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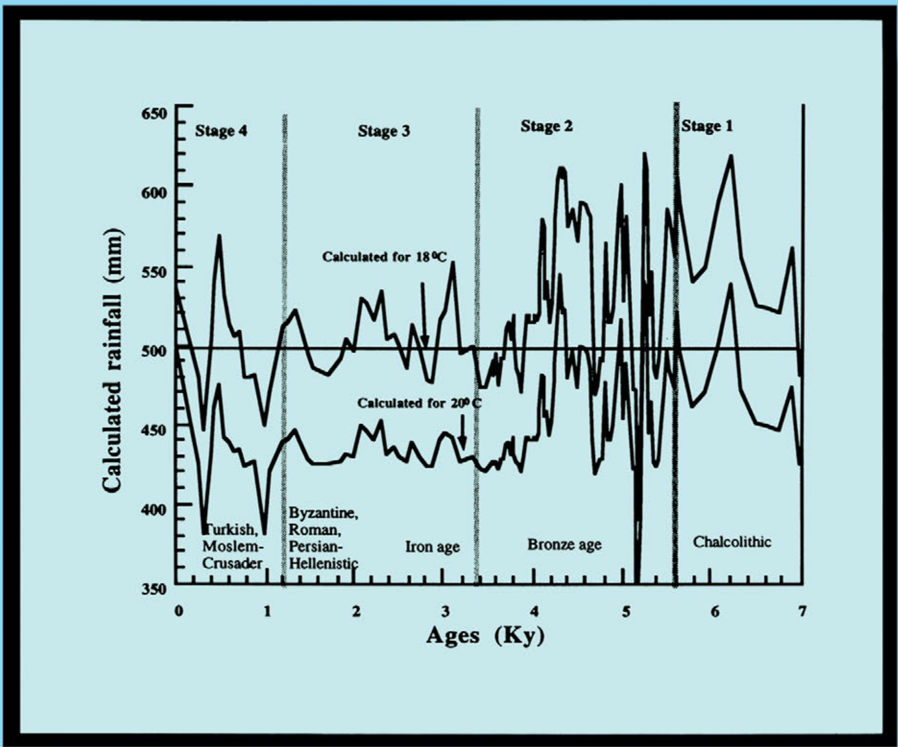
WATER, ENVIRONMENT AND SOCIETY IN TIMES OF CLIMATIC CHANGE

edited by

Arie S. Issar and Neville Brown



WATER, ENVIRONMENT AND SOCIETY IN TIMES OF CLIMATIC CHANGE

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WATER, ENVIRONMENT AND SOCIETY IN TIMES OF CLIMATIC CHANGE

Contributions from an International Workshop within the framework of International Hydrological Program (IHP) UNESCO, held at Ben-Gurion University, Sede Boker, Israel from 7–12 July 1996

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Introduction

Since the greenhouse effect emerged as a predictable threat, necessitating the evaluation of its future impact on the environment in the various parts of the globe, interest in the climate changes during the Holocene has gained momentum. The background can be summarized by the sentence: The past is a key to the future. As a matter of fact, this sentence is in the opposite direction, on the dimension of time, to the principle adopted by the founders of the science of geology. They proposed that geological processes in the present should be used as a key for understanding the past.

Another reason for the interest in the history of the climate of the Holocene can be described as the renaissance of a modified deterministic approach to the interrelation between physical and human geography. This relates in the first place to the fact that various investigations, especially as carried out by Hubert Lamb, showed that the sequence of climate changes previously suggested by Blytt and Sernander for Europe and adopted by most Holocene climatologists was far too general, and that there were more climate changes during recent history than previously taken account of. In the second place it was found out that these changes had had an impact on the history of human communities. Thus, one can conclude that once the taboo on geographical determinism (*i.e.* causal linkage between climate changes and history) was lifted, scientists have gained once more a tool to investigate the recent past with regard to the interaction between the physical and the human environment including socio-economic systems, this with a view to forecasting the future. At present this interaction works in two directions. The physical environment influences the human socio-economic systems while society itself seems to be influencing the climate.

But although one can point out the positive aspects of investigating the causal relations between climate and history in the past, one has to be careful not to become an extreme determinist. Therefore one should try to modify the determinist paradigm into a state which can be called neo-determinism. Assuming this approach is adopted, a word of caution has still to be said with regard to the danger of circular argument. This can happen when historical change is taken first as a consequence of climate change and later as a proof of it. Caution has also to be practiced when one investigates time series of environmental data, which may be influenced by human society, like changes in pollen assemblages and rates of erosion of soils. These pitfalls can be avoided by starting with time series of proxy data, which cannot be in anyway anthropogenic. For this purpose environmental isotopes, mainly $^{18}O/^{16}O$ ratios and to a certain extent $^{13}C/^{12}C$ ratios, are optimal. These have to be correlated with other proxy data as well as with archaeological and historical data. The results can then be compared with those supplied by the climate models. However, such modeling without field data to calibrate the results is itself extremely problematic when it comes to decisions about mitigating strategy and resource allocation. One conclusion is that the effort to understand the past in order to forecast the future has to be thoroughly interdisciplinary.

The workshop which was convened at Sede Boker as part of this global effort was organized with the interdisciplinary requirement firmly in mind.

The reasons for climate changes during the Holocene are discussed in the paper by Robert and Reid Bryson. These authors maintain that the main cause for climate changes during the Holocene was volcanic eruptions, which loaded the atmosphere with aerosols. These affect atmospheric optical depth and thus modulate insolation. The authors calculate the rate of loading, and its impact on insolation and optical depth as the basis for an index of radiocarbon dated volcanism. This they use to predict ice volume, glacial area and albedo, the values of which are combined to model seasonal hemispheric mean temperatures. By a series of calculations they arrive at time series for atmospheric circulation features at different locations. These they calibrate using modern relationships between monthly precipitation averages and circulation features. They then move back through time to estimate past precipitation.

