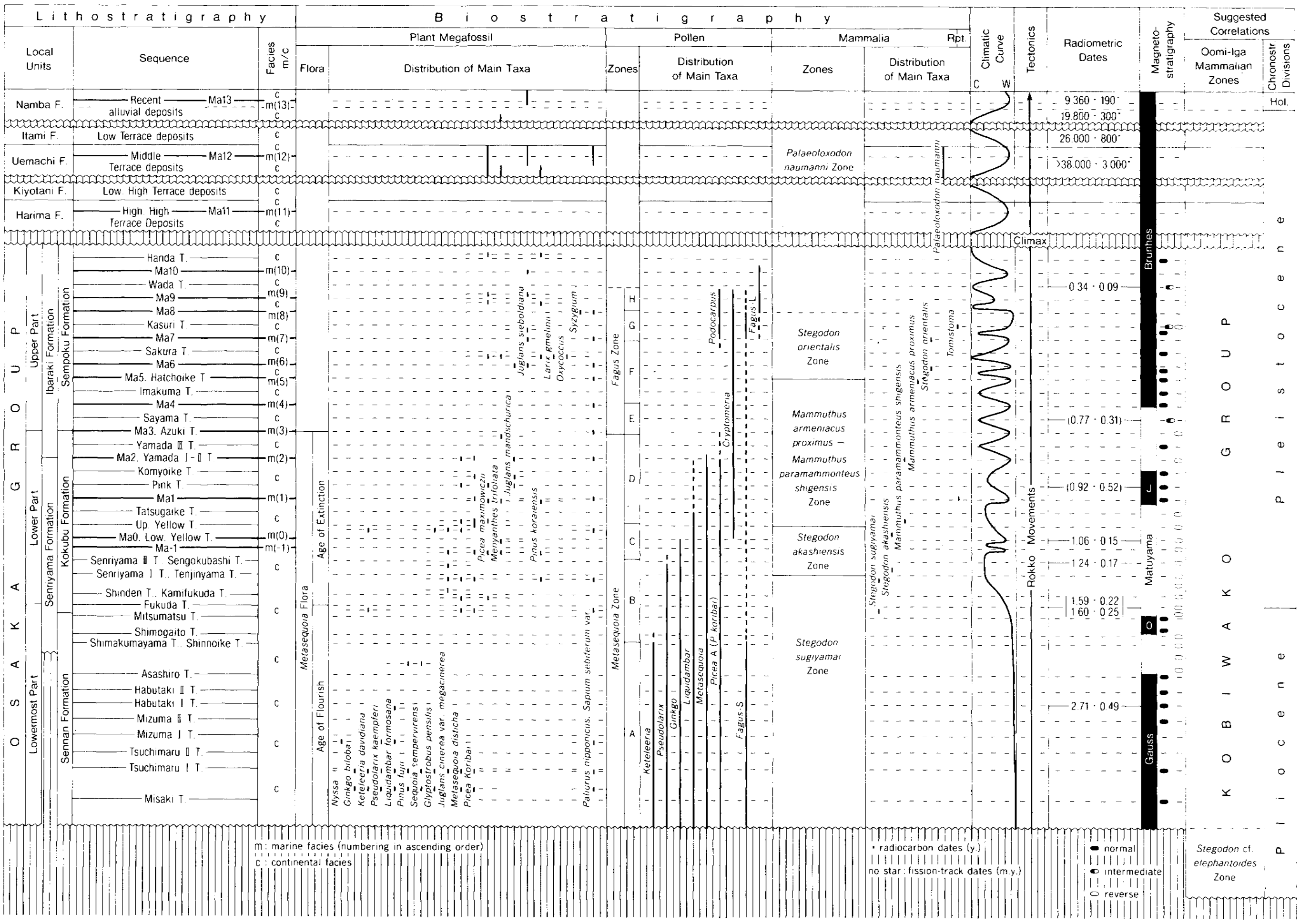

Figure 24.2. Plio–Pleistocene chronology of the Osaka group.

Metasequoia flora, a new kind of flora appeared. For instance, *Juglans mandschurica* replaced *Juglans cinerea* var. *megacinerea*, and then was replaced in turn by *Juglans sieboldiana* (Nirei, 1975).

Climate fluctuations, indicated by alternations of cold-climate megafloras and warm-climate megafloras, intensified after the disappearance of the *Metasequoia* flora (i.e., at the time of deposition of the upper part of the Osaka Group), implying the beginning of glacial conditions. This is exemplified by occurrences of *Larix gmelini, Pinus koraiensis*, and *Oxycoccus palustris* in the interval between the Ma6 and Ma7 beds (cold), followed by *Syzygium buxifolium* and *Podocarpus nagi* in the Ma8 bed (warm), then *Pinus koraiensis* and *Picea maximowiczii* from the interval between the Ma8 and Ma9 beds (cold), and finally *Pinus koraiensis* from the horizon above the Ma10 bed (cold). The warm-climate species *Paliurus nipponicus* and *Sapium sebiferum* are associated with the successive marine beds Ma3 through Ma8, supporting the conclusion that high sea levels documented by the marine clay beds in the Osaka Group were synchronous with warm-climate conditions.

Pollen

Tai (1973) subdivided the Osaka Group into the *Metasequoia* and *Fagus* zones, with the boundary at the base of the Ma3 bed. According to Tai, the onset of climatic deterioration can be detected within the B subzone of the *Metasequoia* zone, especially at the top part of that subzone. It is noticeable that the top part of the B subzone nearly corresponds to the transition between the period of flourishing and the period of reduction and eventual extinction of *Metasequoia* flora, as identified by Itihara (1961). The floral succession seen in the palynological analyses closely parallels that of plant megafossil studies.

Mammalia

The Osaka Group has long been noted for its fossil mammals (Ikebe, Chiji, and Ishida, 1966; Kamei and Setoguchi, 1970; Kamei and Otsuka, 1981). Kamei (1984) found that the Osaka Group and its correlative, the Kobiwako Group, could be subdivided into the following mammalian zones, in ascending order:

8. *Sus scrofa* and *Cervus (Sika) nippon* zone (Holocene)
7. *Mammuthus primigenius* zone (latest Pleistocene)
6. *Palaeoloxodon naumanni* zone
5. *Stegodon orientalis* zone
4. *Mammuthus paramammonteus shigensis–Mammuthus armenaicus proximus* zone
3. *Stegodon akashiensis* zone
2. *Stegodon sugiyamai* zone
1. *Stegodon* cf. *S. elephantoides* zone (middle Pliocene)

All of these zones were founded on mammals from the Osaka Group except for the earliest (zone 1) and latest (zones 7 and 8). It should be noted that at present, the taxonomy of *Stegodon* cf. *S. elephantoides* has been revised to *Stegodon shinshuensis*, a species closely related to *Stegodon zdanskyi* of northern China. At the same time, the two species of stegodont, *S. akashiensis* and *S. sugiyamai*, which characterize the succeeding zones in the lowermost Osaka Group, are both recognized to be conspecific with *Stegodon aurorae* (Kamei, 1991). Finally, Dubrovo (1981) has maintained her opinion that both *Mammuthus paramammonteus shigensis* and *M. armenaicus proximus* are synonymous with *Palaeoloxodon naumanni*.

According to Kamei, zone 1 is characterized by the presence of elements found in the Gaozhuang fauna of the Yushe Basin in northern China, approximately correlatable with zones MN-14 and MN-15 of the western European scale (Qiu, 1989). The faunas of zones 2 and 3 are more endemic, with remnants of the zone 1 assemblage, but are enhanced by immigrants found also in the Nihewan fauna of northern China, such as *Elaphurus, Axis, Rusa*, and so forth. Zone 4 contains transitional and endemic forms of temperate-forest affinities, including *Apodemus argenteus*. Zone 5 is made up almost equally of extinct and living taxa, with temperate-forest elements predominant, but with a few immigrants from warm-temperate faunas, such as *Stegodon orientalis* and *Rhinoceros sinensis*, which are abundant in the Wanhsien fauna of southern China. The giant crocodilian *"Tomistoma" machikanense* from just below the Ma8 level belongs to the zone 5 assemblage. Another crocodilian is also found below the Ma1 horizon.

It is worth notice that the level of occurrence of the deer *Elaphurus* of the Nihewan fauna in zone 3 corresponds to the final extinction of the *Metasequoia* flora.

Magnetostratigraphy

Paleomagnetic studies of the Osaka Group by Torii, Yoshikawa, and Itihara (1974), Maenaka et al. (1977), Maenaka (1983), and Itihara et al. (1984) are summarized in Figure 24.1. The data suggest that the Osaka Group represents the time from the Gauss chron to the early part of the Brunhes chron. One subzone of normal paleomagnetic polarity, from the Ma1 bed to the Komyoike tuff layer, has been thought to correspond to the Jaramillo subchron. Another subzone of normal polarity, from about 2 m above the Shimogaito tuff layer to about 5 m above the Mitsumatsu tuff layer, has been correlated with the Olduvai subchron.

Radiometric dating

Radiometric age determinations for the tuff layers of the Osaka Group, mainly using the fission-track method (Nishimura and Sasajima, 1970; Matsuda, 1980), have been reexamined by Suzuki (1988), with findings that appear to be substantially different from the previously reported ages. It appears that on the basis of direct radiometric dating, the age span of the Osaka Group probably ranges from about 3.0 Ma to 0.3 Ma and that the age of extinction of the *Metasequoia* flora, the level at which the Pliocene–Pleistocene boundary has been correlated in earlier studies (e.g., Itihara, 1961), was in fact about 0.9 Ma. That age is somewhat younger than the age for the top of the normal-polarity zone identified as the Olduvai subchron, in which the Vrica definition is located. It should be noted that the Vrica horizon appears to correspond fairly closely to the level between the Fukuda and Kamifukuda tuffs at which the *Metasequoia* flora first began to decline.

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25 The Pliocene–Pleistocene boundary in Japan: stratigraphy in the Boso Peninsula, central Japan

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Outline of geology

The Boso Peninsula is situated in the southern part of the Kanto region in central Honshu, between two major tectonic provinces of the Japanese Islands: northeastern Honshu, extending to the north, and southwestern Honshu, extending to the west. Thick marine sediments accumulated in a depositional basin that appeared in the middle Pliocene in the southern part of the Kant region; this Plio–Pleistocene section is well exposed in the Boso Peninsula.

The Tertiary and Pleistocene deposits in the Boso Peninsula consist of the Mineoka, Hota, Miura, Kazusa, and Shimosa groups, in stratigraphic order, together with younger terrace formations and volcanic-ash layers. Each of the groups is unconformable with the others. Planktonic foraminifera indicate that the Miocene–Pliocene transition is in the uppermost part of the Amatsu Formation.

After deposition of the Anno Formation of the uppermost Miura Group, the area was uplifted, and a new basin, in which the Kazusa Group was deposited, appeared to the north of the emerged area. In the central-to-eastern parts of the peninsula, the Kazusa Group comprises the Kurotaki, Katsuura, Namihana, Ohara, Kiwada, Otadai, Umegase, Kokumoto, Kakinokidai, Chonan, Mandano, and Kasamori formations, in ascending order. Those formations consist of sandstone and siltstone, with pyroclastic intercalations, and the classification of the formations is based on the predominant lithology in the interbedded rocks.

The pyroclastic intercalations of the Kazusa Group have been carefully traced in the Boso, Choshi, and Miura peninsulas, where they are used as key beds. Lateral changes in lithology and local stratigraphy have been well established (Mitsunashi and Yazaki, 1958, 1961; Mitsunashi et al., 1959, 1961, 1976a,b, 1979; Yazaki and Mitsunashi, 1962; Mitsunashi and Yazaki, 1968; Ishiwada et al., 1971). The Kazusa Group has yielded many fossils, including mollusks, brachiopods, bryozoans, corals, echinoids, foraminifera, calcareous nannoplankton, and plants. Land-mammal fossils have also been found in the Umegase and younger formations.

The Shimosa Group is distributed in the northern part of the peninsula, where it underlies the Shimosa Plateau. The Shimosa Group is made up of the Jizodo, Yabu, and Narita formations, though many other subdivisions have been proposed for this group. The Jizodo Formation overlies the Kasamori Formation, with parallel unconformity. In the western area, fossil valleys were found at the base of the Jizodo Formation. The Shimosa Group consists largely of loose sands, but includes pebbly sand and clayey silt layers, representing sedimentary cycles. Generally, each of the cyclic deposits consists of basal coarse sands, lower-to-middle cross-laminated sands, and upper sandy-to-clayey silts. Slight sedimentary breaks are recognized at the bottoms of some cyclic layers. Each of the formations of the Shimosa Group includes one to several sedimentary cycles.

The Shimosa Group has yielded abundant molluscan and other fossils. The molluscan assemblages indicate that the environment varied between littoral and neritic under the influence of cool, moderate, and warm currents and brackish water (Oyama, 1952; Hatai, 1958; Ogose, 1961; Aoki and Baba, 1980). Oxygen-isotope paleotemperatures obtained from mass-spectroscopic analysis of mollusk shells from the Shimosa Group vary between 10°C and 20°C (Masuda and Taira, 1974).

Magnetostratigraphy and biostratigraphy

The Neogene and Pleistocene of the Boso Peninsula have been classified into magnetozones according to the stratigraphic sequence of polarity measurements in detrital remanent magnetization (Nakagawa, Niitsuma, and Hayasaka, 1969; Niitsuma, 1976; Nakagawa and Niitsuma, 1977). The magnetozones are designated by letters, in alphabetical order downward from the youngest, with the subzones in each magnetozone identified by a numeral following the letter of the magnetozone (Figure 25.1).

The distributions of planktonic microfossils have been examined in the Neogene and Pleistocene sections of the Boso and Choshi peninsulas by various authors (Takayama, 1967, 1973; Sakai, 1972; Koizumi and Kanaya, 1976; Oda, 1977). The Choshi Peninsula is situated in the northeastern part of the Plio–Pleistocene basin in which the Kazusa and Shimosa groups were deposited, and the section exposed in the Choshi Peninsula is

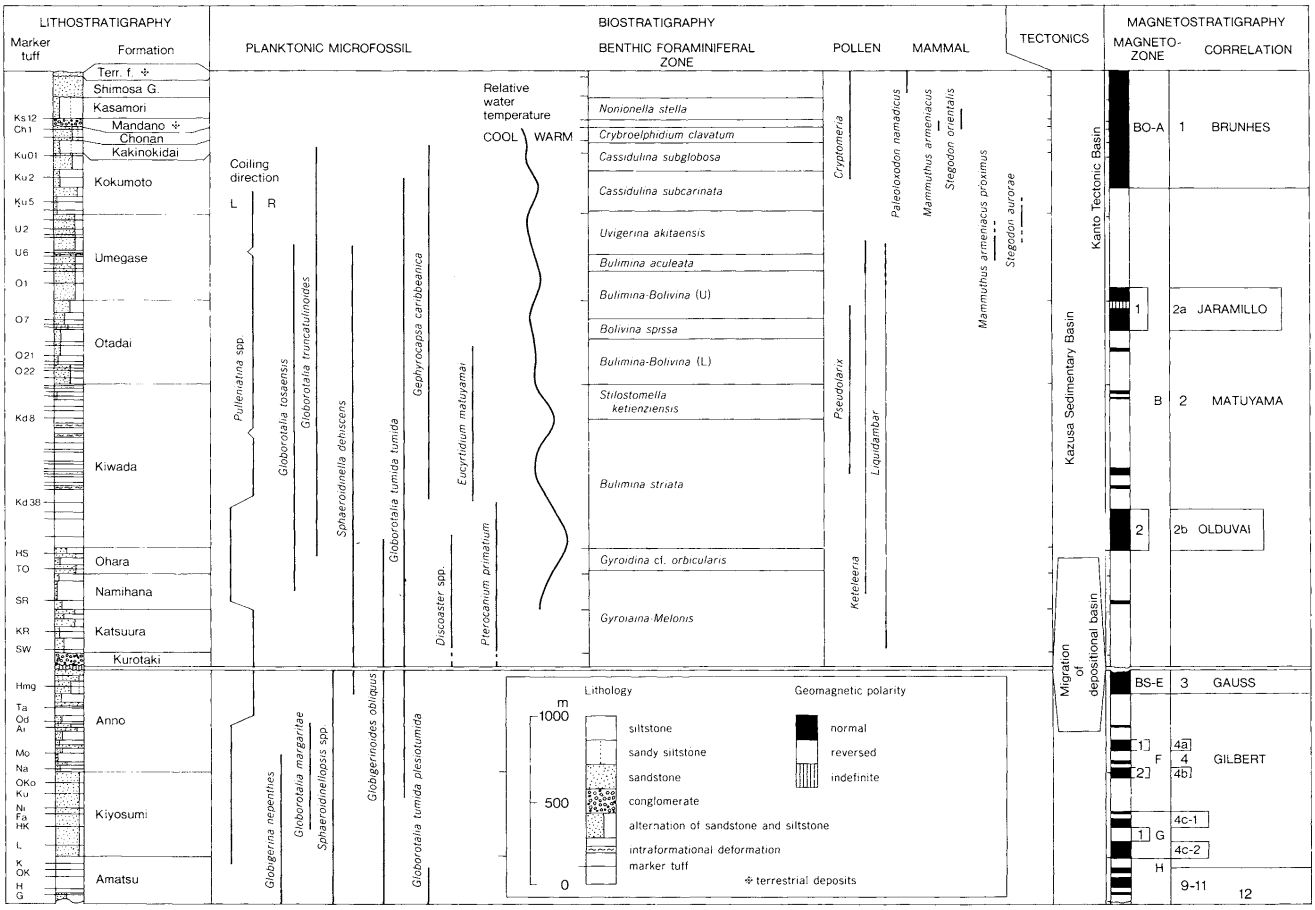

Figure 25.1. Stratigraphy of the Plio–Pleistocene series of the Boso Peninsula.

equivalent to that of the Boso Peninsula. Several pryoclastic key beds of the Kazusa Group have been traced directly from the Boso Peninsula to the Choshi Peninsula, allowing correlation of the stratigraphic section with the magnetozones. The Plio-Pleistocene sediments of the Choshi Peninsula have yielded both calcareous and siliceous microfossils, whereas the sediments of the Boso Peninsula are poor in siliceous microfossils.

Figure 25.1 shows the lithostratigraphy, biostratigraphy, and magnetostratigraphy for the Neogene and Pleistocene of the Boso Peninsula. The stratigraphic distribution of the siliceous microfossils is based mainly on occurrences in the Choshi Peninsula (Sakai, 1972; Koizumi and Kanaya, 1976). The benthic foraminifera zones of Aoki (1968) and the distribution of pollen by Onishi (1969) are included in the same figure.

In the Kazusa Group, the benthic foraminifera assemblage gradually changes upward and also westward to that of a shallow-water environment, indicating that the basin filled from the west. Cold water influenced the middle part of the Kiwada, the upper part of the Umegase, and the lower part of the Kasamori formations. Among the planktonic foraminifera, cold-water species are recognized in the transitional part from the Otadai to Umegase formations and in the upper part of the Umegase Formation.

Thus, the lower part of the Kiwada Formation, which is correlated with the early part of the Olduvai normal-polarity subzone in deeper-water sediments, was deposited in a warm-water environment, but the end of the Olduvai in the middle Kiwada saw cold-water conditions develop.

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26 The base of the Quaternary in China

ZHANG SHOUXIN

Introduction

The Upper Pliocene and Lower Pleistocene sediments in eastern China include various types of deposits, providing an ideal region for study of this part of the geologic record. The cave fillings, loess deposits, and strata of fluviolacustrine, coastal-plain, and marine origin combine to reveal the biological history, topography, climate, and active tectonics during that interval. This chapter presents an overview of the sections that are the most significant for separating the Pliocene and Pleistocene.

Starting in 1949, most Chinese geologists adopted the proposal made at the 1948 XVIII International Geological Congress in London to place the Pliocene–Pleistocene boundary at the first immigration of "cold guests" into the marine faunas of the Mediterranean region, exemplified by the paleontology at the base of the Calabrian Stage of Italy, and to consider that level correlative to the base of the Italian Villafranchian Stage in continental deposits. Because the Plio–Pleistocene vertebrate faunas in China have long been famous and better known (Teilhard de Chardin and Piveteau, 1930) than the marine sequences, the continental Villafranchian concept, rather than the marine Calabrian concept, was widely adopted as the basis for recognition of the base of the Pleistocene in China. It is now recognized that in Italy the earliest Villafranchian is almost twice as old as the base of the Calabrian Stage (Azzaroli et al., Chapter 11, this volume). The Chinese workers have maintained, however, that the Villafranchian definition is in agreement with the first cold-climate faunas of both marine and continental environments in that region, coinciding closely with the Gauss–Matuyama paleomagnetic reversal at about 2.6 Ma.

Nonmarine stratigraphic sections

The Lochuan Loess section

The Lochuan Loess occupies a basin typical of the many large basins in the center of the loess district in northern China. The Lochuan section is located at Heimugou, about 6.5 km south of the capital town of Lochuan county, Shaanxi province. Loess accumulated in that area continuously to a thickness of 136 m over an uneven erosional surface cut into the Hipparion Red Clay Formation. By lithostratigraphic classification, Liu and Chang (1961) divided the Lochuan Loess into three subunits: the Wucheng Loess, the Lishi Loess, and the Malan Loess.

Fossils found in the lower or Wucheng Loess subunit are mainly mammals (Figure 26.1). The species *Prosiphneus intermedius* and *Hipparion (Proboscidipparion) sinense*, found in the unconformably underlying Hipparion Red Clay, do not persist into the Wucheng Loess. At the same time, we note the first appearance of *Myospalax tingi* and *Allocricetus* (Zheng et al., 1985) in the lowest part of the loessic deposits. Magnetostratigraphic study of the Lochuan Loess sections shows that the Wucheng Loess subunit predates the Olduvai subchron of the Matuyama chron (Heller and Liu, 1982). Other studies in the Lochuan area have identified the Gauss–Matuyama boundary just below the base of the Wucheng Loess (Zheng et al., 1985).

Nihewan Series

The Yangyuan and Yuxian basins are intermontane basins in the northern part of Hebei province, where Pliocene and Pleistocene fluviolacustrine sediments are classified as the Nihewan Series. The "Nihewan fauna" described by Teilhard de Chardin and Piveteau (1930) comes from the sands and clays of this unit in the vicinity of Nihewan and Xiaodukou, but recent investigations have discovered additional faunal remains ranging from the base to the top of the 90-m-thick sequence. Faunal and paleomagnetic studies indicate that those strata represent a span of time that overlaps both the customary boundary in China and the revised boundary as recommended by IGCP-41. In modern studies, the lower part of the original Nihewan Formation has been recognized as a separate formation according to lithologic and biostratigraphic evidence, so that the entire sequence is properly termed the Nihewan Series.

At the margin of the basin, near Dongyaozitou, Yuxian, the Nihewan Series rests unconformably on the upper Hipparion Red Clay (there named the Yuxian Formation), in which *Hipparion houfense, Antilospiroides houfense, Sinoryx, Gazella blacki*, and *Viverra* have been found (Wang and He, 1982; Zheng and Cai, 1991). The lowermost Nihewan Series, resting on the

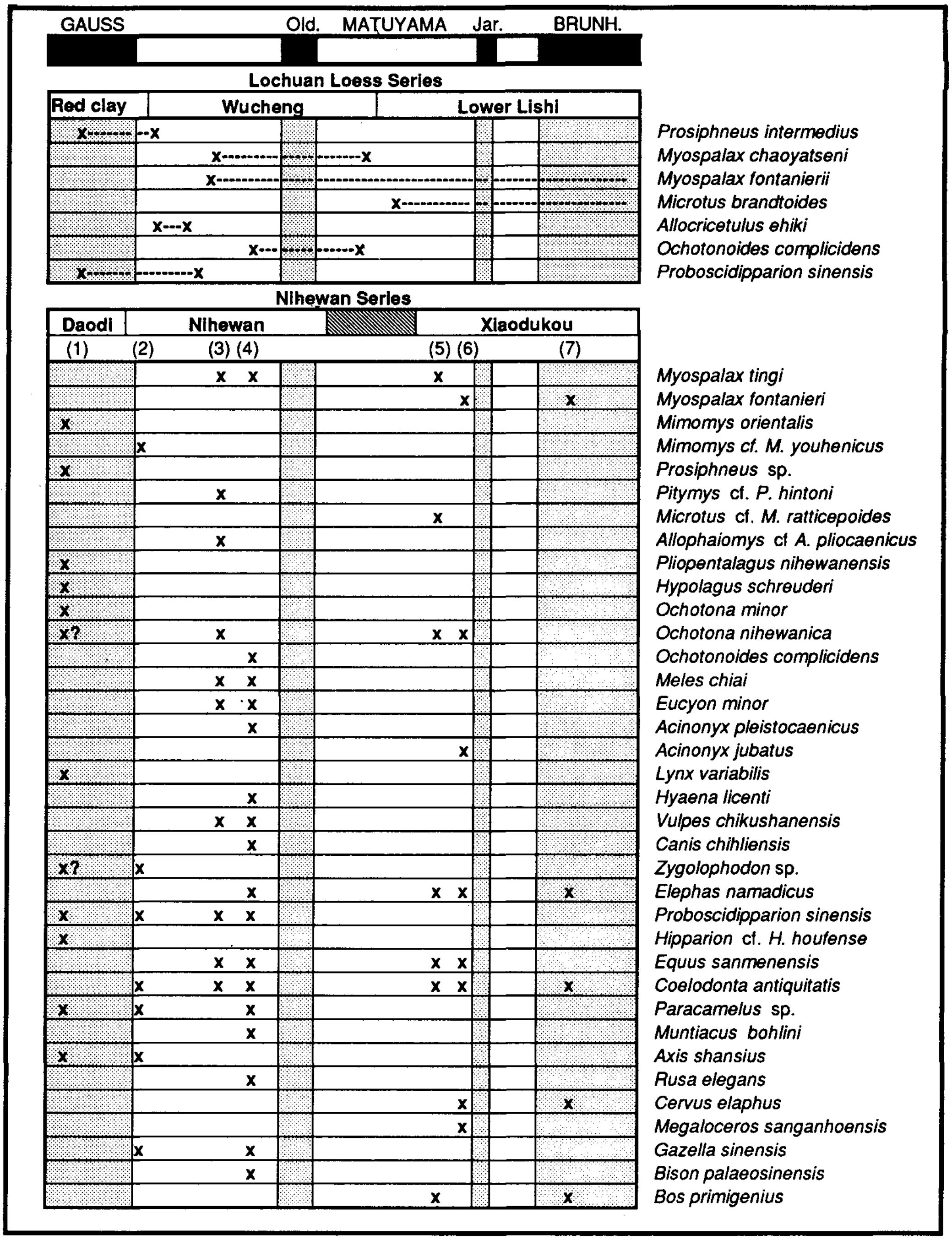


Figure 26.1. Plio–Pleistocene mammal biostratigraphy, on the basis of selected taxa from paleomagnetically analyzed sections in the plateau loess of Lochuan county and fluviatile deposits in the Nihewan region, northern China. Many recorded taxa are omitted from this summary. Data on the Lochuan Loess adapted from Zheng et al. (1985), and data on Nihewan Series adapted from Zheng and Cai (1991). Faunal levels: 1, Daodi; 2, Dongyaozitou; 3, Danangou; 4, Xiashagou; 5, Donggutuo; 6, Xiaodukou; 7, Xujiawa. The Xiashagou l.f. is from reoccupied sites of typical "Nihewan fauna" of Teilhard de Chardin and Piveteau in the Yangyuan Basin (R. H. Tedford, personal communication, 1993). The Donggutuo and Xiaodukou l.f. are from exposures in the interfluve between the Yangyuan and Yuxian basins, and the remaining local faunas are from the Yuxian Basin (Zheng and Cai, 1991).

eroded Yuxian surface, consist of brown and yellow sands and clays, overlain by strikingly cross-bedded gravels. Originally assigned to the Nihewan Formation, those two units were named the Dongyaozitou Formation by Tang and Ji (1983), but more recently the basal brown and yellow sands have been designated the Daodi Formation by Du et al. (1988), and the cross-bedded gravels are recognized as the base of the Nihewan Formation *sensu stricto*.

Several sites in the Daodi Formation have produced a rich and apparently homogeneous fauna (Cai, 1987; Zheng and Cai, 1991) termed the Daodi fauna. The small-mammal assemblage is characterized by *Mimomys orientalis* and *Prosiphneus* (*? = P. paratingi*) and includes representatives of *Sorex, ?Beremendia, Eucastor, Nannocricetus, Germanomys, Orientalomys, Apodemus, Mus, Rattus, Paralactaga, Sminthoides, Hypolagus schreuderi, Pliopentalagus nihewanensis, Ochotona* cf. *O. lagrelii, O. minor*, and *O. erythrotis* (*? = O. nihewanica*). Large mammals include *Lynx variabilis, Hipparion* cf. *H. houfense, Proboscidipparion sinense, Paleotragus progressus, Antilospira yuxianensis, Axis shansius*, and an undetermined elephantid.

The Dongyaozitou local fauna, restricted from the original sense of Tang (1980) according to Zheng and Cai (1991), occurs at the top of the Daodi Formation; it differs from the Daodi fauna, with *Mimomys* cf. *M. youhenicus* in place of *M. orientalis*. Occurrences of *Nyctereutes* cf. *N. sinensis, Zygolophodon, Coelodonta antiquitatis, Paracamelus*, and *Gazella sinensis*, so far unrecorded from the Daodi fauna, are noted at that level. The only known equid is *Proboscidipparion sinense*, with no evidence of either *Hipparion* or *Equus*. A possible record of the latter, cited by Du et al. (1988), is considered invalid (R. H. Tedford, personal communication, 1992). The fossil fish *Pungitius nihowanensis* is also first recorded from this level (Tang and Ji, 1983).

In the cross-bedded gravels at the base of the Nihewan Formation *s.s.* at Dongyaozitou, and in beds just above the gravels, Zheng (1981) and Li (1984) discovered a younger fauna called the Danangou local fauna, in which the first reliably identified remains of *Equus* are found. There, *E. sanmenensis* occurs together with *Proboscidipparion sinense, Ochotona nihewanica, Orientalomys nihowanicus, Canis chihliensis minor, Meles chiai, Coelodonta antiquitatis*, and *Gazella sinensis*. From present indications (Zheng and Cai, 1991) that fauna is essentially the same age as the classical Nihewan fauna of Teilhard de Chardin and Piveteau (1930); R. H. Tedford (personal communication, 1992) notes that 80% of the Danangou species-level taxa (including both equids) are also found in the larger Nihewan fauna, as summarized by Zheng and Cai (1991). Key small-mammal taxa from the Nihewan fauna include *Myospalax tingi, Pitymys* cf. *P. hintoni, Allophaiomys* cf. *A. pliocaenicus*, and *Ochotonoides complicidens*. Among the large mammals, aside from those previously noted, which also occur at Danangou, the presence of *Equus teilhardi, Elephas (Palaeoloxodon) namadicus, Elaphurus bifurcatus, Eucladoceros boulei*, and *Rusa elegans* is of some significance.

A widespread unconformity separates the beds with the typical Nihewan fauna from upper levels, where fossils at Donggutuo (Wei, Meng, and Cheng, 1985) and at Xiaodukou (Wei, 1983) document the initial occurrences of *Microtus raticepoides, Myospalax fontanieri, Acinonyx jubatus, Megaloceros, Bos primigenius*, and *Cervus elaphus*. At Donggutuo and nearby sites, abundant stone tools have also been discovered (Wei, 1985; Wei et al., 1985; Schick et al., 1991), apparently among the earliest evidences of human occupation in eastern Asia (Aguirre, Chapter 10, this volume).

Paleomagnetic analysis of the Nihewan beds indicates that the upper Daodi Formation, at the approximate level of the Dongyaozitou local fauna, contains the Gauss–Matuyama boundary (Du et al., 1988). The lower Nihewan Formation containing the Danangou l.f. and the classic Nihewan faunas shows the reversed paleomagnetic polarity of the lower Matuyama chron (Dong et al., 1986). Above the stratigraphic hiatus in the middle of the Nihewan Formation, the upper Nihewan faunas and Paleolithic artifacts occur in beds just below the Jaramillo subchron. Strata recording the Olduvai subchronozone are not present, because of the mid-Nihewan hiatus.

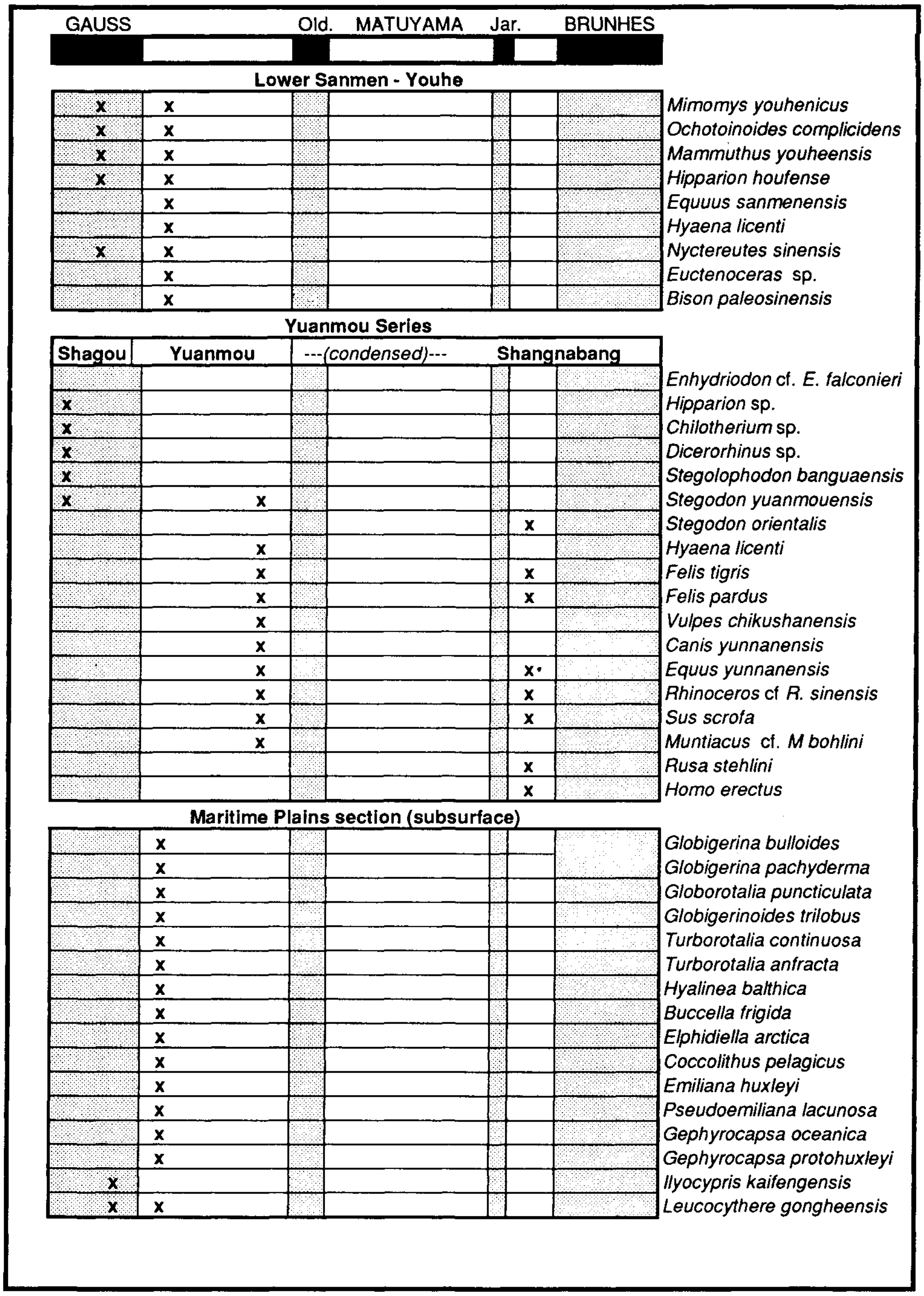
Tang and Ji (1983) believed that the Dongyaozitou fauna was clearly transitional between the fauna of the Hipparion Red Clay, considered as later Pliocene in China, and the classic Nihewan faunal unit, which has been dated to the Pleistocene according to the Chinese usage. The mammals of the Dongyaozitou fauna (of Tang and Ji, 1983) can be correlated with the Youhe faunal unit of the lower Sanmen Formation at Weinan, Shaanxi, and also can be correlated to the later part of the Montopoli unit (Étouaires fauna) of the early Villafranchian, dated to the earliest part of the Matuyama in Europe. The classic Nihewan fauna contains *Myospalax tingi*, which indicates a correlation of the lower Nihewan Formation to the Wucheng Loess in Shaanxi (Zheng et al., 1985); *Allophaiomys* cf. *A. pliocaenicus, Pitymys* cf. *P. hintoni*, and *Elephas (Palaeoloxodon) namadicus* are elements in common between the Nihewan fauna and the pre-Olduvai Villafranchian in central Asia and Europe (e.g., Saint-Vallier). The presence of *Microtus, Bos*, and *Cervus* in the fauna collected at Donggutuo and Xiaodukou is consistent with the late Villafranchian age suggested by the pre-Jaramillo paleomagnetism in those strata.

Sanmen Formation

The fluviolacustrine sediments developed in the valleys of the Huang He (Yellow River) and Wei He have been named the Sanmen Formation, after Sanmenxia, Henan province. The mammalian fossils in the Sanmen Formation are similar to those from the Nihewan Formation. The lower part of the Sanmen Formation in Weinan, Shaanxi, contains a mammalian fauna known as the Youhe faunal unit, with *Mammuthus ("Archidiskodon") youheensis, Hipparion houfense*, Rhinocerotidae, *Cervavitus* sp., *Sus subtriquetra, Nyctereutes sinensis, Mimomys youhenicus, Cricetulus* sp., and *Ochotonoides* cf. *O. complicidens*. The Youhe faunal unit, as noted, compares most closely with the Dongyaozitou faunal unit below the typical Nihewan Formation in the Yuxian Basin.

Yuanmou section

The Yuanmou section is exposed on the southern bank of the Jinsha River, in an intermontane basin in northeastern Yunnan province. In that basin, Pliocene and Pleistocene fluviolacustrine sediments some 600 m in thickness consist mainly of sands and gravels, as well as sandy clays with lignite beds. Beginning in 1938, with discovery of teeth of *Equus yunnanensis*, the Yuanmou beds were taken as the type section for the Lower Pleistocene in southern China, equivalent to the Nihewan Series of northern China. In 1961, Zhou Ming-zhen concluded that the faunal assemblage of the Shagou Lignite, found in the lower part of the Yuanmou section, could be correlated to the Dhok Pathan fauna of the Siwaliks, based on the presence of *Enhydriodon* cf. *E. falconeri*, and that the Shagou level was therefore of "late Pliocene" age. The Yuanmou section was subsequently divided by You et al. (1978) into three formations: a lower (Shagou), a middle (Yuanmou), and an upper (Shangnabang).



GAUSS		Old.	MATUYAMA	Jar.		BRUNHES	
			Lower Sanmen - Youhe				
x	x						*Mimomys youhenicus*
x	x						*Ochotoinoides complicidens*
x	x						*Mammuthus youheensis*
x	x						*Hipparion houfense*
	x						*Equuus sanmenensis*
	x						*Hyaena licenti*
x	x						*Nyctereutes sinensis*
	x						*Euctenoceras* sp.
	x						*Bison paleosinensis*
			Yuanmou Series				
Shagou	**Yuanmou**		---*(condensed)*---		**Shangnabang**		
							Enhydriodon cf. *E. falconieri*
x							*Hipparion* sp.
x							*Chilotherium* sp.
x							*Dicerorhinus* sp.
x							*Stegolophodon banguaensis*
x	x						*Stegodon yuanmouensis*
					x		*Stegodon orientalis*
	x						*Hyaena licenti*
	x				x		*Felis tigris*
	x				x		*Felis pardus*
	x						*Vulpes chikushanensis*
	x						*Canis yunnanensis*
	x				x•		*Equus yunnanensis*
	x				x		*Rhinoceros* cf *R. sinensis*
	x				x		*Sus scrofa*
	x						*Muntiacus* cf. *M bohlini*
					x		*Rusa stehlini*
					x		*Homo erectus*
			Maritime Plains section (subsurface)				
	x						*Globigerina bulloides*
	x						*Globigerina pachyderma*
	x						*Globorotalia puncticulata*
	x						*Globigerinoides trilobus*
	x						*Turborotalia continuosa*
	x						*Turborotalia anfracta*
	x						*Hyalinea balthica*
	x						*Buccella frigida*
	x						*Elphidiella arctica*
	x						*Coccolithus pelagicus*
	x						*Emiliana huxleyi*
	x						*Pseudoemiliana lacunosa*
	x						*Gephyrocapsa oceanica*
	x						*Gephyrocapsa protohuxleyi*
x							*Ilyocypris kaifengensis*
x	x						*Leucocythere gongheensis*

Figure 26.2. Plio–Pleistocene biostratigraphy according to selected taxa from paleomagnetically analyzed sections in Sanmenxia (Shaanxi) and Yuanmou (Yunnan), and from boreholes S-5 and Bo-3 on the northern coastal plain, Shunyi county, near Beijing.

At present, the Shagou Formation has yielded a fossil assemblage that indicates late Pliocene–early Pleistocene (i.e., early Villafranchian) age, with *Hipparion, Dicerorhinus, Chilotherium yunnanensis*, the proboscideans *Serridentinus, Stegolophodon*, and *Stegodon*, and *Enhydriodon* species (Figure 26.2). In the overlying Yuanmou Formation, which You et al. (1978) believed to be restricted to early Pleistocene age, a different assemblage is recorded, with a few holdovers from Shagou time (most notably *Stegodon*), but with many new taxa, such as *Equus yunnanensis, Sus scrofa, Felis (Panthera) tigris, Felis (Panthera) pardus*, and *Rhinoceros* cf. *R. sinensis*, which are unknown from the older levels. The rare mammals from the Shangnabang horizon indicate correlation to the Middle Pleistocene *Stegodon orientalis* fauna of southern China.

Paleomagnetic studies by Li et al. (1976) and Cheng et al. (1977) agreed on placing the boundary between the Gauss and Matuyama chrons at the boundary between the Shagou and Yuanmou formations, which is consistent with the overall Villafranchian character of the Yuanmou vertebrate assemblage. Fossil teeth of *Homo erectus* from the Shangnabang faunal level were initially assigned an age of more than 1.6 Ma, on the basis of paleomagnetic analysis, but Liu and Ding (1983) and Jiang, Sun, and Liang (1988) have argued that the normal-polarity sediments below the beds from which these specimens were collected correspond to the Brunhes, not the Olduvai, and that the Shangnabang fossils are therefore younger than 0.7 Ma. In that view, the stratigraphic interval yielding Yuanmou fauna (i.e., the Yuanmou Formation only) spans the entire Matuyama, from 2.6 to 0.7 Ma, and thus it is a possibility that the Yuanmou fauna is mixed from levels of pre-Olduvai and post-Olduvai age. The stratigraphy, however, is rather unclear, and as Aguirre et al. (Chapter 9, this volume) have pointed out, the faunal and paleomagnetic data are also consistent with correlation of the normal-polarity strata below the Shangnabang fauna to the Jaramillo event at about 1.0 Ma. The Shangnabang *H. erectus* might thus come from uppermost Matuyama strata.

Many Chinese vertebrate paleontologists consider that it is suitable to place the faunal levels of early Villafranchian age in the early Pleistocene. In the Nihewan and equivalent fluviolacustrine sequences in northern China, there are thus two choices for the reference point for the N/Q boundary (the base of the Quaternary), according to the preferred concept of the Pleistocene in that region. One option is to place it at the base of the Nihewan Formation (*sensu stricto*), or Bed 6, at the first-appearance datum of *Equus sanmenensis*, and the other is in Nihewan Bed 4, at the last-appearance datum of *Zygolophodon* and the fish *Pungitius nihowanensis*. In both options, the N/Q boundary is close to the Gauss–Matuyama boundary at 2.6 Ma. However, it is also admitted that the Youhe–Dongyaozitou mammalian faunas could mark the Upper Pliocene, in view of the present decision to define the N/Q boundary at Vrica, at a level equivalent to the top of the Olduvai normal chron and thus well above the lower Nihewan, lower Sanmen, and lower Yuanmou levels.

The maritime-plain sections

Plio–Pleistocene sediments in the North China Plain consist mainly of fluviolacustrine deposits interbedded with marine layers. Drilling in these strata has shown that the Plio–Pleistocene interbeds of marine and nonmarine sediments were deposited in coastal environments.

S-5 borehole section. Important information has come from examination of the main stratigraphic features of samples taken from the S-5 borehole in Shunyi county, near Beijing (Wang and He, 1982; An et al., 1979). Jurassic basement was encountered at 793.7 m. The Matuyama–Brunhes geomagnetic-polarity reversal was placed at a depth of 160 m, and the Gauss–Matuyama reversal at a depth of 468 m. At 428 m, equivalent to an age of about 2.3 Ma, a marine bed 10 m thick is composed of gray-green to blue-gray medium-coarse silty sand, with thin layers of sandy clay. That marine layer yields benthic and planktonic foraminifera such as *Globigerina bulloides, G. pachyderma, Hyalinea baltica, Buccella frigida*, and *Elphidiella arctica*. An assemblage of calcareous nannofossils in the same bed includes *Cyclococcolithus leptoporus, Coccolithus pelagicus, Emiliana huxleyi, Pseudoemiliana* cf. *P. lacunosa, Gephyrocapsa oceanica*, and *G. protohuxleyi*. A similar marine interval has been observed in many borehole sections east of Beijing, and paleomagnetic study has shown those occurrences to be correlative to the marine bed in the S-2 core section.

Chinese paleontologists recognize that the calcareous nannofossils of the Beijing plain can be correlated to the entire range of the NN19 through NN21 nannofossil zones, which encompasses the Quaternary. Therefore, they prefer to place the Pliocene–Pleistocene boundary at the base of the *Hyalinea–Globigerina* assemblages, as seen in the S-2 core, essentially coincident with the Gauss–Matuyama polarity reversal at 468 m.

Bo-3 borehole section. In a 600-m core from the Bo-3 borehole (39°15'N, 118°30'E), in the northern coastal plain, three major paleomagnetic polarity zones were recognized by Li and Wang (1982). The first zone showed normal polarity from the surface to a depth of 171 m, with the exception of a few reversed samples from 15 m and 103 m. From 171 m to 493 m, the polarity of the remanent magnetization in the samples is reversed, and in the normal-polarity zone below 493 m there is a zone of mixed polarities from 572 m to 588 m. From the different lines of evidence, these three polarity zones appear to correspond to the Brunhes normal, Matuyama reversed, and Gauss normal chrons, respectively.

In the lower part of the core, three marine beds occur in an 80-m interval. At a core depth of 505 m, about 15 m below the assumed Gauss–Matuyama boundary, an ostracode assemblage is found that represents a transition from the older *Leucocythere gongheensis* assemblage to the succeeding *Ilyocypris* assemblage. The exact position for the N/Q boundary, equivalent to the boundary in Chinese continental sequences, should thus be at a depth of 515 m, at the first appearance of *Ilyocypris kaifengensis*.

In the same Bo-3 borehole, there is a marked change in pollen flora at a depth of 464 m, from a mixed conifer and broad-leaf forest assemblage with *Ulmus, Pinus, Betula*, and others to a dry steppe assemblage with *Chenopodia ceae* and *Artemisia*. According to the magnetostratigraphy described earlier, the changes in ostracoda and vegetation observed in the Bo-3 core happened near the Gauss–Matuyama boundary. Those data roughly agree with the results obtained from the S-5 core section, in which the layer bearing *Hyalinea baltica* is just above the Gauss–Matuyama reversal, at about 2.3 Ma.

In 1978, Zhang and co-workers studied the 600-m Ca-13 core section. Brunhes, Matuyama, Gauss, and Gilbert polarity zones were recorded. Those authors put the lower boundary of the Pleistocene at the Mammoth subchron of the Gauss normal-polarity chron.

Marine section

Pliocene and Pleistocene sediments in the South China Sea and around Taiwan Island belong to the neritic environment.

Research in Taiwan shows that the first-appearance datum levels for the planktonic foraminifera *Globorotalia truncatulinoides* and the calcareous nannofossil *Gephyrocapsa oceanica*, together with the last-appearance datum of *Discoaster brouweri*, are more or less synchronous. These events may be correlated with the lower boundary of the *G. truncatulinoides* zone (N22 zone) and the lower boundary of the *Pseudoemiliana lacunosa* zone (NN19 zone) of the Mediterranean (Rio et al., Chapter 5, this volume).

Conclusions

The beginning of loess deposition in China, which coincided with major changes in the large- and small-vertebrate faunas equivalent to the Middle Villafranchian, dates to the same level as the Pretiglian cold-climate phase in western Europe and to the earliest loessic deposits in Stranzendorf, Austria, and in Tadjikistan (Sibrava, 1992). Following Teilhard de Chardin and Piveteau (1930), this level was adopted many years ago by Chinese stratigraphers as a logical position for the Pliocene–Pleistocene boundary. The resolution of the 1948 IGC that established the criteria for defining the Pliocene–Pleistocene boundary in Italy considered that the change in continental climates in the middle Villafranchian was coeval with the appearance of cold-climate fossils in the marine deposits of the Calabrian Stage. Although this correlation appears to be incorrect for the Mediterranean Basin, it describes the situation in China rather accurately. Chinese micropaleontologists observed the appearance of the *Hyalinea baltica–Globigerina pachyderma* cold-climate assemblage in northern China to coincide with the Gauss–Matuyama boundary, and therefore to coincide with the mid-Villafranchian change in mammal faunas. That microfossil datum is found above the Olduvai normal-polarity subchron in the upper Matuyama in Mediterranean sections, including sections of Calabrian age (Pasini and Colalongo, Chapter 2, this volume; Azzaroli et al., Chapter 11, this volume). The difference in the times of these two placements indicates that the first appearances of *Hyalinea baltica* and *Globigerina pachyderma* in China (or elsewhere in Asia) were the first evolutionary appearances of those taxa, and their first appearances in the Mediterranean were delayed by regional environmental effects.

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27 Plio–Pleistocene deposits and the Quaternary boundary in sub-Saharan Africa

H. BASIL S. COOKE

Introduction

According to the definition of the lower boundary of the Pleistocene epoch at Vrica (Pasini and Colalongo, Chapter 2, this volume), the top of the Olduvai subchron can be used as a convenient approximation of the Quaternary boundary in nonmarine areas. In Africa, several highly fossiliferous stratigraphic sequences are known in which the limits of the Olduvai subchron can be placed, either by direct paleomagnetic observations or through good radiometric control. The faunal associations in those sequences can then be used for correlating other deposits in which paleomagnetic or radiometric age data are not available. The major sequences with geochronological control are in East Africa, whereas the occurrences in southern (and northern) Africa are generally deficient with regard to isotope age determinations and paleomagnetic data.

East African Plio–Pleistocene deposits

The highland areas of East Africa, including Ethiopia, have been strongly affected by tectonic and volcanic activity since at least the early Miocene, and the rift valleys and downwarped areas have provided unusually good traps for the accumulation of fossil-bearing sediments, in many cases associated with lavas or volcanic tuffs that furnish radiometric or fission-track ages. Geochemical "fingerprinting" of tuffs has also been employed successfully as a correlation tool. The broad geological framework has been described by a number of specialists, and the fossil mammals have been treated extensively (Maglio and Cooke, 1978).

The most important deposits of Plio–Pleistocene age occur in six areas: (1) the Awash Valley and adjoining areas in north-central Ethiopia, (2) the lower Omo Basin in southwestern Ethiopia, north of Lake Turkana, (3) the areas flanking the northern half of Lake Turkana, Kenya (the Koobi Fora area to the east and the Nachukui area to the west), (4) the Olduvai Gorge, Laetoli, and Lake Natron areas in northern Tanzania, (5) the Lake Baringo area in north-central Kenya, (6) western Kenya and the Western, or "Albertine," Rift in Uganda, and the Upper Semliki area in Zaire. The principal localities are shown in Figure 27.1. Magnetostratigraphic sequences that include the Matuyama (Plio–Pleistocene) interval have been most carefully worked out in the Omo Basin and Olduvai and are reasonably well established at Koobi Fora and in parts of the Awash and Baringo successions.

Awash Valley, Ethiopia

A portion of the Middle Awash region, opening into the Afar depression, is flanked by sediments ranging in age from mid-Miocene to Holocene. The formations recognized in that sequence have been designated the Awash Group by Kalb et al. (1982), although the term has not been widely used by others. The Hadar Formation, some 200 m thick, contains important hominid fossils, and for that reason it has been described in detail (Aronson et al., 1977; Johanson, Taieb, and Coppens, 1982; Tiercelin, 1986), and the fauna has been analyzed (White, Moore, and Suwa, 1984). Although the exact age range of the Hadar Formation was a subject of controversy for some years (Brown, 1982; Aronson, Walter, and Taieb, 1983), the modern consensus is that the unit ranges between 3.5 Ma and 2.9 Ma (Haileab and Brown, 1992; Walter, 1994) and thus is clearly mid-Pliocene in age. Substantiation comes from geochemical correlations of the Sidi Hakoma Tuff (SHT) in the upper Hadar with (1) the Tulu Bor Tuff at Koobi Fora, interpolated to 3.39 Ma in the orbitally tuned time scale (Walter and Aronson, 1993), (2) Tuff B in the Shungura Formation of the Omo area, and (3) marine tuffs in the Gulf of Aden (Sarna-Wojcicki et al., 1985).

The Matabaietu Formation in the Awash Group is not dated radiometrically, and its stratigraphic relationship to the Hadar Formation is not established, but it has a younger, apparently late Pliocene, fauna. It is overlain by the Wehaietu Formation, also undated, which has a fauna indicative of a time range beginning in the later part of the early Pleistocene (Kalb and Mebrate, 1993). The Bodo member, in the lower part of that formation, has yielded Acheulean artifacts and the cranium of an archaic *Homo* (Conroy et al., 1978; Kalb et al., 1980). There are other significant paleoanthropological occurrences in the same valley (Clark et al., 1984) and on the plateau near Gadeb (Clark, 1987).

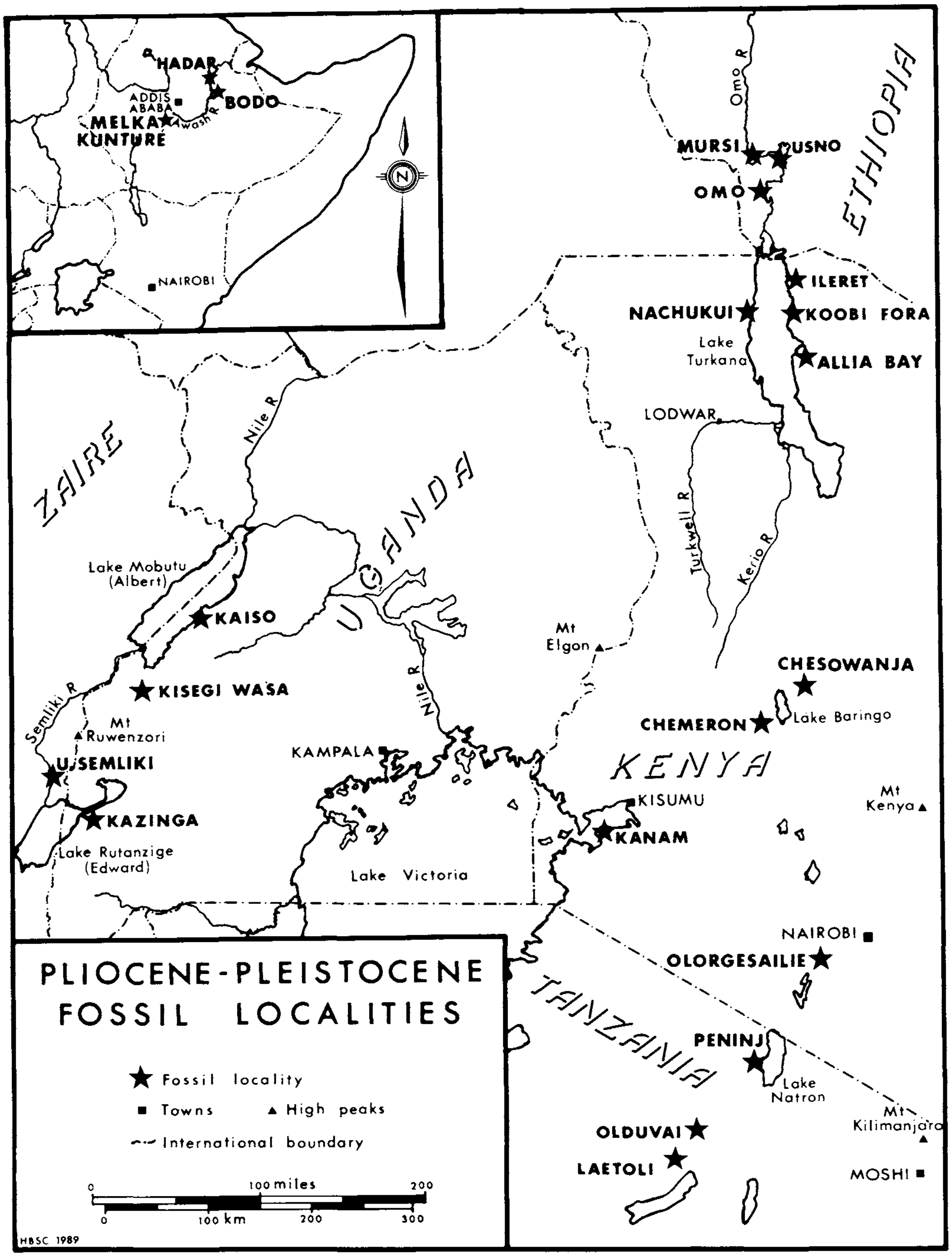

Figure 27.1. Principal fossil localities of Plio–Pleistocene age in East Africa. The inset shows the major localities in Ethiopia.

Terraces on the upper Awash River, 50 km south of Addis Ababa at Melka-Kunturé in the central Ethiopian Rift, have yielded important artifact assemblages on occupation floors, as well as hominid and other (as yet largely undescribed) mammal fossils, referable to eight cultural levels spanning the entire Pleistocene (Chavaillon et al., 1979). Radiometric dating of the tuffs and magnetostratigraphy at Melka-Kunturé (Cressier, 1980) indicate that the oldest levels are within the midpart of the Matuyama reversed-polarity zone, from 1.6 Ma to about 1.0 Ma. Chavaillon et al. (1979) stressed the similarity between the Melka-Kunturé "Olduwan" culture and that of Beds I–II at Olduvai Gorge, but the reversed polarities of the sediments in Melka-Kunturé Beds 1 and 2 suggest that they are equivalent only to the later "Oldowan" cultural levels in upper Bed II (discussed later). Evidently, the base of the sequence is very close to the Pliocene–Pleistocene boundary, as defined at Vrica.

Another Ethiopian Plio–Pleistocene sequence was described at Konso-Gardula (Asfaw et al., 1992). The upper levels of this section, dated within the range 1.50–1.40 Ma, contain an abundant fauna, with early *Homo erectus* and Acheulean-type tools, closely comparable to Olduvai Bed II. The Chari Tuff of Koobi Fora (1.39 Ma) is recognized at the top of the fossiliferous sequence, while the level of the proposed base of the Pleistocene may be interpolated to lie somewhat below the fossil beds and slightly above an exposure of the KBS Tuff (1.88 Ma). The Plio–Pleistocene interval is also documented in Djibouti, at the mouth of the Afar Rift, where a diverse large-mammal fauna at Anabo Koma, with Olduvai Bed I affinities, is dated to 1.9 Ma (de Bonis et al., 1988), and an Acheulean butchering site at Barogali, with remains of *Homo erectus*, has been given an age of 1.5 Ma (Amosse et al., 1991).

Lower Omo River basin, Ethiopia

The Omo River drains into the north end of Lake Turkana (formerly Lake Rudolf). The lower reaches of the river within Ethiopia are flanked by more than 800 m of tilted and faulted fossiliferous sediments of the Shungura Formation; there are also two isolated areas where the Usno and Mursi formations are exposed (Brown, de Heinzelin, and Howell, 1970). The geological setting has been described in detail (de Heinzelin, 1983), and the fauna has been reviewed (Coppens and Howell, 1985). Numerous tuffs are interbedded with the Shungura strata, the most prominent of which are used as marker horizons to divide the sequence into alphabetic members, with the respective "marker" at the base of the member. Many radiometric ages between 5 Ma and 0.5 Ma and a detailed paleomagnetic study provide a very well controlled section that has become the reference standard for the Pliocene and early-to-middle Pleistocene of East Africa (Brown et al., 1985; Feibel, Brown, and McDougall, 1989). The top of the Olduvai subchron almost coincides with Tuff J of the Shungura Formation, so that the boundary between Members H and J (there is no member I) is effectively very close to the Pliocene–Pleistocene boundary. The Jaramillo subchron has not been recorded and is considered to be slightly younger than the top of Member L. The succession is set out in the general correlation diagram in Figure 27.2.

In the northeast corner of the basin, near the Fejej police post, a fossiliferous sequence has been described by Asfaw et al. (1991, 1993) straddling the Pliocene–Pleistocene boundary. Locality FJ-5, with an abundant fauna, is dated to 1.7 Ma by the presence of the Koobi Fora Orange Tuff, while FJ-1, with both abundant fossils and tools, is dated to 1.9 Ma on the basis of its relationship to a tuff identified with Shungura Tuff H-1 (Asfaw et al., 1993).

Lake Turkana basin

The fossil-rich Koobi Foora Formation is exposed in extensive badlands on the northeast side of Lake Turkana (Bowen and Vondra, 1973). The aggregate thickness of the formation is about 450 m, but there are breaks within the sequence and wide alluviated gaps between different exposure areas that render correlation difficult (White et al., 1981). Tuffs are as common as in the Shungura Formation, but their use has been hindered by the physical stratigraphic problems, and as a result, some of the key horizons were initially miscorrelated (Cerling et al., 1979; Brown and Cerling, 1982). On the basis of geochemical "fingerprinting," which identifies key tuffs occurring in both the Koobi Fora and Shungura formations, Brown and Cerling (1982) proposed a revised scheme that has culminated in a major revision of the stratigraphic nomenclature (Brown and Feibel, 1986). This is backed by a substantial number of radiometric dates (McDougall, 1985; Feibel et al., 1989). Paleomagnetic data initially were somewhat inconclusive (Brock and Isaac, 1976), but now confirm the revised stratigraphy and correlations (Hillhouse, Cerling, and Brown, 1986; Feibel et al., 1989).

The Koobi Fora hominids and a portion of the mammalian fauna have been described in monographs (Leakey and Leakey, 1978; Harris 1983). The Okote and Morutot tuffs (parts of a closely spaced complex that also includes the lower Koobi Fora Tuff) are radiometrically dated to 1.64 and 1.65 Ma, respectively, and are geochemically correlated with tuffs in the lower part of Member J of the Shungura Formation, just above the Olduvai subchron and thus close to the Pliocene–Pleistocene boundary (Figure 27.2). The KBS Tuff, which is dated firmly at 1.88 Ma and is physically correlated with Shungura Tuff H-2, is at the base of the Olduvai normal-polarity interval at Omo.

Prospecting on the west side of Lake Turkana has disclosed a sequence of deposits generally similar to those of the Omo and Koobi Fora areas, and that has led to the discovery of a remarkable partial skeleton of an australopithecine (Walker et al., 1986). Named the Nachukui Formation, the deposits are divided into members by prominent tuffs, several of which can be correlated geochemically with those of the Koobi Fora Formation (Harris, Brown, and Leakey, 1988). The Natoo Member has the Lower Koobi Fora Tuff at its base – one of the units of the Okote complex, dated to 1.63 Ma – and thus nearly coincides with the Pliocene–Pleistocene boundary (Figure 27.2). The paleogeography of the whole Turkana Basin has been outlined (Brown and Feibel, 1988) and shows that the basin was not occupied continuously by a lake.

Olduvai, Laetoli, and Lake Natron

North of Lake Eyasi in Tanzania are two important sets of beds representing a single succession from the Lower Pliocene to the uppermost Pleistocene. The classic section at Olduvai Gorge comprises about 100 m of sediments divided by Reck (1914) into a series of "Beds," numbered I, II, III, IV, and V. The geology has been discussed in detail by Hay (1976). The Olduvai beds were the first sedimentary strata in Africa to be dated radiometrically (Leakey, Evernden, and Curtis, 1961) and the first to be analyzed for remanent paleomagnetism. The originally measured ages for the upper and lower limits of the normally magnetized zone in Bed I and lowermost Bed II, termed the "Olduvai event" by Grommé and Hay (1971), were 1.65 Ma and 1.8 Ma, respectively, based on 57 K/Ar ages. These limits have been revised by subsequent laser-fusion $^{39}Ar/^{40}Ar$ analysis to 1.75

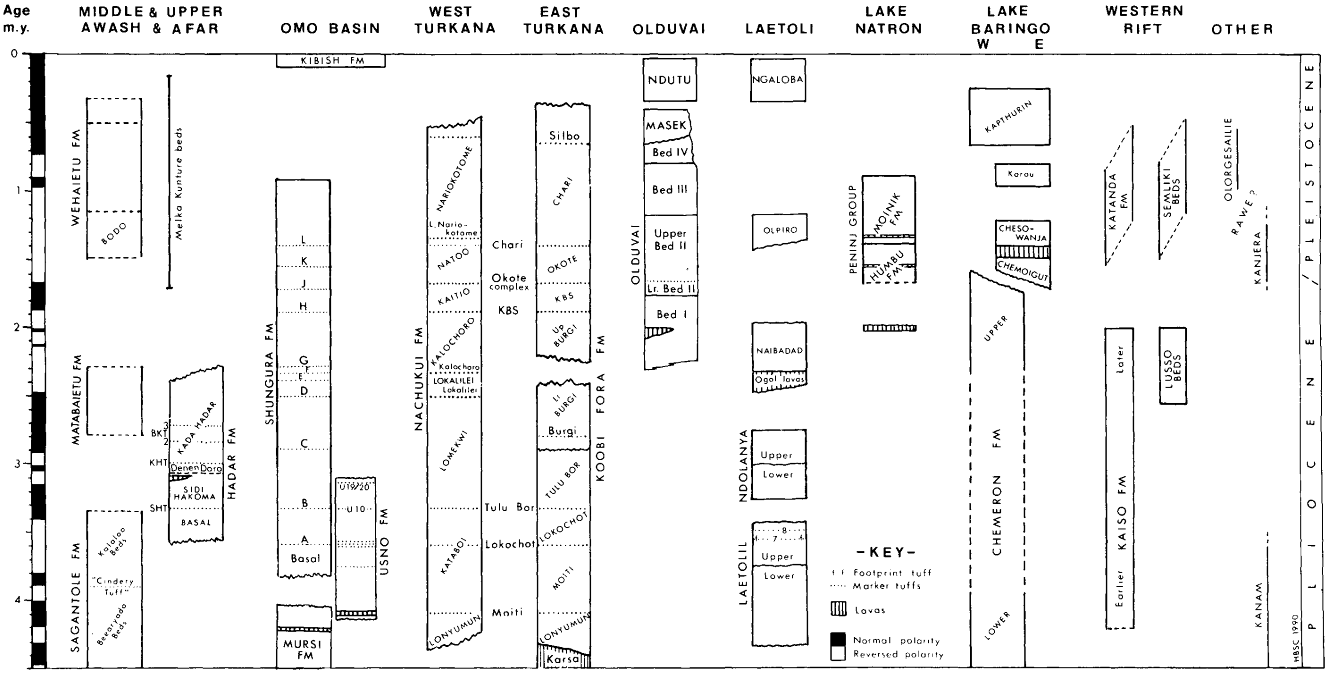

Figure 27.2. Provisional correlation of the main fossiliferous stratigraphic units of Plio–Pleistocene age in East Africa. Most of the sequences are controlled by radiometric dates and/or paleomagnetic records (age values are those of the referenced studies; for orbitally tuned values, see the Preface to this volume). Faunal data are also employed in correlation between sites.

and 2.0 Ma by Walter et al. (1991), closely comparable to the values obtained by astronomical calibration, as noted in the Preface to this volume.

In the type locality for the Olduvai subchron, the top is now identified within the Lemuta Member, which forms the upper part of Lower Bed II (Michael, 1992). The Lemuta Member represents an accumulation of largely eolian material deposited during a relatively dry interval and truncated by a regional disconformity marking the base of Upper Bed II. The marked change in the fauna above the Lemuta Member was associated with a change in the environment, but that does not seem to have been a regional trend. At Laetoli, to the south of Olduvai, the Laetolil Formation contains important hominid fossils, as well as a hominid trackway preserved just below an ash horizon dated at 3.6 Ma (Leakey, 1981; Drake and Curtis, 1987; Hay, 1987). The lower part of the succeeding Ndolanya unit is barren, but the upper part has a good fauna, overlain unconformably by the Ogol lavas, dated to 2.4 Ma. The unconformably overlying Naibadad beds have a fauna comparable with that of Olduvai Upper Bed II. The Pliocene–Pleistocene boundary would thus occur in the Laetoli section within the erosion interval below the Naibadad beds.

Some 150 km to the northeast of Olduvai Gorge lies Lake Natron, on the west side of which there is a series of largely lacustrine sediments known as the Peninj Group, from which a fine australopithecine mandible and numerous artifacts were collected in the lower unit, the Humbu Formation (Isaac, 1967). That unit contains a basalt within a normal-polarity interval, whereas the sediments above and below it are magnetically reversed. Thouveny and Taieb (1987) suggest that the basalt was extruded during the Olduvai subchron, which is in agreement with fossil evidence that the main Peninj fauna, just above the normal-polarity zone, is comparable to that of the middle of Bed II at Olduvai (Gentry and Gentry, 1978; Geraads, 1987). A basalt near the base of the overlying Moinik Formation is fairly well dated at 1.38–1.33 Ma. It is thus probable that the main fauna of the Peninj Group is early Pleistocene in age, dating to about 1.6 Ma.

Lake Baringo basin

The important sequence exposed on both sides of Lake Baringo consists of about 3,000 m of well-dated volcanics with seven intercalated fossil-bearing units ranging in age from Miocene to mid-Pleistocene (Chapman and Brook, 1978; Hill, Curtis, and Drake, 1986). The lower part of the Chemeron Formation has a fauna that is probably early Pliocene to mid-Pliocene in age, but finds of a younger fauna suggest that the formation may range up to the equivalent of lower Bed II of the Olduvai sequence, of earliest Pleistocene age. On the east side of Lake Baringo, the Chemoigut Formation has furnished Oldowan tools and a hominid fragment. It is overlain by the Chesowanja Formation, the lower basalt of which is dated at 1.42 Ma (Hooker and Miller, 1979). The Chesowanja beds are of limited extent, but have furnished a partial cranium of an australopithecine (Carney et al., 1971; Bishop, Hill, and Pickford, 1978), as well as artifacts from an occupation site with suggestive traces of fire (Gowlett et al., 1981). Both the Chemoigut and the Chesowanja beds are Lower Pleistocene.

Western Kenya, Uganda, and Zaire

At the foot of the Homa volcano in the Winam (= Kavirondo) Gulf in western Kenya, Pliocene and Pleistocene strata crop out in lakeside gullies at Kanam and Kanjera. One of the Kanam sections, Kanam West, has yielded a distinctive fauna that Pickford et al. (1990) considered to be earliest Pliocene and correlative to the lower part of the Kaiso Series (discussed later). Strata resting on the Pliocene have a mammal fauna correlative to upper Bed II at Olduvai and include remains of *Homo erectus*.

The Kaiso Formation occurs in a belt east of the Semliki River from the type area on Lake Mobutu (formerly Lake Albert) to eastern Lake Rutanzige (formerly Lake Edward) (Bishop, 1965), as well as in the Kazinga Channel (Krommenhoek, 1969) and on the northwestern side of Lake Mobutu at Sinda-Mohari in Zaire. The deposits are mainly ferruginous and calcareous oolitic beds deposited in well-aerated fresh-water swamps; they have a good molluscan fauna in which Gautier (1970) recognized 7 or 8 age-related associations. The various collections of fossil mammals from a number of localities originally were divided into an "earlier" and a "later" faunal association (Cooke and Coryndon, 1970), but recently Pickford et al. (1988, 1990) have shown that the sequence is more complex. Two "earlier" units, the Nkondo Beds and Warwire Beds, containing faunas similar to that of Kanam West, have been distinguished with ages estimated at around 5–3 Ma. Two of the Turkana Basin tuffs, the Lomugol (3.60 Ma) and the Lokochot (3.40 Ma), have been identified in the Warwire Beds (Pickford et al., 1991). Overlying these deposits unconformably, the Kaiso Village beds, correlative to the Kyeoro and Hohwa formations, have a fauna equivalent to, or slightly older than, that of Olduvai Bed I. The succeeding Museta beds contain fossils and artifacts comparable to those in Upper Bed II at Olduvai (Pickford et al., 1990), and a tuff layer in correlative strata at Kagusa has been chemically identified as the lower Natoo Tuff of West Turkana, dated to about 1.5 Ma (Pickford et al., 1991). The Museta beds, representing the Lower Pleistocene in Uganda, appear to be part of what was originally named the "Epi-Kaiso" or the Kanda Formation (Cooke and Coryndon, 1970). In Zaire, flanking the upper Semliki River, artifacts have been found in the Lusso beds, which correspond faunally to the Kaiso Village beds, whereas the overlying Semliki beds are probably equivalent to the Katanda Formation (Boaz, 1990). The Pliocene–Pleistocene boundary, based on the foregoing correlations of fossils and tuffs, is just older than the Museta beds, while the Kaiso Village and Lusso beds are of late Pliocene age.

Southern African Plio–Pleistocene deposits

In northern Malawi, the Chiwondo beds (Clark, Stevens, and Coryndon, 1966) contain numerous horizons with fossil mammals that suggest a time range from the early or middle Pliocene to the earliest Pleistocene (Kaufulu, Vrba, and White, 1981; Bromage, Schrenk, and Juwaweyi, 1995). They are overlain unconformably by the unfossiliferous but artifact-rich Chitimwe beds of middle or late Pleistocene age; artifacts have been recovered from the paleosurface at the contact, but despite hopeful claims (Kaufulu and Stern, 1987; Clark, 1990), none have been clearly documented from within the Chiwondo beds (Juwaweyi and Betzler, 1995). Analysis of the age-diagnostic large-mammal taxa in the Chiwondo fossil assemblages, constrained by geology, indicates three main paleontological levels: unit 2, with a fauna dated to 4.0 Ma or earlier; unit 3A, with fossils (including a mandible of primitive *Homo*) which indicate ages between 3.76 Ma and 2.0 Ma or younger; and unit 3B, with a fauna of 1.6 Ma to 1.5 Ma (Bromage et al., 1995; Betzler and Ring, 1995). The Pliocene–Pleistocene boundary appears to fall between 3A and 3B.

Transvaal caves

The so-called cave breccias of the Transvaal and northern Cape Province are important sources of fossil mammals, especially of the extinct hominids known as the Australopithecinae. The first specimen to be found was a fissure filling in a secondary limestone tufa in the Buxton-Norlim quarry at Taung, some 130 km north of Kimberley; the age probably is late Pliocene, but dating and correlation are difficult (Cooke, 1990). The major localities are in the Sterkfontein area, 50 km west of Johannesburg, where the three most important sites (Sterkfontein, Swartkrans, and Kromdraai) are separated by only 1–2 km. Another significant occurrence is at Makapansgat, 250 km northeast of Johannesburg. All those Transvaal breccias are essentially infillings of subsurface caves and fissures, into which external soil was carried and then firmly cemented. Erosion has removed most of the former cave roofs, so that the pink and brown breccias, locally rich in bone, are now exposed at the surface. The process of formation has been clearly set out by Brain (1958, 1993).

The stratigraphy of the Sterkfontein deposit was formalized by Partridge (1978), who distinguished several successive members (1–4), starting with a basal travertine, as well as a younger unit (Member 5) separated by a substantial interval. Deposits of unknown age have been reported from deeper levels in the cave system (Wilkinson, 1985). At Makapansgat, Partridge (1979) proposed a division similar to (but not correlated with) that of Sterkfontein, again with four successive members and a later fifth unit. Maguire (1985) suggested that in some cases facies changes have been confused with stratigraphic succession. The complexity of the cave deposits is well illustrated at Swartkrans, where two members, separated by extensive erosion, were originally distinguished (Butzer, 1976; Brain, 1981). Subsequent work has led to the recognition of five units (Brain, 1985, 1988, 1993; Brain et al., 1988). Member 1, which contains abundant hominid and other fossils, as well as "Developed Oldowan" tools, was heavily eroded, leaving a "Hanging Remnant" unit adhering to the north wall of the original cave, and a "Lower Bank" unit on the floor. The space was then filled by the brown breccia facies of Member 2 and calcified before another cycle of erosion cut a deep gully, which was then filled with Member 3. Although there were some changes in the fauna and the artifacts, those three members do not appear to differ very greatly in age. Similar cycles led to the formation of Member 4, with Middle Stone Age tools, and Member 5, of late Pleistocene or Holocene age. At Kromdraai, there are two sites, a "faunal site" (Kromdraai A) and the "australopithecine locality" (Kromdraai B), which are not of the same age. Excavations at Kromdraai B disclosed two areas, not demonstrably connected, now termed Kromdraai B West (KBW) and Kromdraai B East (KBE). At KBE, Partridge (1982) named five members, of which Member 3 is the source of the hominid material and of the fauna (Vrba, 1982).

The faunas from all the sites have been discussed by Brain (1981), although not from the point of view of dating. Various analyses of the faunas have indicated that the Makapansgat and Sterkfontein are the oldest and correspond best with the lower part of the Shungura Formation in East Africa, whereas the Swartkrans Member 1 and Kromdraai KBE faunas have closer affinities with that of Bed II at Olduvai or that of the upper–middle part of the Shungura succession (Cooke, 1974; Vrba, 1975, 1976, 1988; Harris and White, 1979; Delson, 1984, 1988). The magnetic properties of the cave breccias are far from ideal for paleomagnetic studies, but results at Makapansgat (McFadden and Brock, 1984) show reversed polarity in Member 1 and lower 2, whereas upper 2 is normal. Member 3 (which is the main source of the fossils) is magnetically poor, but Member 4 shows two narrow reversed zones followed by a normal sequence. In conjunction with the faunal evidence, that suggests an age in the middle of the Gauss normal-polarity chron involving the Kaena or Mammoth subchron, providing an age close to 3 Ma. However, the possibility of lateral facies variation makes precision impossible. Samples from Sterkfontein Members 2, 3, and 4 are magnetically rather unsatisfactory. They show mainly normal polarity, but there are several intermediate or reversed polarities, and the best fit is probably in the middle to upper part of the Gauss (Jones, Brock, and McFadden, 1986). The samples from Swartkrans are unreliable. Kromdraai KBE shows predominantly reversed polarity and is likely to lie within the Matuyama chron (Jones et al., 1986).

A tentative correlation of the breccia sites is given in Figure 27.3. The Pliocene–Pleistocene boundary of Vrica would seem to lie close to Member 5 at Makapansgat and Sterkfontein, and perhaps within Swartkrans Member 1. Kromdraai KBE may be very late Pliocene, and Sterkfontein and Makapansgat mid-Pliocene.

Climatic and environmental changes

In sub-Saharan Africa as a whole, the fossil faunas represent varying aspects of the mosaic of grassland, open bush, and gallery forest that characterizes the region today. In the

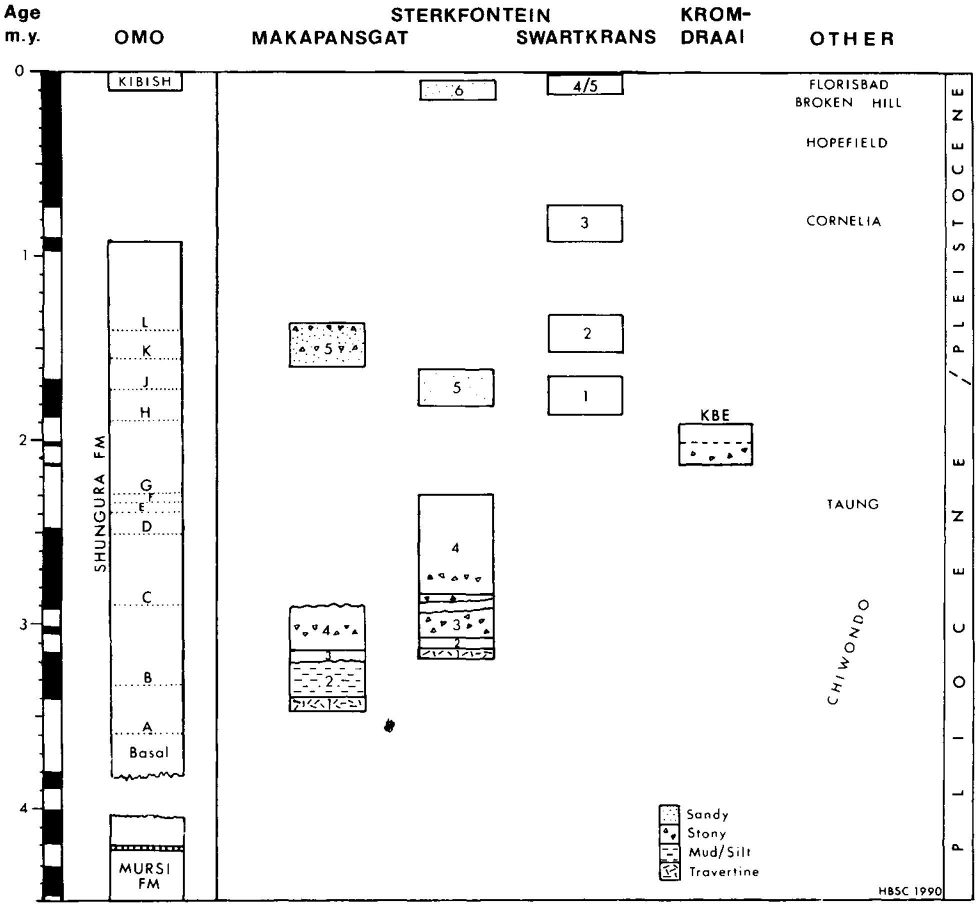

Figure 27.3. Tentative correlation of the main fossiliferous sequences or sites in southern Africa. There are no reliable radiometric ages, and the correlation and relative dating rest mainly on faunal comparisons with the well-controlled East African sequences (typified by Omo). The available paleomagnetic evidence is subject to ambiguous interpretations.

collections from each location there are fluctuations in the proportions of animals that preferred one or another of those three habitats, and there were changes in the local ecological conditions from time to time in each area. Although they probably were due to climate changes, there is no clear evidence of major cyclic events. In many of the major faunal groups, the most marked faunal turnover took place at approximately 2.5–2.4 Ma. The small mammals at Omo show a rapid shift from more mesic to more xeric conditions within Member D (Wesselman, 1984), and that is consistent with the general nature of the change at that time. Palynological data support the observed trend (Bonnefille, 1976, 1983, 1984; Bonnefille and Vincens, 1985). The Bovidae have provided useful analyses, and the case for a global 2.5-Ma "event" reflected in the continental environments of Africa is well made by Vrba (1988, 1992). The genus *Equus* made its first appearance slightly later, around 2.3 Ma. Bovidae have also been used to provide evidence for major changes in local habitat preference and ecology at that time (Vrba, 1982, 1985, 1992; Shipman and Harris, 1988; Bromage et al., 1995). There were minor faunal turnovers at about 1.8 Ma and again at 1.5 Ma, but nothing that can be said to mark the Pliocene–Pleistocene boundary as a period of significant change. There is also isotopic evidence from carbonates for dramatic changes in rainfall in the Lake Turkana region between 2.0 and 1.8 Ma, and at Olduvai a minor rainfall change at 1.8 Ma, and a major one between 0.5 and 0.6 Ma (Cerling, Hay, and O'Neil, 1977). The change at Olduvai at 1.8 Ma was approximately synchronous with the aforementioned faunal change, but the latter seems to have been due to local rather than regional trends, and its coincidence with the Quaternary boundary adopted at Vrica probably does not signal an important event in Africa, unlike the climatic and faunal changes at 2.5 Ma.

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28 Plio–Pleistocene reference sections in Indonesia

FRANÇOIS SÉMAH

Introduction

Quaternary research in Indonesia developed rapidly beginning in the late 1970s because of the activities of Indonesian, French, Dutch, and Japanese teams (e.g., Saint-Marc and Suminta, 1979; Sémah et al., 1980; De Vos et al., 1982; Sartono, Sémah, and Djubiantono, 1984; Watanabe and Kadar, 1985; F. Sémah, 1986). Understandably, research focused mainly on Java, where sites with human remains have been known for more than a century. Comparable information on the Plio–Pleistocene strata in other areas of Indonesia is not yet available to give a complete synthesis of the region, and this chapter is therefore restricted to the island of Java.

The manuscript for this chapter was initially submitted in 1984, with revisions to certain essential points in 1991. The time-scale values have been corrected in the text according to orbitally tuned calibration, as discussed in the editor's Preface. It was not possible, however, to include a balanced review of all the work that has been published in the past decade, and this report must therefore be considered as a reflection of the state of the art in the mid-1980s.

Geological background

Exhaustive field work by Dutch geologists during the first half of the century was fairly well summarized by Van Bemmelen (1949) and Marks (1957). This report follows the standard terminology of Marks (1957) for Javanese formations, in order to maintain a connection with previous works, without implying acceptance of the chronostratigraphic meanings and correlations proposed by earlier authors. In our view, the assignment of formation names has been overdone in publications dealing with Javanese stratigraphy, particularly with regard to the Plio–Pleistocene. Recent work has shown that many stratigraphic units have such limited geographic extensions, or vary in age so greatly from place to place, that they hardly deserve the name of "formation." This is particularly the case in the Sangiran Dome area (discussed later), where lateral facies changes are very abrupt.

Java can be divided into several structural units, roughly oriented E–W, following the axis of the island (Figures 28.1 and 28.2). The most nearly complete Plio–Pleistocene sections are exposed in the central depression (Solo zone) and near the axial anticlinorium (Bogor zone, North Seraju and Kendeng ranges). We shall describe here two of the areas which illustrate the Pliocene–Pleistocene boundary, namely the Bumiayu area of central Java south of Tegal, and the Surakarta region of central Java on the upper Solo River.

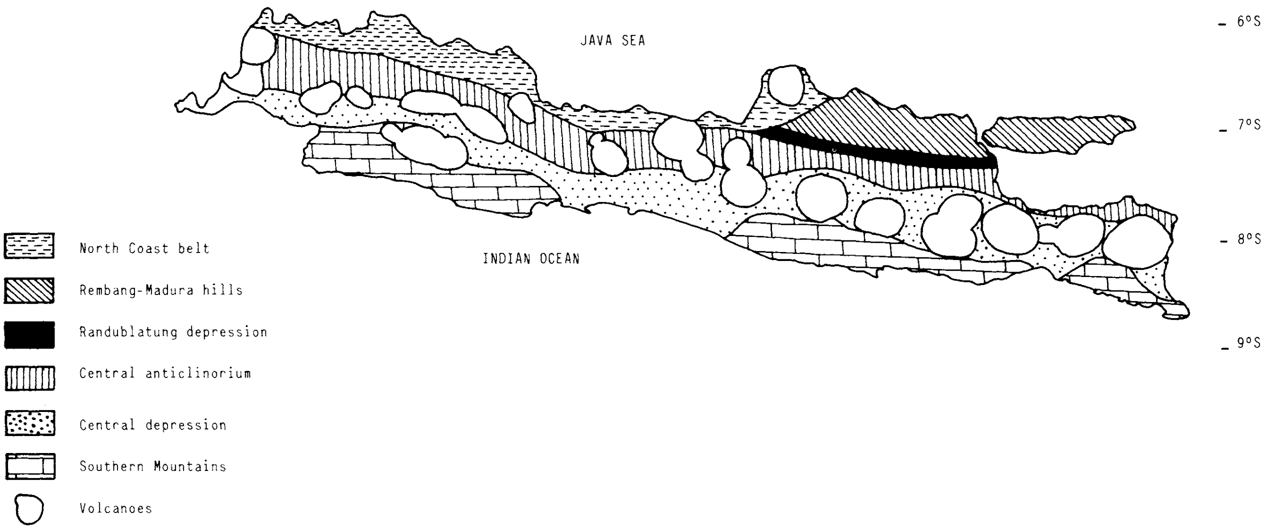
The stratigraphic succession in these two areas is shown in Table 28.1. Exposures, paleontology, and other features are discussed next under separate headings.

Elevation of the Southern Mountains of Java had already taken place before the late Cenozoic (Van Bemmelen, 1949). The main factor in the Plio–Pleisticene history and stratigraphy of central Java was the emergence of the axial anticlinorium, for instance, the Kendeng Hills (e.g., Saint-Marc, Paltrinieri, and Situmorang, 1977). The present consensus among geologists working in Java is that tectonic uplift progressed gradually from west to east, with most of the island being emergent by the end of the Pliocene. Other important factors in the stratigraphy were eustatic shifts in sea level, beginning about 2.5 Ma, and local activity in the inner volcanic arc, which began at least 2 m.y. ago (Bandet et al., 1989).

Bumiayu area, western central Java

At the foot of the North Seraju mountains, in the Bumiayu region, sections in the valleys of the Kali Glagah and Ci Saat rivers exemplify the sequences of the Plio–Pleistocene interval. Resting on Miocene and Lower Pliocene units, the Kalibiuk Formation, with marine mollusks, is followed by the Kali Glagah–Mengger–Gintung synorogenic series, in Ter Haar's terminology (1934) (Figure 28.3). The Kalibiuk–Kali Glagah boundary is located, in the Ci Saat Valley, at the first occurrence of continental sediments. The Kali Glagah and Mengger formations contain sandy, conglomeratic, and clayey layers, whereas the Gintung has a mainly conglomeratic and volcanic breccia aspect.

At least two vertebrate horizons have been recognized in that area, which is the type for the so-called Kaliglagah fauna. The lower one is situated a few tens of meters above the Kalibiuk–



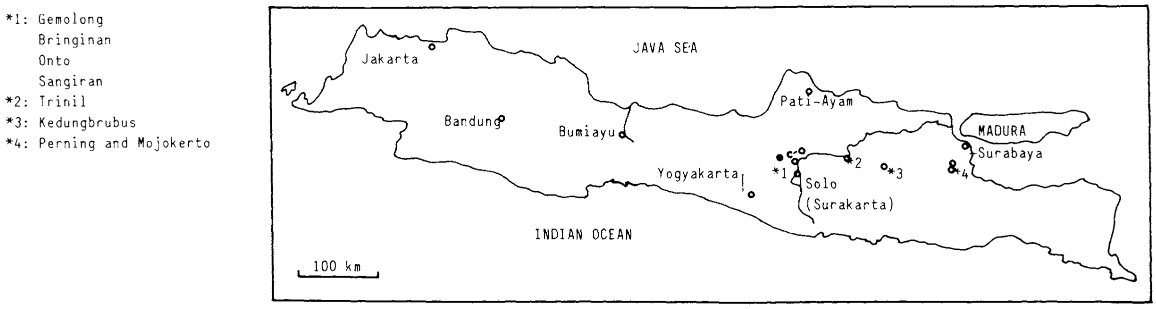


Figure 28.1 (top). Structural geology of Java. (Adapted from Van Bemmelen, 1949.) **Figure 28.2 (bottom). Locations of sections discussed in this chapter: 1, Gemolong, Bringinan, Onto, Sangiran; 2, Trinil; 3, Kedung Brubus; 4, Perning and Mojokerto.**

Table 28.1. *Summary of Plio–Pleistocene stratigraphy in the studied areas of central Java*

West Central Java (Bumiayu): Ci Saat, Kali Glagah valleys	
Gintung	Conglomerate.
Mengger	Continental sand, gravel, clay.
Kali Glagah	Continental sand, gravel, clay.
Kali Biuk	Marine sands and clays.
East Central Java (Surakarta): Solo River basin, Sangiran dome	
Notopuro	Volcanic breccias.
Kabuh	Tuffaceous sands, clays, gravels; basal *Grenzbank* conglomerate.
Pucangan	Black estuarine and paludal clay, basal lahar over *Corbicula* beds.
Kalibeng	Blue clays, with littoral deposits at top.
Globigerina marls	Open-marine deposits.

Kali Glagah boundary (von Koenigswald, 1933, 1934; Marks, 1957), while the upper one is less clearly localized in the stratigraphy. According to Sondaar (1981, 1984), the upper horizon reflects several faunal exchanges with the mainland that are not evident in the lower one. That might indicate one or more glacial intervals, with sea level decline and exposure of Sundaland interconnections.

Paleobotanic data presented by A.-M. Sémah (1986) show the dominance of a mangrove-forest association in the upper Kalibiuk unit, with a maximum in the lignitic layer which marks the Kalibiuk–Kali Glagah boundary along the Ci Saat River. The pollen data show a trend to a drier climate in the upper Kalibiuk beds; in the lower Kali Glagah, whereas the persistence of mangrove elements indicates proximity of the shoreline, the pollen spectra indicate development of a true rain forest on the higher ground.

Paleomagnetic studies in the Ci Saat Valley (F. Sémah, 1983, 1986) have shown a long reversed sequence (100 m thick) in the upper Kalibiuk marine layers. This long interval appears to correlate with the early Matuyama chron, on the basis of preliminary micropaleontological data (Sumarso and Suparyono, 1974). Several normal sequences that appear within the Kali Glagah series (F. Sémah, 1983) are tentatively assigned to the normal events (Réunion and Olduvai subchrons) which interrupted the reversed polarity of the Matuyama chron between 2.15 and 1.77 Ma. Higher in the section, reversed polarities persist through the Mengger up to the Gintung unit. Up to now, no direct geochronological dating has been carried out in the Bumiayu area, but the paleomagnetic interpretation indicates that the lower of the two vertebrate-fauna horizons in the Kali Glagah dates to the early Matuyama, between 2.1 and 1.8 Ma.

Surakarta (Solo) area, eastern central Java

South of the Kendeng Hills, near the town of Surakarta, the Solo central depression contains giant volcanoes active from the late Pleistocene to the present, and also deeply eroded and faulted domal uplifts that are the remnants of much older volcanic complexes, such as Sangiran, Gemolong, Bringinan, and Onto. In the uplifted domes, exposed Plio–Pleistocene sections are divided into the so-called formations of Kalibeng, Pucangan, Kabuh, and Notopuro, following the classic terminology (Marks, 1957) in that part of Java.

The Solo "formations" have little to do with the type sections defined some 100–150 km to the northeast (Duyfjes, 1936), and stratigraphic correlations based strictly on these lithologies are extremely problematical. For instance, blue marine clays, which are commonly attributed to the "Upper Kalibeng," have a late Pliocene age at Sangiran, but were formed in other parts of the Solo area at various times from the earliest Pleistocene

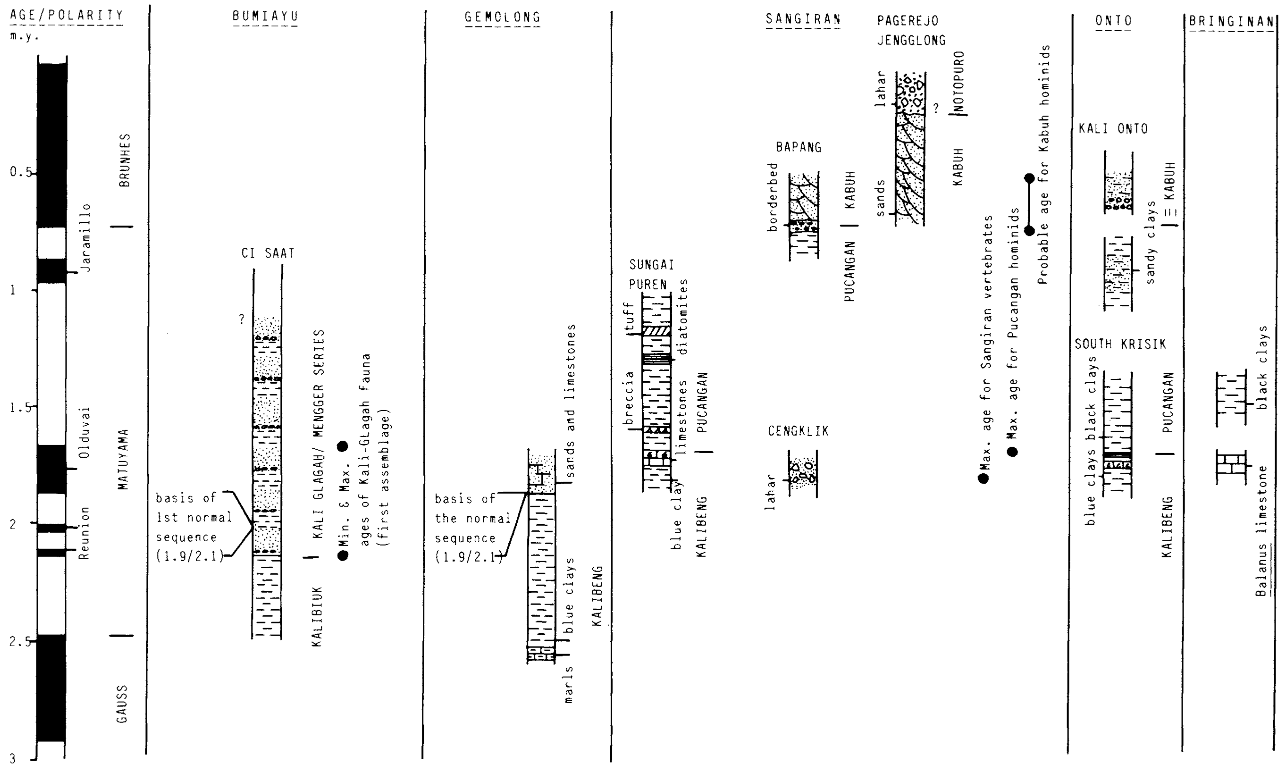


Figure 28.3. Proposed correlations of Plio–Pleistocene sequences of central eastern Java. For locations, see Figure 28.2. (Adapted from F. Sémah, 1984b.)

(Gemolong section) to the beginning of the middle Pleistocene (Kaliuter section, 10 km north of Gemolong) (Djubiantono and Sémah, 1991). These sediments derive from the shallow-marine environments that persisted in the Solo area until the final uplift of the Kendeng Hills. It is obvious that to group such different exposures on a geological map as a regional formation would be illogical and misleading, but rather than create a separate formation name in each exposure, we use these "formation" names in the sense of facies units. It should be noted, however, that Itihara, Kadar, and Watanabe (1985a) have proposed new names for units based on type sections in the Sangiran Dome, namely the Puren, Sangiran, Bapang, and Pohjajar formations.

The whole of the stratigraphic sequence is not exposed in every dome, and for the sake of clarity we shall discuss here a synthetic section (Figure 28.3). Besides the fundamental publications noted earlier, detailed sections and descriptions of the Solo area can be found in the works of F. Sémah (1983, 1986) and Watanabe and Kadar (1985).

Stratigraphy

The upper Kalibeng beds in the Sangiran-Gemolong area begin with blue clays containing shallow-marine mollusks and thin tuffaceous beds. Those rest on open-marine lower Kalibeng *Globigerina* marls, seen only in the Gemolong Dome (Reinhold, 1937). In the Sangiran area, the blue clays pass upward into sandy littoral deposits including calcareous *Turritella* sands and *Balanus* limestones, capped by a bed rich with the fresh-water (swamp) mollusk *Corbicula*.

In the Sangiran section, lahars (volcanic mudflow breccias) mark the boundary between the Kalibeng and Pucangan beds. The lahars, which cover an irregular topography and range from 0.5 m to more than 50 m thick (Indonesia–Japan Research Cooperation Program, 1979), played an important part in ponding up the drainage to form a lagoon in the Solo area. The lagoonal deposits of the lower Pucangan beds reflect low-energy, poorly oxygenated sedimentation dominated by bluish-gray and black clays, with a fauna consisting mainly of fresh-water mollusks (*Corbicula, Viviparus, Unio*), together with estuarine intercalations characterized by marine mollusks and diatoms (Reinhold, 1937; Itihara et al., 1985a) and several fine-grained tuffaceous layers. The upper Pucangan beds consist of poorly stratified black clays. A drastic change is seen at the top of the Pucangan unit, with the transition to high-energy fluviatile sedimentation of the Kabuh beds, mainly sands, gravels, and tuffs. Conspicuous cliff-forming conglomerates, or Grenzbanks ("boundary beds"), mark the end of marine/lagoonal sedimentation at the base of the Kabuh in the Sangiran Dome. The Kabuh beds are unconformably overlain by the Notopuro volcanic breccias and lahars.

Paleontology

The oldest vertebrate remains in the Sangiran Dome seem to come from the basal Pucangan lahar, which includes much sandy clastic material (Van Es, 1931). Above that level, Sondaar (1984) recognized a faunal succession within the Pucangan black clays, equivalent to the upper Kali Glagah succession of the Bumiayu area, with the so-called Satir and Ci Saat faunas. The basal Grenzbank and the lower part of the Kabuh formation (in the terminology of De Vos et al., 1982) yield a Trinil fauna, and examples of the Kedung Brubus fauna have been collected from the upper Kabuh.

Hominid remains from Sangiran are not precisely located in the stratigraphy. The oldest, including the Sangiran *"Meganthropus,"* appear in the upper Pucangan formation (Sartono, 1982) and might be more than 1 m.y. old. A few hominid remains are found in the Grenzbank, although that level yields very abundant fossils of other mammals, and most of the hominid specimens are from the lower and middle Kabuh formation. Of the approximately 18 hominids from Sangiran, all are classed as *Homo erectus*, in the wide sense.

According to A.-M. Sémah (1982, 1984, 1986), the palynological record begins in the upper Kalibeng blue clays in the Sangiran and Onto Domes, with indications of mangrove forest and thus a nearby shoreline. Basal Pucangan layers show a marked impoverishment of the flora, possibly due to the effects of repeated volcanism. The palynology of the remainder of that formation indicates renewed diversification of near-shore swamp vegetation and a land flora that alternated between tropical rain forest and more open, seasonal forest (possibly corresponding to glacial and interglacial climates, respectively). The Kabuh beds have an overall palynological association indicative of open forest and seasonal rainfall. It is not definitely known whether those variations are to be connected with local phenomena (tectonism, volcanism) or with slight climatic changes. Sémah and Rahardjo (1984) interpreted the change in palynoflora at the base of the Kabuh formation to have an environmental cause, parallel with the change in sedimentology, in agreement with Sondaar (1984), who also found indications of a drier climate in correlative layers. The Australasian tektite event, which appears to correlate with the middle Kabuh (discussed later), is associated in deep-sea cores with paleoclimatic indicators of glacio-eustatic sea-level lowering and cold climate (Schneider, Kent, and Mello, 1992).

Paleomagnetic studies

F. Sémah (1983; Sémah et al., 1980; Sartono et al., 1984) found that the Notopuro and Kabuh formations were of normal polarity and should be related to the Brunhes epoch. That conclusion was based on study of several parallel Kabuh sections. Measurements made after careful thermal cleaning have shown several mixed and even reversed samples, mainly in the Jengglong and Pagerejo sections (F. Sémah, 1986), but the polarities in those samples show little stratigraphic continuity or statistical consistency. The transition from predominantly reversed to predominantly normal paleomagnetic polarities found by these authors near the top of the Pucangan unit at Bapang (southwestern part of the Sangiran Dome) is interpreted to be a reflection of the Matuyama–Brunhes boundary.

On the other hand, the report of Shimizu et al. (1985, p. 286), which actually refers to work completed in 1981, showed a consistent mixed-polarity interval in the lower Kabuh and the uppermost Pucangan. That would suggest that the base of the Brunhes might be located in their analyzed section at a level in the middle Kabuh, very close to levels that have been radiometrically dated to about 0.71 Ma, which is in fact not greatly different from the age of the Matuyama–Brunhes boundary (0.78 Ma). The need for caution in lithostratigraphic correlations in continental-volcanic sequences, even locally, has been emphasized (Lizon-Sureau, 1979; A.-M. Sémah, 1984), and there is the strong possibility that the observed Pucangan–Kabuh boundary, with its Grenzbank facies, may actually be diachronous with regard to the Matuyama–Brunhes polarity boundary.

The Jaramillo subchron has not been recognized in the Sangiran Dome. In the Onto dome, however, F. Sémah (1983) observed a short doublet of normal polarity in the upper part of the local Pucangan facies that possibly could be evidence of that episode. F. Sémah et al. (1980) found consistent reversed polarities in samples taken below the upper beds of the Pucangan unit in Sangiran and referred those levels to the Matuyama in the post-Olduvai interval, a conclusion further supported by the work of Shimizu et al. (1985).

In the basal Pucangan lahars at Sangiran the polarities are not consistent, being reversed in some sections and normal in others. It would seem clear that the volcanism, which is evidenced in all parts of the Dome, was diachronous with regard to the upper boundary of the Olduvai (F. Sémah, 1986). Laterally consistent normal polarities, seen in samples from a significant interval in the upper Kalibeng formation (F. Sémah, 1983; Shimizu et al., 1985), are correlated with the Olduvai event. Samples in the lower part of the blue Kalibeng clays show reversed polarity, corresponding to the early part of the Matuyama chron. Finally, normal-polarity samples have been taken from the base of the blue clays and down into the *Globigerina* marls, which may represent the top of the Gauss chron.

Radiometric ages

Radiometric dating at Sangiran is summarized in Table 28.2. It is beyond the scope of this chapter to discuss all of the results, but a few comments are in order. It must be noted that the Sangiran sequence, despite its formation in a very active volcanic area, includes almost no tuff deposits suitable for radiometric dating. The age determinations we have been able to obtain are therefore not as well substantiated as would be desirable.

Overall, the dating is in accordance with the interpretations of the paleomagnetic succession. Among the fission-track ages obtained by different teams in the tuffaceous layers of the Pucangan facies, only the determinations by Suzuki et al. (1985) require consideration, as the other published fission-track age values would correlate the reversed-polarity Pucangan sediments with the normal-polarity Brunhes epoch. As regards the Kabuh facies, the most coherent dates have been obtained by FT (fission-track) and K/Ar (potassium-argon isotope ratio) analysis of tektite samples. All of the tektites dated by those two methods are close to 0.7–0.8 Ma, even though several of the dated tektites are from different levels. The Kabuh tektites have been geochemically correlated to a microtektite horizon found in deep-sea cores about 12 k.y. below the Matuyama–Brunhes boundary (Glass and Heezen, 1967; Schneider et al., 1992). The specimens found in the Kabuh facies at Sangiran are of large size and rather rare. Only three specimens are reported to have been found in situ in the middle Kabuh beds (Itihara et al., 1985a), while the others were from higher in the sequence. Those heavier stones have been considered by most workers to represent specimens that remained at the primary tektite horizon, but it is possible that all of the Kabuh tektites were reworked from an even older level, and the dating of the middle Kabuh on this basis is not yet established beyond question.

An age of 0.8 Ma for the middle Kabuh is too old if the tektite horizon is within the Brunhes, according to one possible paleomagnetic interpretation (i.e., F. Sémah, 1983), but it could agree with the findings of Shimizu et al. (1985), as noted previously. Itihara et al. (1985a) justified placing the Matuyama–Brunhes boundary close to the level of the lowest-occurring tektites in the middle Kabuh in the northern Sangiran exposures, according to the fact that Curtis (1981) obtained an average K/Ar age of 0.83 Ma on 4 middle Kabuh pumices, and Suzuki et al. (1985) obtained an FT date of 0.78 Ma on zircon from the Kabuh Middle Tuff, just below the lowermost tektite occurrences (Table 28.2); in modern time-scale calibrations, that dating supports correlation to the Matuyama–Brunhes (0.78 Ma) reversal more strongly than before.

In the long reversed-polarity interval of the Pucangan, Suzuki et al. (1985) obtained stratigraphically consistent FT ages between 1.16 Ma and 1.51 Ma from pyroclastic zircon. This is appropriate for the upper Matuyama below the Jaramillo, as noted earlier. An age of 2.99 Ma has been obtained for the lower Kalibeng within the basal normal-polarity interval, as is appropriate for the upper Gauss. Finally, Table 28.1 shows the only age proposed for the Notopuro lahars, at 0.25 Ma. Overall, the dating cited here is consistent with interpretations that make the Olduvai normal event, 1.95 Ma to 1.77 Ma, equivalent to the normal-polarity interval in the upper Kalibeng formation.

Micropaleontology

Micropaleontological data are not as numerous in the Surakarta region as in the Bumiayu region. According to Kadar (Watanabe and Kadar, 1985), the foraminifera of the upper blue clays of the Kalibeng date to the late Pliocene near the top of Zone N.21 in the planktonic foraminifera zonation. That is not in contradiction of the interpretation that the Olduvai event, with the Pliocene–Pleistocene boundary near its top, is represented in the shallow-water sediments above the blue clays, at the top of the Kalibeng facies. In eastern Java, biostratigraphy of the *Globigerina* marls (Saint-Marc and Suminta, 1979) near Ngawi shows that the deposition of those marls lasted until the end of the Pliocene, in contrast to the central area, because of the

Table 28.2. *Radiometric ages from Sangiran*

No.	Sample	Age. Ma	Method	Reference
1.	Notopuro (pumice)	0.25 +/- 0.07	FT	Suzuki et al. 1985
2.	Kabuh (tektite)	0:67 +/- 8.3%	FT	Nishimura et al. 1980.
3.	Kabuh (tektite)	0.70 - 0.72	FT	Suzuki et al. 1985
4.	Kabuh (tektite)	0.71 +/- 0.03	K/Ar	Orchiston & Siesser, 1982
5.	Kabuh (pumice, mean value of 4 ages)	0.83	K/Ar	Jacob & Curtis, 1971; Jacob, 1975.
6.	Kabuh (tuff)	0.78 +/- 0.15	FT	Suzuki et al. 1985
7.	Kabuh (tuff)	0.47 +/- 0.02	FT	Nishimura et al. 1980.
8.	Kabuh (tuff)	0.5 +/- 0.02	FT	Nishimura et al. 1980.
9.	Kabuh (pumice)	1.05 +/- 0.1	K/Ar	Obradovich & Naeser, 1981
10.	Kabuh (pumice)	1.6 +/- 0.7	FT	Obradovich & Naeser, 1981
11.	Pucangan	0.4 - 0.5	FT	Orchiston & Siesser, 1982
12.	Pucangan (tuff)	0.57 +/- 0.32	FT	Nishimura et al. 1980.
13.	Pucangan (tuff)	0.67 +/- 0.04	FT	Nishimura et al. 1980.
14.	Pucangan (tuff)	1.16 +/- 0.24	FT	Suzuki et al. 1985
15.	Lower Pucangan (tuff)	1.49 +/- 0.32	FT	Suzuki et al. 1985
16.	Lower Pucangan (tuff)	1.51 +/- 0.25	FT	Suzuki et al. 1985
17.	Pucangan (basal lahar)	2.0 +/- 0.6	K/Ar	Nishimura et al. 1981
18.	Upper Kalibeng (tuff)	2.99 +/- 0.06	FT	Suzuki et al. 1985

Table 28.3. *Radiometric ages from eastern Java localities*

No.	Sample	Age, Ma	Method	Reference
19.	Pati-Ayam (Muriah basalt)	0.50 +/- 0.02	K/Ar	Koenigswald, 1964
20.	Pati-Ayam (basal Slumprit breccias)	0.85 +/- 0.02	K/Ar	Bandet et al. 1989
21.	Trinil ("Pucangan" lahar)	0.5 +/- 0.3	K/Ar	Bartstra, 1978
22.	Perning/Modjokerto (tuff)	1.9 +/- 0.4	K/Ar	Jacob & Curtis, 1971
23.	Kedungbrubus (Pucangan andesite)	1.91	K/Ar	Jacob, 1978
24.	Kedungbrubus/Gunung Butak (top lower breccias)	1.87 +/- 0.04	K/Ar	Bandet et al. 1989

progressive emergence. That observation is consistent with the placement of the Gauss–Matuyama boundary near the top of the *Globigerina* marls in the Gemolong Dome.

Conclusion

In this chapter we have treated two lithological sequences that include the Olduvai subchron. The adopted level for the Pliocene–Pleistocene boundary at Vrica is associated with the uppermost part of this normal-polarity event, at about 1.8 Ma (Pasini and Colalongo, Chapter 2, this volume). In the Bumiayu area, the boundary probably lies within the continental synorogenic sediments of the Kali Glagah unit. In the Surakarta region, it is situated slightly below the apparently diachronous boundary between the littoral Kalibeng facies and the lahars produced by the volcanism that marked the initiation of the brackish-lagoonal and swampy facies of the Pucangan unit. Global climate changes of the Pleistocene, indicated by successive periods of isolation and mixing in the vertebrate fauna that appear to be related to glacio-eustatic shifts in sea level, may be evident in the Bumiayu sequence.

Several other areas in eastern Java are likely to include the same boundary, according to preliminary dating (Table 28.3) in fossiliferous sites like Perning (Mojokerto), Kedung Brubus, Pati-Ayam, and Trinil.

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29 The Pliocene–Pleistocene boundary in New Zealand

ANTHONY R. EDWARDS

Introduction

Since 1953 the Pliocene–Pleistocene boundary in New Zealand has been set at the base of the Nukumaruan (regional) Stage because that coincides with the abrupt first appearance of the subantarctic bivalve *Chlamys delicatula* (Hutton) in marine sequences of central New Zealand. However, it is now estimated, on the basis of biostratigraphic, magnetostratigraphic, and radiometric data, that the Nukumaruan commenced at about 2.4 Ma and ended about 1.3 Ma and that the Olduvai subchron approximately coincides with the middle of the Nukumaruan. In the stratotype Nukumaruan, and in correlative sections, a marked lithologic change (interpreted as the result of glacioeustatic regression) following the first appearance of *Gephyrocapsa sinuosa* Hay and Beaudry may be correlative with the top of the Olduvai and with the Vrica boundary-stratotype.

Historical background

In New Zealand, the regional stages for Upper Pliocene to Middle Pleistocene strata, in upward sequential order, are Waipipian, Mangapanian, and Nukumaruan. The Nukumaruan Stage is sometimes divided into the Hautawan and (above) Marahauan substages. More or less continuous marine and nonmarine sequences extend from the Pliocene into the Pleistocene in various parts of the country. Internal correlation between the sequences is poor, and for external paleontological correlation, for example with Italy, only the marine sequences are of value. Modern reviews of the Pliocene–Pleistocene boundary have been published by Vella (1975), Jenkins (1975), Hornibrook (1976), Edwards, Hornibrook, and Te Punga (1981), and Te Punga (1981).

A cooling trend indicated by the arrival of the bivalve *Chlamys delicatula* (Hutton), a member of the circum-subantarctic *C. patagonica* species group, was an important factor in the decision of New Zealand stratigraphers in 1953 to adopt the base of the Nukumaruan Stage as the base of the Pleistocene. Since then, despite evidence of earlier coolings and inadequate correlation with the boundary in Italy, there has been a reluctance to change the assignment. Even with substantial information on the Le Castella, Santa Maria di Catanzaro, and Vrica sequences, Edwards et al. (1981) noted that "at least four different correlations . . . can be supported."

Data published on the biostratigraphy and paleomagnetic characterization of the proposed stratotype boundary at Vrica (Backman, Shackleton, and Tauxe, 1983; Tauxe et al., 1983; Zijderveld et al., 1991) provide a new opportunity to correlate between New Zealand and Italy. According to the most recent analyses, the boundary-stratotype at Vrica occurs just below the top of the Olduvai normal subchron at approximately 1.81 Ma.

Nonmarine sequences

On North Island, the nonmarine rocks attributed to the Nukumaruan Stage include lowland clastic deposits that are similar to parts of the upper Nukumaruan (including the stratotype) in the Wanganui Basin (Figure 29.2). On South Island, uplift of the main mountain ranges during the Pliocene and Pleistocene resulted in many local accumulations, sometimes more than 1,000 m thick, of gravel (now deeply weathered) and finer-grained alluvium.

Palynomorphs are used to assign those deposits to regional stages, although correlation with the marine strata on which the stages are based is imprecise. In particular, the stratigraphic range of the "Hautawan" (lower Nukumaruan) palynomorph assemblage zone is uncertain. Its lower limit is probably in the upper Mangapanian (Hornibrook, 1981, pp. 280–281), and its upper limit somewhere near the middle of the Nukumaruan (D. C. Mildenhall, personal communication). Sequences in the northwestern region of South Island are important for their inclusion of glacial deposits attributed to the Ross and Porika glaciations. Those of the Ross Glaciation (Gage, 1945; Bowen, 1967) are within gravel closely overlying a lignite that has a palynoflora similar to those in the mostly pre-"Hautawan" Moutere Gravel of Nelson (Mildenhall and Suggate, 1981). The gravel at Ross lies, apparently conformably, about 30 m above a thin shell bed with Waipipian Stage mollusks (Suggate et al., 1978), so that the Ross Glaciation probably occurred in Waipipian or early Mangapanian time. At Timaru in South Canterbury, cool palynofloras, possibly coeval with the Ross

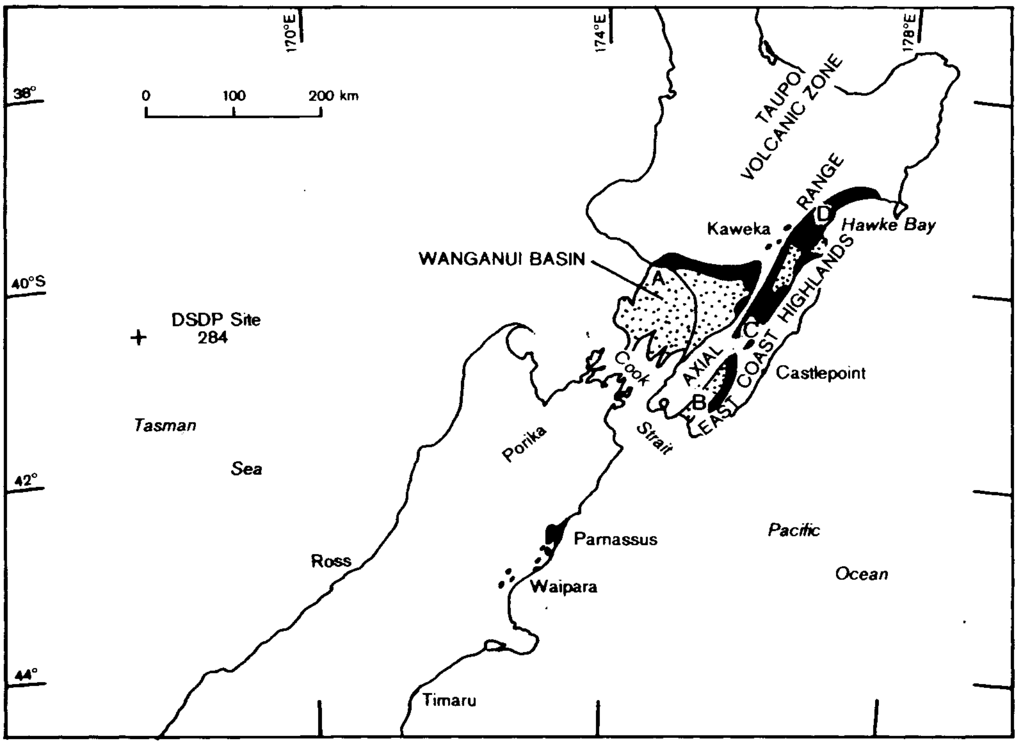

Figure 29.1. Distribution of marine Nukumaruan strata in New Zealand (black is exposed; stippled pattern is subsurface) and locations of the stratigraphic columns in Figure 29.2 as follows: A, Nukumaru Beach; B, Mangaopari Stream; C, Totara Road; D, Devils Elbow. Also shown is DSDP site 284. Small outliers of marine strata occur near Castlepoint, Kaweka, and Waipara; important nonmarine sequences occur near Porika, Ross, and Timaru.

Glaciation, overlie warmer palynofloras of Waipipian age (D. C. Mildenhall, personal communication) and underlie a reverse-polarity basalt (Gair, 1961) with a whole-rock K/Ar date of 2.47 ±0.38 Ma (old constants; Mathews and Curtis, 1966).

Glacial deposits from the Porika Glaciation overlie the Moutere Gravel and contain "Hautawan" palynomorphs (Mildenhall and Suggate, 1981); thus the Porika Glaciation probably occurred in late Mangapanian or early Nukumaruan time, coeval with the first arrival of *Chlamys delicatula* in marine sequences of New Zealand.

The marine Nukumaruan

Nukumaruan marine strata are well developed in central New Zealand (Figure 29.1), where basins continued to subside through the Pliocene and into the Pleistocene. Later uplift and erosion have provided many good sections, especially in the southern part of North Island. In the Wanganui Basin, up to 900 m of very shallow-marine sandstone, mudstone, coquina, and limestone are followed by up to 100 m of nonmarine (and rare marginally marine) clastic strata. In the eastern basins, the Nukumaruan strata typically consist of 100–600 m of more or less cyclically deposited shallow-marine mudstone, shelly sandstone, and limestone; small areas dominated by deep-water conglomeratic sandstone, alternating sandstone/mudstone, or massive mudstone are also known. Most sections contain minor breaks in sedimentation and lithologic changes attributed to glacio-eustatic changes (Beu and Edwards, 1984).

The four on-land sequences summarized in Figure 29.2 are as follows:

1. Nukumaru Beach, Wanganui Basin (Fleming, 1953; Beu and Edwards, 1984): stratotype of the Nukumaruan Stage, very shallow marine, becoming nonmarine upsection.
2. Mangaopari Stream, southern Wairarapa (Kennett, Watkins, and Vella, 1971; Beu and Edwards, 1984): the best-understood late Cenozoic section in New Zealand.
3. Totara Road, southern Hawkes Bay (Hornibrook, 1981): located near a paleo-seaway between the eastern and western basins.
4. Devils Elbow, northern Hawkes Bay (Hornibrook, 1981; Beu and Edwards, 1984): a sequence showing well-developed lithologic cycles.

Together, those four sequences are fairly representative of the shallow-marine Nukumaruan; although a few deep-water sequences are known, they are all incomplete, poorly exposed, or tectonically disturbed.

Biostratigraphy

The New Zealand Upper Pliocene–Middle Pleistocene regional stages are based on the mollusk-rich shallow-marine strata of the Wanganui Basin. Many of the key mollusks have wide but somewhat patchy distributions elsewhere in New Zealand. In the deeper-water strata of inland eastern North Island, the molluscan biostratigraphic scheme can be related to a recently developed microfossil scheme only partly recognizable in the Wanganui Basin (Hornibrook, 1981; Beu and Edwards, 1984). The resultant integrated biostratigraphy provides a reliable

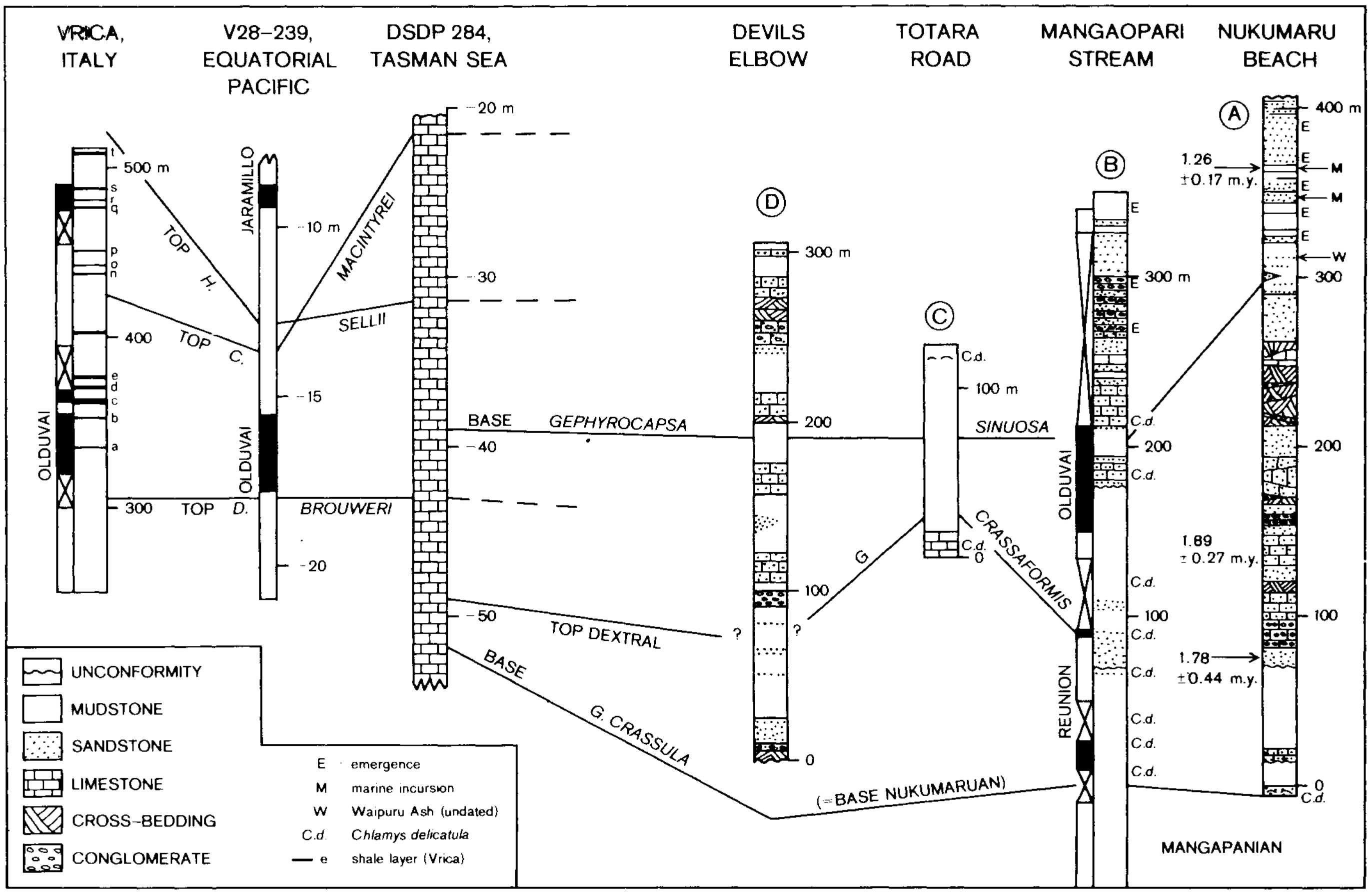

Figure 29.2. Primary stratigraphic correlations between the Nukumaruan strata of four New Zealand on-land sequences (Hornibrook, 1981; Beu and Edwards, 1984) and the Italian Vrica sequence (Backman et al., 1983; Tauxe et al., 1983); the short normal interval above bed *e* observed by Zijderveld et al. (1991) is not shown.

The upper bathyal sequence at DSDP site 284 is adapted from Kennett et al. (1975) and Hornibrook (1982), modified according to new data. The abyssal sequence of core V28–239 is adapted from Shackleton and Opdyke (1976) and Backman and Shackleton (1983).

means of recognizing the New Zealand stages beyond their stratotypes, although not without some difficulties.

There are few widely distributed biostratigraphic events that are reliable for correlation purposes. In upward stratigraphic order, the Nukumaruan bioevents are as follows:

1. First appearance of *Chlamys delicatula* (Hutton): traditionally accepted, but not without controversy, as indicating the base of the Nukumaruan in many sequences.
2. First appearance of *Globorotalia crassula* Cushman and Stewart: currently the most accurate means of recognizing the base of the Nukumaruan, but possibly unreliable in some shallow-marine sequences.
3. Last appearance of predominantly dextral populations of *Globorotalia crassaformis* (Galloway and Wissler).
4. First appearance of *Gephyrocapsa sinuosa* Hay and Beaudry: very useful in a wide variety of lithofacies.

The positions of these events in four representative New Zealand sequences are given in Figure 29.2. Other relevant bioevents are as follows:

5. First appearance of typical populations of *Globorotalia truncatulinoides* (d'Orbigny): useful in some sequences where it occurs between bioevents 3 and 4 mentioned earlier. This species gradually evolved from *G. tosaensis* Takayanagi and Saito during late Mangapanian and early Nukumaruan times. In subtropical-to-tropical environments, it first occurs just above the Gauss–Matuyama boundary (Barron et al., 1985).
6. Last appearance of *Discoaster brouweri* Tan Sin Hok: not useful for internal correlation purposes; inferred from a few very rare and sporadic occurrences to be between bioevents 3 and 4. This bioevent occurs not far below the Olduvai subchronozone in Pacific cores (Backman and Shackleton, 1983).
7. Last appearance of *Calcidiscus macintyrei* (Bukry and Bramlette): possibly a useful late Nukumaruan bioevent, but not yet well documented; definitely lies above bioevent 4 in section C. In Pacific cores it is recorded just above the Olduvai subchronozone (Backman and Shackleton, 1983), as well as in the Mediterra-

nean (Rio, Raffi, and Backman, Chapter 5, this volume).

8. Last appearance of *Helicosphaera sellii* (Bukry and Bramlette): a very useful but imprecisely located late Nukumaruan bioevent; definitely lies above bioevent 4 in sections A and C. In the deep sea it is found above event 7, close to the Jaramillo subchronozone (Gartner, 1977; Raffi, Rio, and Backman, Chapter 5, this volume).

For further information on these and numerous other bioevents in New Zealand, see Suggate et al. (1978), Hornibrook (1976, 1981), Beu, Grant-Taylor, and Hornibrook (1977), Hoskins (1982), and Beu and Edwards (1984).

The three radiometric dates shown against the Nukumaru Beach column in Figure 29.2 were obtained from elsewhere in the Wanganui Basin and positioned by lithostratigraphic correlation (Seward, 1976; Beu and Edwards, 1984). The lowest, from the Ohingaiti Ash, is a zircon fission-track date (Seward, 1979); the middle date is the K/Ar age of a pebble collected from a slightly higher (possible Nukumaru Limestone equivalent?) horizon (Mathews and Curtis, 1966); the highest, from the Mangahou Ash, is a glass-shard fission-track date (Seward, 1974).

Nukumaruan paleomagnetic observations have been made only at Mangaopari Stream (Kennett et al., 1971); the interpretation in Figure 29.2 is according to A. R. Edwards (in Hornibrook, 1981; Beu and Edwards, 1984). The combined paleontologic, paleomagnetic, and radiometric data are consistent with the Nukumaruan Stage commencing at about 2.4 Ma and ending at about 1.3 Ma (Beu and Edwards, 1984).

External correlations

Magnetostratigraphy appears to be a straightforward method to correlate between the Mangaopari Stream and Vrica sequences (Figure 29.2). The identification of the Olduvai subchron in New Zealand is still largely dependent on biostratigraphy (Beu and Edwards, 1984), as noted earlier.

In contrast to the magnetostratigraphic correlation, biostratigraphic correlations between New Zealand and Vrica are indirect. The first correlation is between New Zealand and the nearby upper bathyal DSDP site 284; the second is between DSDP site 284 and Vrica via the geographically intermediate abyssal core V28–239 (Figure 29.2). The change in the order of the LAD (last-appearance datum) of *C. macintyrei* and the LAD of *H. sellii* between core V28–239 and DSDP site 284 is not regarded as critical. The LAD of *H. sellii* probably is the less reliable of these two bioevents in the Pacific (Backman and Shackleton, 1983).

From that, the proposed Pliocene–Pleistocene boundary-stratotype (the base of the claystone resting on bed *e* in section B at Vrica) can be correlated to an undefined level within the upper part of the Nukumaruan and above the first occurrences of *G. sinuosa* in the New Zealand region, largely because of improvements in our understanding of the microbiostratigraphy of this interval, thanks to Backman and colleagues, as summarized by Rio, Raffi, and Backman (Chapter 5, this volume). Beu and Edwards (1984), using inferences of glacio-eustatic changes, have estimated where the boundary should lie in three of the New Zealand sequences shown in Figure 29.2. Their estimates are at about 30 m (Devils Elbow), 45 m (Mangaopari Stream), and 50 m (stratotype of the Nukumaruan Stage at Nukumaru Beach) above the first appearance of *G. sinuosa*; all three estimates fall at obvious lithologic changes.

Acknowledgments

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30 The Pliocene–Pleistocene boundary in continental sequences of North America

EVERETT H. LINDSAY

Introduction

Stratigraphic sections coeval with the Pliocene–Pleistocene boundary section at Vrica have been securely identified in two long terrestrial sequences in the southwestern USA: the San Pedro Valley sequence in Arizona and the Anza-Borrego Badlands sequence in southern California. Those sequences, with excellent chronologic resolution afforded by magnetostratigraphy, biostratigraphy, and isotope dating, provide a framework for interpreting the evolutionary and paleoclimatic changes in North America at the Pliocene–Pleistocene boundary, as well as for interpreting the sequence and effects of intercontinental dispersal events in the late Cenozoic.

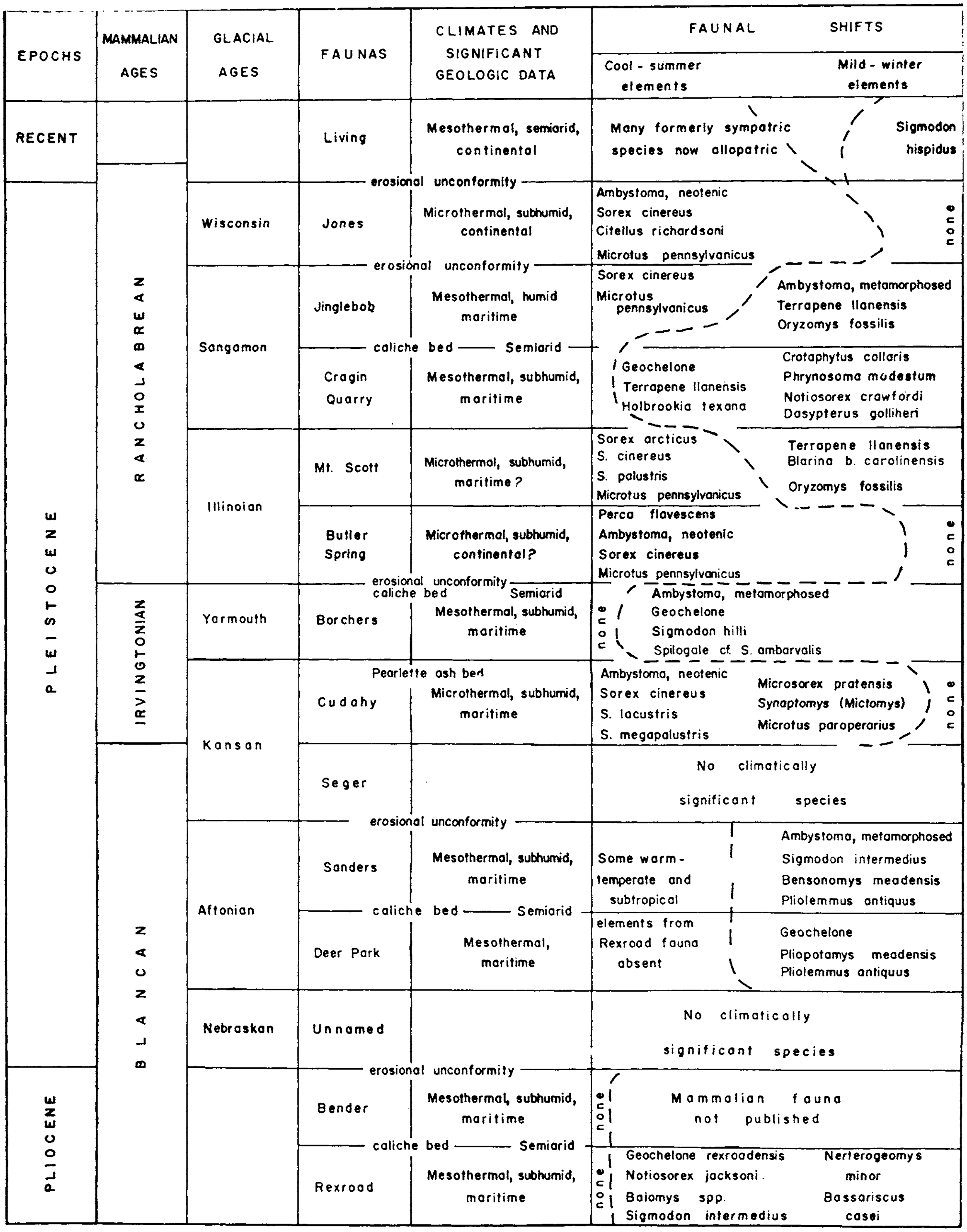
Glacial climate change and continental Plio–Pleistocene correlations

In the late 1930s, the so-called Wood Committee (Wood et al., 1941) assigned the Blancan Provincial Land Mammal Age to the late Pliocene and inferred that the Pliocene–Pleistocene boundary was located at the end of the Blancan. Although no formal Pleistocene land mammal age was proposed in the 1941 report, a list of Pleistocene mammals was given, for which the term "Rancholabrean" (characterized by the famous remains from the Rancho La Brea tar pits in Los Angeles) later came into general use. In 1951, D. E. Savage identified the Irvington fauna of the San Francisco Bay area as the type for a new "Irvingtonian" land mammal age, for early Pleistocene, post-Blancan and pre-Rancholabrean fossil mammal faunas. Thus, by the early 1950s, three late Cenozoic land mammal ages (Blancan, Irvingtonian, and Rancholabrean) had been identified in the late Cenozoic of North America, and the Pliocene–Pleistocene boundary (in North American vertebrate paleontology) was placed at the Blancan–Irvingtonian boundary. Nevertheless, the age and even the terms of recognition for the epoch boundary in North American contintental sequences remained controversial, because the criteria for its basic definition in the Old World were in conflict (Berggren and Van Couvering, 1979).

Despite controversies, improvements in the biostratigraphic and chronostratigraphic resolution of the late Cenozoic continental record continued, and in 1965 Hibbard and co-workers published a comprehensive summary of the North American late Cenozoic vertebrate faunal sequence, in which the mammal record was correlated with the four classic North American glacial ages (Nebraskan, Kansan, Illinoian, and Wisconsinan). In that report, the Pliocene–Pleistocene boundary was placed within, rather than at the end of, the Blancan, because late Blancan faunas appeared to represent a shift to colder climates (Figure 30.1). The chronologic-climatic framework for that report was based primarily on the faunal sequence in southwestern Kansas and northwestern Oklahoma, which was established by Hibbard and his students. A key part of that framework was the correlation of the distinctive and widespread Pearlette Ash, which was recognized in post-glacial sediments attributed to the retreat of the Kansan ice sheet. The warm-adapted Cudahy fauna, from below the Pearlette Ash, was held to be an early Irvingtonian assemblage and was correlated with late Kansan warming trends. The cold-adapted Borchers fauna, from above the Pearlette Ash, was interpreted as late Irvingtonian and was correlated with the end of the Yarmouth interglacial.

Work since 1965 (e.g., Izett, 1981) has shown that the "Pearlette Ash" above the Cudahy fauna is a volcanic unit different from the "Pearlette Ash" below the Borchers fauna, with a radiometric age more than 1 m.y. younger, and that the Borchers fauna is older (not younger) than the Cudahy fauna. In fact, three "Pearlette Ash" units are now identified, based on stratigraphy near the source area in Yellowstone Park, Wyoming: the Lava Creek *b* ash (equivalent to Pearlette type O), dated to 0.6 Ma; the Mesa Falls ash (Pearlette type S), dated to 1.3 Ma; and the Huckleberry Ridge ash (Pearlette type B), dated to 2.0 Ma (Izett, 1981). The Lava Creek *b* ash overlies the Cudahy fauna, and the Huckleberry Ridge ash underlies the Borchers fauna. The Cudahy fauna is now considered late Irvingtonian, and the Borchers fauna is considered latest Blancan or earliest Irvingtonian.

At the time, Hibbard suspected that the 1965 framework might be wrong (Zakrzewski, 1975, p. 123), and by 1972 he openly questioned his 1965 correlations with the glacial sequence (Skinner and Hibbard, 1972, pp. 131–136). For one thing, Izett et al. (1970) had already noted the occurrence of two superposed "Pearlette ashes" at Cerro Summit, Colorado, and for another, Skinner and Hibbard (1972, p. 136, fig. 60) had come to recognize



EPOCHS	MAMMALIAN AGES	GLACIAL AGES	FAUNAS	CLIMATES AND SIGNIFICANT GEOLOGIC DATA	FAUNAL SHIFTS: Cool - summer elements	FAUNAL SHIFTS: Mild - winter elements
RECENT			Living	Mesothermal, semiarid, continental	Many formerly sympatric species now allopatric	Sigmodon hispidus
				erosional unconformity		
PLEISTOCENE	RANCHOLABREAN	Wisconsin	Jones	Microthermal, subhumid, continental	Ambystoma, neotenic Sorex cinereus Citellus richardsoni Microtus pennsylvanicus	none
				erosional unconformity		
		Sangamon	Jinglebob	Mesothermal, humid maritime	Sorex cinereus Microtus pennsylvanicus	Ambystoma, metamorphosed Terrapene llanensis Oryzomys fossilis
				caliche bed — Semiarid		
			Cragin Quarry	Mesothermal, subhumid, maritime		Geochelone Terrapene llanensis Holbrookia texana Crotaphytus collaris Phrynosoma modestum Notiosorex crawfordi Dasypterus golliheri
		Illinoian	Mt. Scott	Microthermal, subhumid, maritime ?	Sorex arcticus S. cinereus S. palustris Microtus pennsylvanicus	Terrapene llanensis Blarina b. carolinensis Oryzomys fossilis
			Butler Spring	Microthermal, subhumid, continental?	Perca flavescens Ambystoma, neotenic Sorex cinereus Microtus pennsylvanicus	none
				erosional unconformity caliche bed Semiarid		
	IRVINGTONIAN	Yarmouth	Borchers	Mesothermal, subhumid, maritime	none	Ambystoma, metamorphosed Geochelone Sigmodon hilli Spilogale cf. S. ambarvalis
				Pearlette ash bed		
		Kansan	Cudahy	Microthermal, subhumid, maritime	Ambystoma, neotenic Sorex cinereus S. lacustris S. megapalustris Microsorex pratensis Synaptomys (Mictomys) Microtus paroperarius	none
	BLANCAN		Seger		No climatically significant species	
				erosional unconformity		
		Aftonian	Sanders	Mesothermal, subhumid, maritime	Some warm-temperate and subtropical elements from Rexroad fauna absent	Ambystoma, metamorphosed Sigmodon intermedius Bensonomys meadensis Pliolemmus antiquus
				caliche bed — Semiarid		
			Deer Park	Mesothermal, maritime		Geochelone Pliopotamys meadensis Pliolemmus antiquus
		Nebraskan	Unnamed		No climatically significant species	
				erosional unconformity		
PLIOCENE			Bender	Mesothermal, subhumid, maritime	none	Mammalian fauna not published
				caliche bed — Semiarid		
			Rexroad	Mesothermal, subhumid, maritime	none	Geochelone rexroadensis Notiosorex jacksoni Baiomys spp. Sigmodon intermedius Nerterogeomys minor Bassariscus casei

Figure 30.1. Late Cenozoic faunal shifts and inferred climatic changes in the southern Great Plains. Note the Pearlette Ash bed between the Cudahy and Borchers faunas. (From Hibbard et al., 1965, figure 2, with permission of Princeton University Press.)

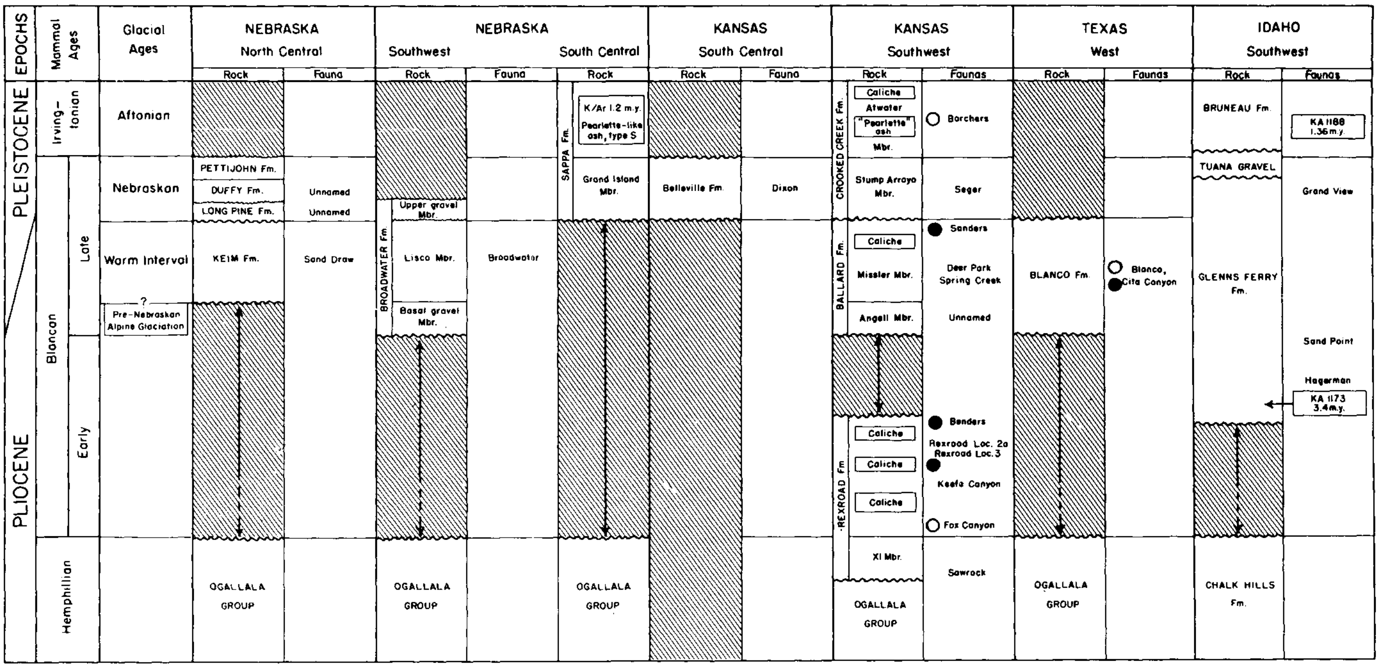


Figure 30.2. Correlation of Blancan deposits and faunas in the Great Plains region and Idaho, according to Skinner and Hibbard (1972, figure 6). Note that the Sand Draw and Broadwater faunas of Nebraska are correlated to a warm interval near the Pliocene–Pleistocene boundary. Radiometric age calibration according to Berggren et al. (1985).

that there were in fact diverse, warm-adapted late Blancan faunas (e.g., Sand Draw in northeastern Nebraska) at relatively high latitudes. That was sufficient evidence for those authors to conclude that most, if not all, Blancan faunas in the Great Plains area were older than the "first glacial climates" (Figure 30.2).

Unfortunately, Hibbard and most other students of North American vertebrate paleontology, up to and including Kurtén and Anderson (1980), were locked into a rigid paleoclimatic concept in which the earliest evidence of any climatic deterioration was considered, by definition, as Nebraskan and, by definition, the beginning of the Pleistocene. As an example, the Broadwater fauna of southwestern Nebraska, a late Blancan warm-adapted fauna very similar to the Sand Draw fauna of Skinner and Hibbard (1972), was considered to be of early Pleistocene age by Schultz, Lueninghoener, and Frankforter (1951), in part because a gravel bed interpreted as a Nebraskan till underlies the strata yielding the fauna.

I heartily endorse the conclusions of Harland et al. (1982, p. 43) regarding the unreliability of Pleistocene glacial sequences: "Attempts to make a stratigraphic sequence out of successive cycles of continental glaciation have proved to be totally inadequate . . . and . . . continental classifications have been shown to be oversimplified and incomplete, based to a large extent on erroneous concepts." Oxygen-isotope changes in oceanic plankton of deep-sea cores represent at least 23 globally significant intervals of cooling during the past 1 m.y. (Shackleton and Opdyke, 1973, 1976; Harland et al., 1982; Thunell and Williams, 1983). Correlation of North American continental glaciation with oceanic cooling intervals is still uncertain in detail; to maintain integrity and accuracy, correlations of North American Quaternary land mammal faunas must be based on superposition, morphologic evolution of common taxa, isotope dating, and magnetic-polarity sequences, *not on climatic inferences*.

Present state of North American Plio–Pleistocene correlations

Among the relatively recent attempts to correlate the North American Quaternary mammal faunas are the following: those of Repenning (1979, 1987; Repenning, Fejfar, and Heinrich, 1990; Repenning and Brouwers, 1992), based on the evolution and dispersal of microtine rodents; that of Schultz, Martin, Tanner, and Corner (1978), based on the grouping of faunas according to taxonomic similarity; and that of Johnson, Opdyke, and Lindsay (1975), based on magnetostratigraphic correlations tied to detailed biostratigraphy in the southwestern USA. It should be pointed out that those and most other studies of North American Quaternary mammals (e.g., Kurtén and Anderson, 1980) have included Blancan as well as Irvingtonian and Rancholabrean land mammal ages, since at least part of the Blancan has frequently been considered Pleistocene.

Microtine rodent event-stratigraphy

Repenning (1979) divided the Blancan into five sequential faunal intervals, and the Irvingtonian and Rancholabrean into two intervals each, characterized by the appearances of immigrant species of microtine rodents closely related to European faunas, or characterized by newly evolved species. In a follow-up, Repenning (1987) expanded on that study with some slight revisions (Figure 30.3). His chronologic sequence was based on a

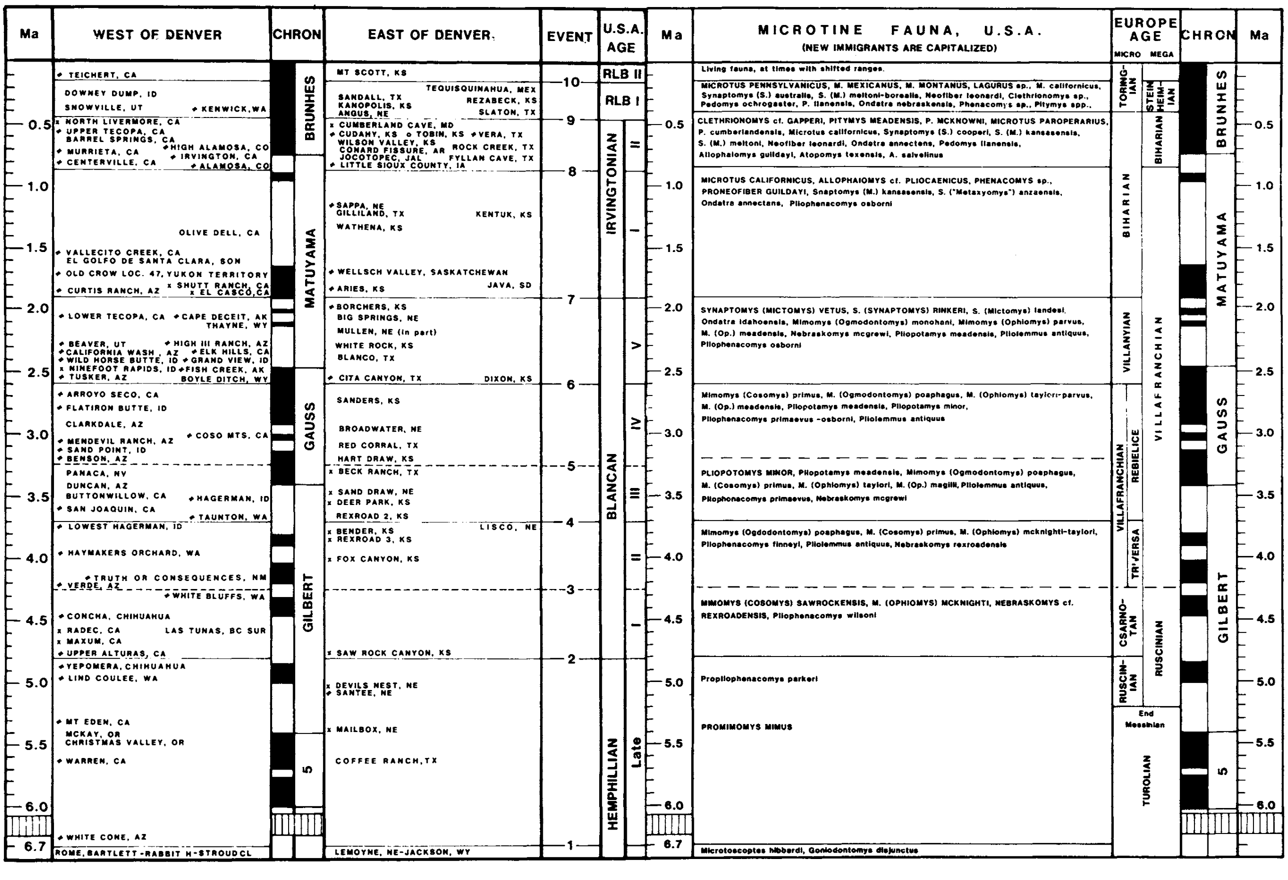


Figure 30.3. Correlation of microtine dispersal events in North America to subdivisions of the land mammal ages. (From Repenning, 1987, figure 1, courtesy of U.S. Geological Survey.)

very comprehensive study of this widespread and abundantly fossiliferous group and, in collaboration with O. Fejfar and others, on correlations with European faunas (Repenning et al., 1990). Repenning's division of Quaternary faunas is a great leap forward toward the chronologic resolution of North American Pleistocene mammal faunas, although I regard the separation of the early Blancan into five parts to be excessive, when three are all that may reasonably be called for. Also, Repenning's dating of Blancan faunas, based on correlations to Europe, conflicts with the direct evidence of local age determinations (Lindsay, Johnson, and Opdyke, 1975; Lindsay, Opdyke, and Johnson, 1984), which date the base of the Blancan to about 4.3 Ma, rather than 4.8 Ma.

Faunal-assemblage subdivisions

Schultz et al. (1978) provided names for the faunal assemblages of Plio–Pleistocene age that are, in effect, subdivisions of the conventional Blancan and Irvingtonian mammal ages (Figure 30.4). Early Blancan faunas were grouped into the Rexroadian sub-age (considered pre-Blancan by Kurtén, 1972), and later Blancan faunas into the Senecan sub-age. Early Irvingtonian faunas were grouped into the Sappan sub-age, and later Irvingtonian faunas into the Sheridanian sub-age; the Cudahy, Port Kennedy, and Conard Fissure faunas were left, however, as an unnamed middle Irvingtonian group. Rancholabrean faunas were not discussed, although early and late Wisconsinan intervals, as well as a Sangamon interglacial division, were illustrated as Rancholabrean (Schultz et al., 1978, fig. 1). In this chapter, species differences (*Smilodon californicus* vs. *S. fatalis*, and *Castoroides kansensis* vs. *C. nebraskensis*) separate Rancholabrean from Irvingtonian faunas. The appearances of modern species of microtines, and the earliest certain record of *Bison*, distinguish Sheridanian from Sappan faunas, and the appearances of the vole *Allophaiomys* and the lemming *Synaptomys* (*Mictomys*) distinguish Sappan from Senecan faunas. Schultz et al. (1978) also questioned whether or not *Mammuthus* appeared at the beginning of the Irvingtonian (Sappan).

Schultz et al. (1978) vacillated on placing the Pliocene–Pleistocene boundary between the Blancan and Irvingtonian, or between the egregious Kimballian (= early Hemphillian) (Lindsay et al., 1984) and Blancan. Schultz and Hillerud (1978) reiterated the Great Plains Pleistocene mammal sequence of Schultz et al. (1951), based on the concept of a five-terrace sequence cut into and built on Pliocene rocks (Figure 30.5). The oldest terrace (Terrace 5) includes the late Blancan Broadwater formation and fauna (noted earlier), underlain by a gravel interpreted as a Nebraskan glacial till; that was given an inferred age of 3.2 Ma, which may or may not be the same age as the beginning of the trend toward Pleistocene climates noted at about that time in mid-Pliocene marine sequences (Thunell and Williams, 1983; Shackleton et al., 1984, 1985; Shackleton, Hall, and Pate, 1995). A critical examination of Schultz and Hillerud (1978) (Figure 30.5) suggests that terraces 1, 2, and 3 are Holocene. All of the significant Pleistocene climatic events must therefore be recorded in Terrace 4.

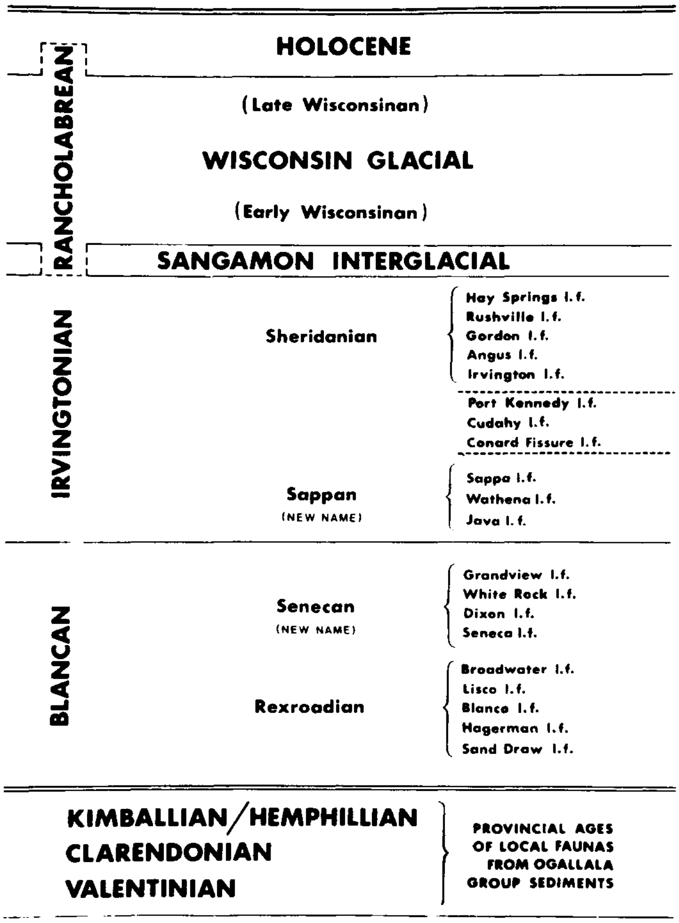


Figure 30.4. Provincial land mammal age subdivisions for the North American Quaternary and late Tertiary proposed by Schultz et al. (1978, figure 1). (Courtesy of Nebraska Academy of Sciences.)

Magnetostratigraphically controlled biochronology

Magnetostratigraphy, in combination with biostratigraphy, in thick, continuous sequences offers a third approach to the analysis of North American Pleistocene terrestrial deposits and faunas.

San Pedro Valley, Arizona

In the San Pedro Valley sequence of Arizona, the magnetostratigraphy published by Johnson et al. (1975) identified 11 reversal events in 150 m of strata in the St. David Formation that could be correlated with the interval from late Gilbert to Brunhes in the magnetic-polarity time scale (Figure 30.6). Also, four biochronologically significant faunal datum events were identified in the biostratigraphy of the vertebrate fossil sites. The first three of these events, the *Sigmodon medius* LSD (lowest

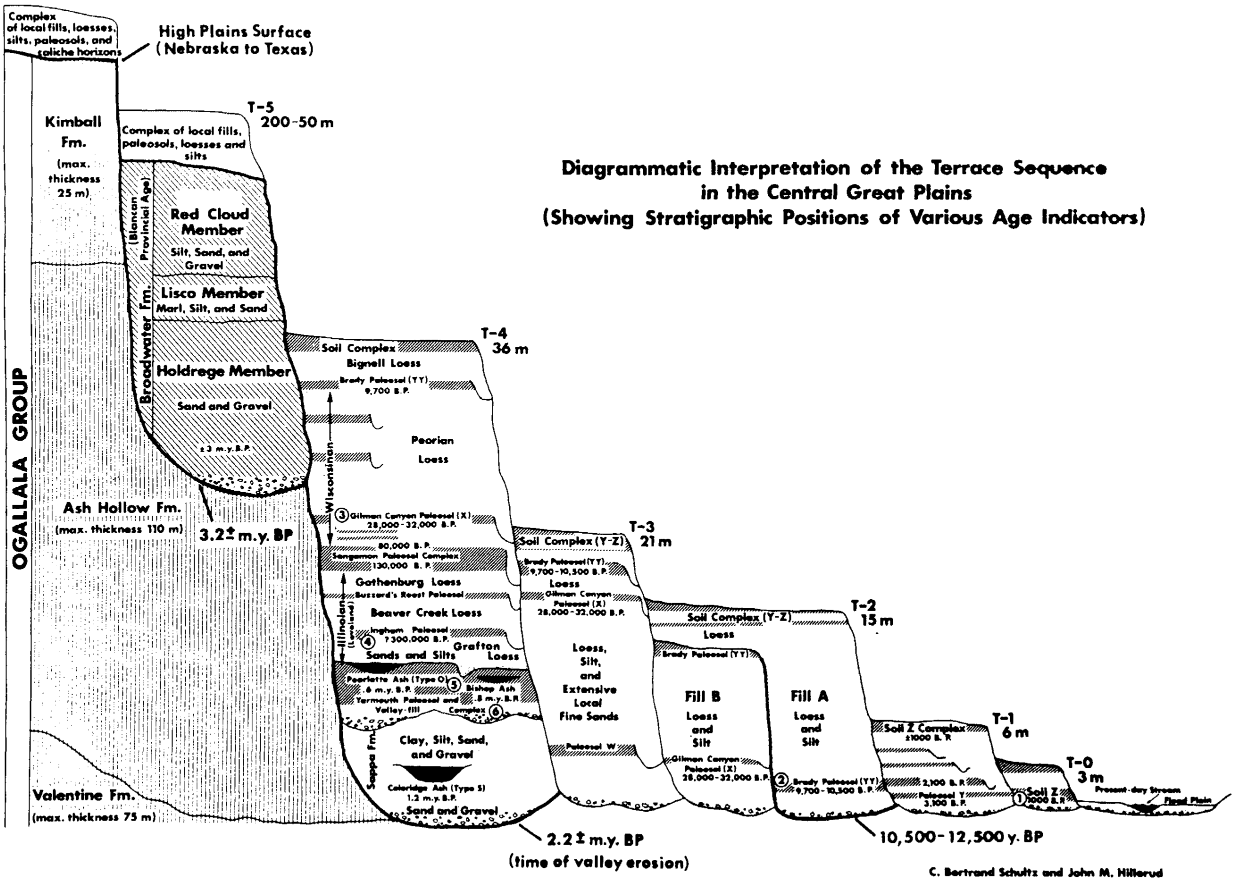


Figure 30.5. Diagrammatic interpretation of the sequence of post-Kimballian terrace fills developed in the Great Plains of Nebraska. (From Schultz and Hillerud, 1978, figure 1, courtesy of Nebraska Academy of Sciences.)

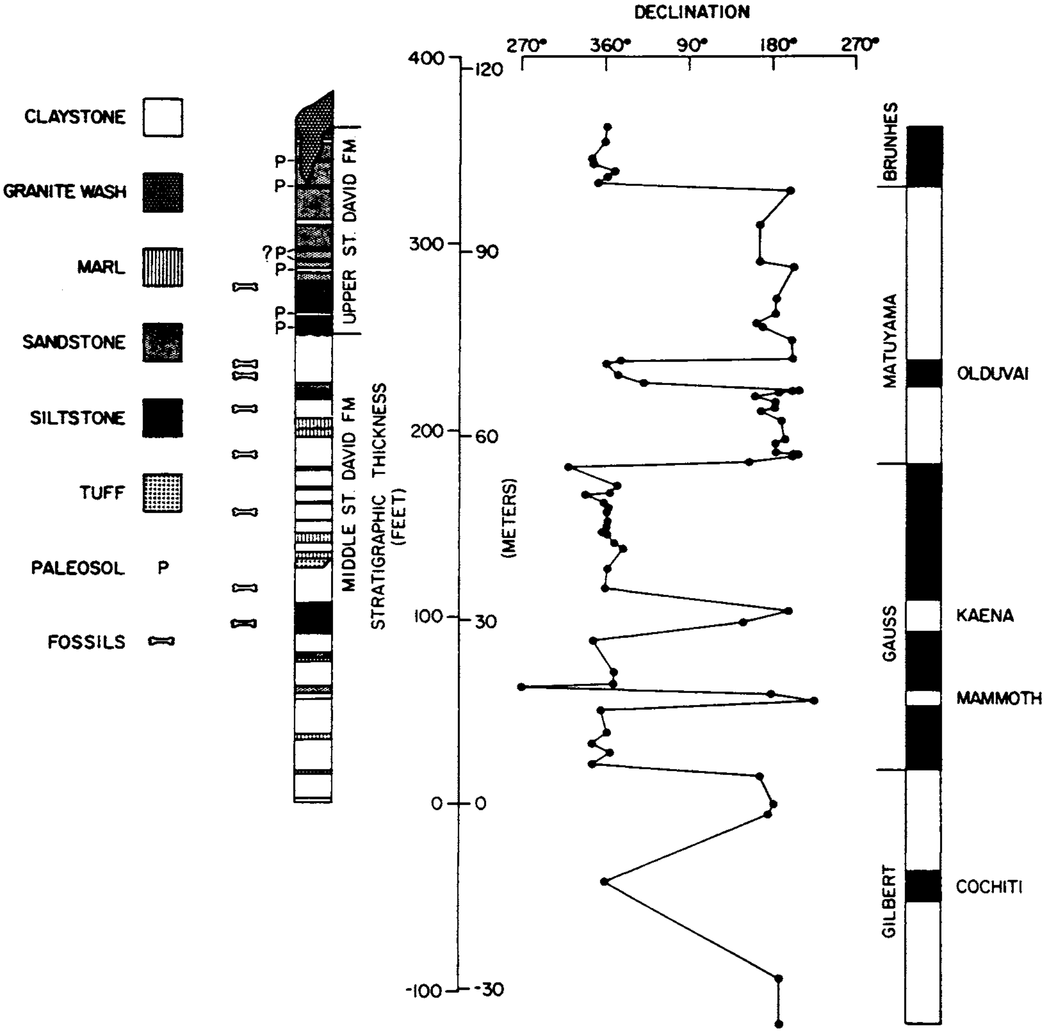


Figure 30.6. Variation of magnetic declination with stratigraphic level at Curtis Ranch. In this and other magnetic-polarity columns, the black sections represent normally magnetized strata, and the white sections represent reversely magnetized strata. Eight fossil levels in the San Pedro Valley sequence are indicated by bone symbols to the left of the lithologic column. These fossil levels, named in order of increasing age, are Prospect, *Glyptotherium* (of Gidley), Gidley level, Johnson Pocket, Cal Tech, Horsey Green bed, Honey's Hummock, and Bonanza. (From Johnson et al., 1975, with permission of the Geological Society of America.)

stratigraphic datum) at the base of the Mammoth subchronozone of the Gilbert chronozone, the *Nannippus* HSD (highest stratigraphic datum) at the base of the Matuyama chronozone, and the *Ondatra idahoensis* LSD near the Réunion subchronozone of the lower Matuyama, are associated with the Blancan land mammal age. The youngest datum event, the *Lepus* LSD at the base of the Olduvai subchronozone, has been associated with the Curtis Ranch fauna of earliest Irvingtonian land mammal age. Unfortunately, the taxon previously recorded as *Lepus* in the Curtis Ranch fauna is now identified as a large species of another rabbit, *Sylvilagus* (White, 1991). A smaller species of *Sylvilagus* appears stratigraphically lower in the San Pedro Valley strata in the California Wash fauna, associated with the *Ondatra idahoensis* datum event (Lindsay et al., 1990). Perhaps a more significant biochronologic event is the replacement of archaeolagine rabbits (e.g., *Hypolagus* and *Pewelagus*) by leporine rabbits (e.g., *Sylvilagus* and *Lepus*) in the late Blancan (White, 1987). It appears the San Pedro Valley faunal sequence documents the transition from the Blancan to the Irvingtonian land mammal age, in which the Curtis Ranch fauna represents the latest Blancan or earliest Irvingtonian faunal interval. The Curtis Ranch fauna is securely placed at the base of the Olduvai subchron and therefore (according to the Vrica calibration) represents the latest Pliocene fauna in that sequence.

Anza-Borrego, California

In the Anza-Borrego sequence in southern California, Opdyke et al. (1977) identified 11 magnetic reversals through 4,000 m of strata in the Palm Springs and upper Imperial formations (Figures 30.7 and 30.8). The Anza-Borrego magnetic sequence is almost a duplicate of the San Pedro Valley magnetic sequence, although with a much higher average sedimentation rate (1.2 m/k.y. vs. 0.04 m/k.y.) (Figures 30.6 and 30.9). Johnson et al. (1983) extended the Anza-Borrego paleomagnetic sequence

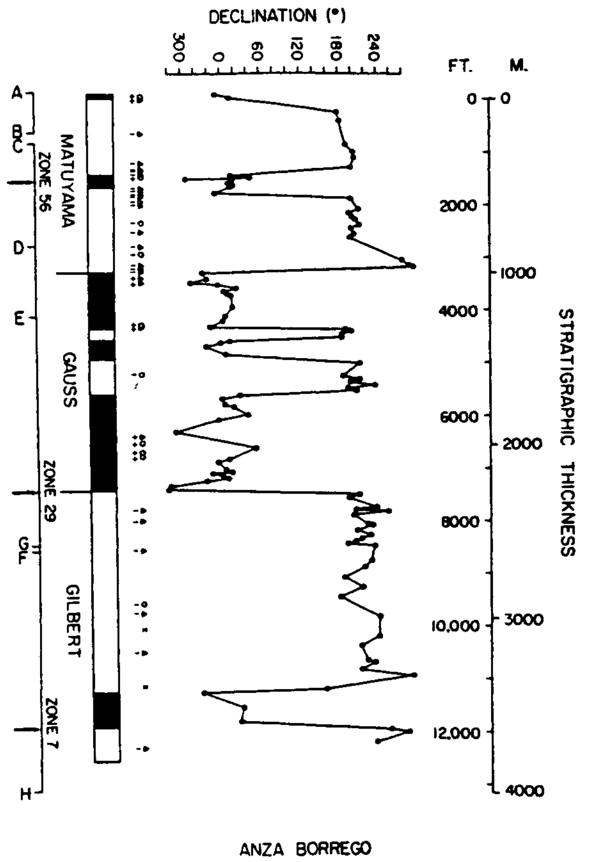


Figure 30.7. Variation of magnetic declination with stratigraphic level in the Anza-Borrego Badlands, California. (From Opdyke et al., 1977, with permission of the University of Washington Press.)

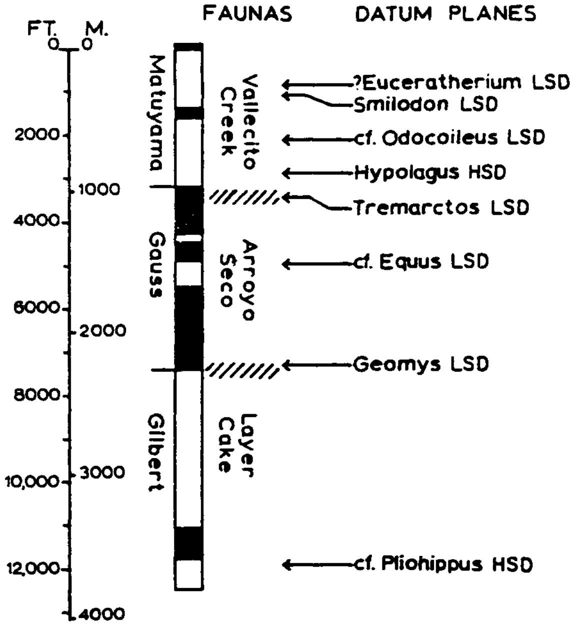


Figure 30.8. Mammal datum planes in the Anza-Borrego Badlands sequence. (From Opdyke et al., 1977, with permission of the University of Washington Press.)

down through the Imperial Formation, after removing a magnetic overprint in the samples by thermal cleaning. They were able to identify 17 magnetic reversals (including the Nunivak subchron of the Gilbert magnetic chron), and with that control they were able to demonstrate a change in the sedimentation rate from 5.5 m/k.y. to 0.5 m/k.y. in an interval of 3.4 m.y. during deposition of a deltaic complex in the Anza-Borrego area (Figure 30.9).

Vertebrate fossils in the Anza-Borrego Badlands have been assigned to Blancan and Irvingtonian land mammal ages (Downs and White, 1968). Eight local datum events were identified in the Anza-Borrego faunal sequence by Opdyke et al. (1977) and dated according to paleomagnetic interpolation (Figure 30.9). The lower six datum events (cf. *Pliohippus* HSD, *Geomys* LSD, cf. *Equus* LSD, *Tremarctos* LSD in the Gilbert chronozone, and the HSD of *Hypolagus* and LSD of cf. *Odocoileus* in the lower Matuyama chronozone) are associated with Blancan land mammal age, while the two highest faunal datum events (*Smilodon* LSD and ?*Euceratherium* LSD near the top of the Olduvai subchronozone in the upper Matuyama) are associated with Irvingtonian land mammal age.

Blancan–Irvingtonian boundary

The transition from the Blancan to the Irvingtonian land mammal age is identified by the appearance of *Ondatra* and perhaps *Sylvilagus* in the San Pedro Valley sequence and by the appearance of cf. *Odocoileus* and *Smilodon* in the Anza-Borrego sequence. Those interpretations suggest that the Blancan–Irvingtonian boundary should be placed below but near to the base of the Olduvai subchronozone, at approximately 2.1 Ma in the orbitally tuned calibration. Thus, the Blancan–Irvingtonian boundary is slightly older than the Pliocene–Pleistocene boundary, as identified at Vrica within the uppermost part of the Olduvai (Pasini and Colalongo, Chapter 2, this volume).

The beginning of the Irvingtonian land mammal age is best characterized by the appearance of Eurasian immigrants, especially *Mammuthus*. Perhaps the earliest North American record of *Mammuthus* is from the Wellsch Valley fauna, which has been correlated to a level within the Olduvai magnetic subchron. To date, *Mammuthus* has not been recorded in the Curtis Ranch sequence in the San Pedro Valley nor in the paleomagnetically dated sequence of the Anza-Borrego Badlands. The current consensus of North American vertebrate paleontologists (cf. Lundelius et al., 1987) is to consider the Blancan–Irvingtonian boundary to be coincident with the base of the Olduvai subchron. This has the effect of placing the Olduvai-age Wellsch Valley and Curtis Ranch faunas, probably the earliest known Irvingtonian faunas, in the latest part of the Pliocene, just below the level of the Vrica definition.

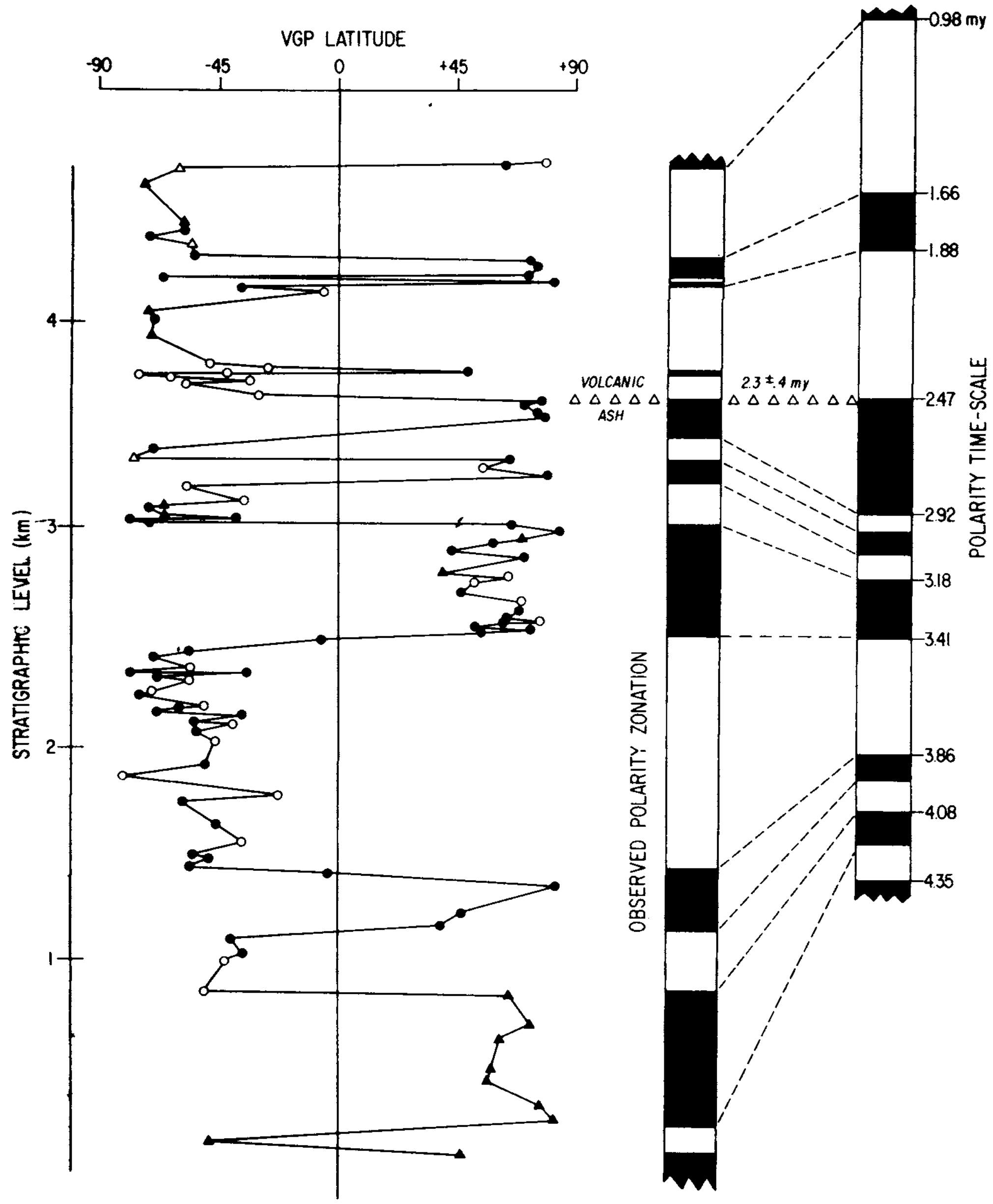


Figure 30.9. Extended Anza-Borrego Badlands sequence, with calibrated magnetic-polarity time scale of Mankinen and Dalrymple (1979) on the right. (From Johnson et al., 1983, with permission of the Geological Society of America.)

Sequence of late Cenozoic mammal faunas in North America

Additional paleomagnetically dated sequences from other areas of the western USA have contributed to a rigorous interpretation of late Cenozoic faunal changes in North America. Those additional paleomagnetic sequences include Meade County, Kansas (Lindsay et al., 1975), and the Snake River plain of Idaho (Neville et al., 1979).

Lindsay et al. (1984) completed a summary of North American late Cenozoic mammal faunas that includes the faunas mentioned earlier, as well as the Chamita section of New Mexico (MacFadden, 1977), the Verde section of Arizona (Bressler and Butler, 1978), the Wikieup section of Arizona (MacFadden, Johnson, and Opdyke, 1979), and several previously unpublished sections. Figures 30.10 and 30.11, taken from that study, illustrate the correlations and sequences of those faunas, which extend the correlation of North American faunas into the late Miocene and Hemphillian land mammal age. The Hemphillian land mammal age includes the interval from about 4.5 Ma to 8 Ma, correlative with magnetic chrons 5, 6, and 7, and half of chron 4 (Gilbert). The Blancan land mammal age includes the interval between about 1.9 Ma and 4.5 Ma, correlative with the lower Matuyama, the Gauss, and the upper half of the Gilbert magnetic chrons. The Blancan land mammal age thus includes most of the Pliocene, although the early Pliocene extends into late Hemphillian age, and the latest Pliocene is equivalent to the basal Irvingtonian.

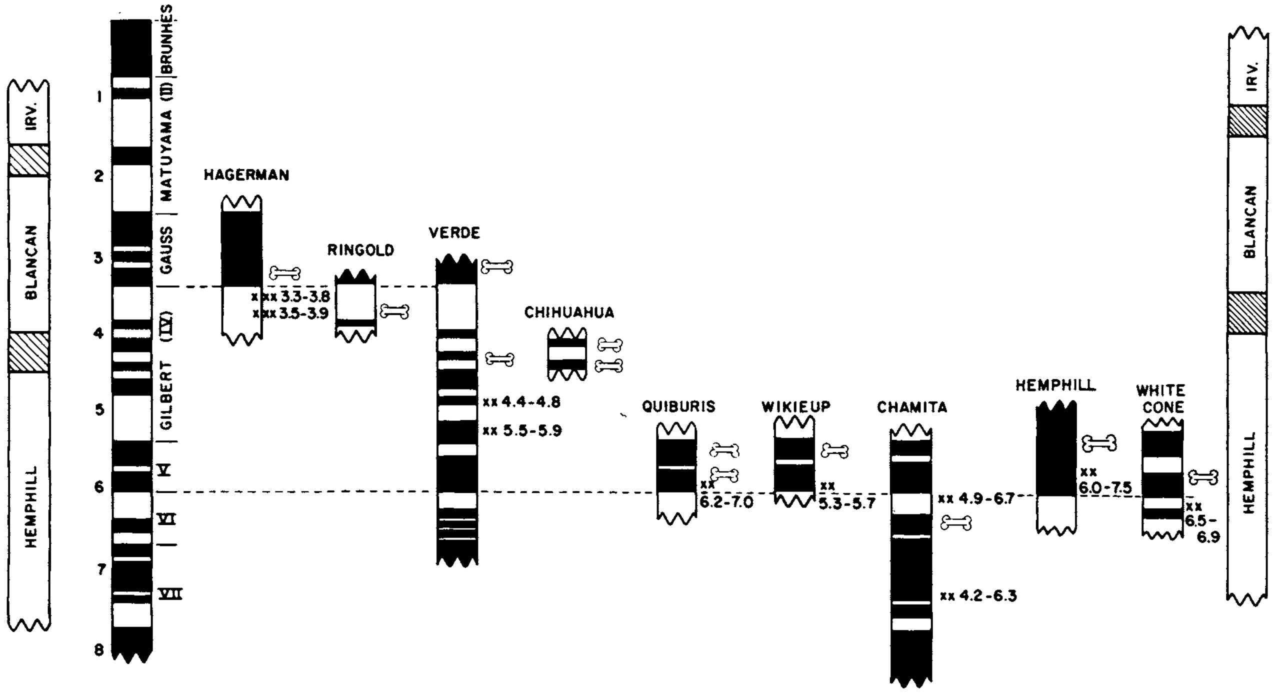


Figure 30.10. Correlation of Blancan and Hemphillian fossil sequences. Magnetic-polarity time scale on the left; positions and error limits of isotopically dated ashes are shown by the symbol XX. Bone symbols indicate faunal levels. Hagerman section from Neville et al. (1979), with Gidley Horse Quarry level; Ringold section from Lindsay et al. (1984), with level of *Ophiomys mcknighti*; Verde section from Bressler and Butler (1978), with level of Clarkdale fauna (above) and Verde fauna (below); Chihuahua section from Lindsay et al. (1984), with level of Concha fauna (above) and Yepómera fauna (below); Quiburis section from Lindsay et al. (1984), with position of Camel Canyon fauna indicated by lower bone symbol; Wikieup section from MacFadden et al. (1979), with level of Wikieup fauna; Chamita section from MacFadden (1977), with level of Rak Camel Quarry; Hemphill section from Lindsay et al. (1984), with level of White Cone fauna.

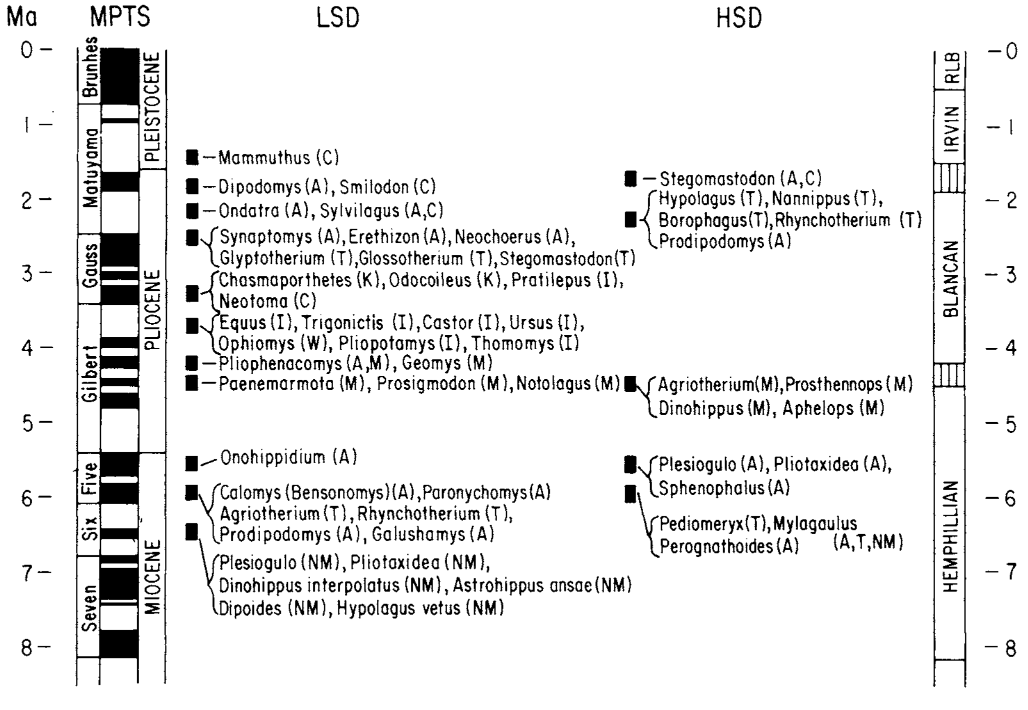


Figure 30.11. North American late Cenozoic mammal-datum events, based on data summarized by Lindsay et al. (1975, 1984).

Chronostratigraphic ordering of these faunas, based on magnetostratigraphy, biostratigraphy, and radiometric dating, provides a framework for interpreting faunal and climatic changes in North America during the late Cenozoic. It also allows correlation with similar faunal changes in other parts of the world and facilitates identification of the timing and sequence of intercontinental dispersal events by land mammals.

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Index