A rather good correlation between calculated and proxy data series was demonstrated for a few test cases. This seems to confirm the basic assumption of the authors that volcanic activity is a major forcing mechanism in climate change during the Holocene.

On the other hand Bas van Geel and Hans Renssen in their paper suggest that the *causum casorum* for a cold period which occurred between *ca.* 850 and 760 calendar years BC, (*ca.* 2,750–2,450 BP on the radiocarbon time scale) was reduced solar activity. This can be evidenced by higher levels of ^{14}C in the atmosphere, which cause divergence between the dates indicated by ^{14}C and those calculated by dendrochronological methods. The reduced solar activity, which means reduced ultraviolet radiation, may also lead to a decline in ozone production in the lower stratosphere. This could have resulted in a curbing of the latitudinal extension of the Hadley Cell circulation, which may lead to a weakening of the monsoons as well as to an expansion of the Polar Cells and the displacement towards the south of the cyclonic tracks. The authors present evidence from all over the world though especially Europe that indeed a cold climate event occurred; and that it had a strong still traceable impact on the environment and society in certain regions.

In Scott Stine's paper the genesis for a Medieval cold period from 850 to 1350 AD is related to shifts in the circumpolar circulation. During the first century of this interval many lakes in west of the U.S.A. dried up. Stine suggests that this regional dry spell was due to a global cold phase, as it correlates with the glacial advance *e.g.* in Alaska and Canada. More complicated changes abruptly began within a half century of 1130 AD. Conditions changed radically again after 1325 when lake levels rose again in this region.

Neville Brown looks at the impact of climate change on history from what he sees as the high point of the Roman Empire to the end of the first millennium. He works from the precept that the influence of climate change cannot be assessed except in relation to the other factors that bear upon historical development. He is therefore bringing general history into historical climatology more comprehensively than anybody has previously done.

He amplifies long-established notions about the influence that climate change had on certain major episodes. One case in point is the extent to which barbarian pressures on the later Roman Empire were modulated by drought cycles in Inner Eurasia. Another is the part that aridity abnormal even by Arabian standards played in the initial emergence of Islam. A third concerns the maritime and continental expansion of the Vikings. In this last case he raises a new issue. Why were the Vikings so reluctant to settle northwards? He proffers an explanation couched in terms of climate becoming colder in Northern Scandinavia, against the general European trend.

Another new theme he touches on is the probability that dust veil events (*i.e.* collisions with the Earth by comets and meteorites) played a significant part in the history of the 5th and 6th centuries AD. He also sees climate amelioration as contributing, perhaps decisively, to the rise of the Frankish Empire under Charlemagne.

Amos Frumkin, Noam Greenbaum and Ascher Schick suggest that their observations point to a 1,000 years cyclicity. They base their conclusions on an investigation of the paleohydrology of two independent catchments, differing in area by four orders of magnitude, as well as in their lithology and topography. The smaller one (area about 0.078 km^2) is part of the underground salt karst system in the salt plug of Mount Sedom on the western shore of the Dead Sea, while the larger one (1,400 km^2) is part of the surface drainage of the Negev mountains towards the Dead Sea. The synchronous paleo floods records since 2,000 BP allow one to infer in which periods the synoptic conditions over the Negev produced many high intensity storms. The main synchronous periods are 2,000 and 1,000 BP, while the Mount Sedom cave drainage system shows a high flood phase around 3,000 BP as well. The two earlier periods were ones in which agricultural settlements were abundant in the Negev Desert. Frumkin *et al.* suggest that the synoptic situations could have been caused either by Mediterranean cyclonic depressions, which are associated with colder climatic conditions or by a strengthening of the Red Sea trough connected with climate warming. We editors favor the former interpretation, in the light of research we have been doing ourselves. This will also correlate well with several other papers in this volume.

In a study, based on reinterpretation of proxy data, as well as on historical and archaeological evidence, Arie Issar proposes a series of cold-wet and warm-dry periods affecting the Eastern Mediterranean region.

These were: The Pre-Pottery Neolithic A (PPNA), which was a cold and humid period. The Pre-Pottery Neolithic B (PPNB), warm and humid period. The Pottery Neolithic period (8,000–6,200 BP) PPN, mainly cool and humid. The Lower and Middle Chalcolithic, (6,200–5,500?) cold and humid. The Upper Chalcolithic (5,500–5,200 BP), warm and dry. The Early Bronze Period (EB) (5,200/5,100–4,000 BP), cold and humid. The Middle Bronze Period (MB) (4,100/4,000–3,500 BP), warm and dry. The Late Bronze Period (3,500–3,200 BP), and early Iron Age (3,200–3,000 BP), mainly cold and humid. The Late Iron period (3,000–2,200 BP), warm and dry. The Roman Period (2,200–1,700 BP), which was cold and humid. A Roman-Byzantine transition period (1,700–1,800 BP), warm. Byzantine Period (1,800–1,500 BP), cool and humid. The Moslem-Arab period (1,300–1,000 BP), warm and dry. The Crusader Period (1,000–800 BP), cold and humid. The Moslem-Turkish Period (800–500 BP), warm and dry. The Little Ice Age (500–100 BP), cold and humid. The Industrial Period (100 BP to present), warmer and drier.

Michael Netser has come to very similar conclusions on the basis of a detailed survey of historical and archeological data he has carried out in some parts of Israel and Palestine.

Uzi Avner summarizes the investigations carried out by himself and other archaeologists in the Uvda Valley in the southern part of Negev Desert of Israel. These show that in this desert valley agricultural settlements existed continuously from *ca.* 10,000 to *ca.* 4,000 years BP. Yet the zenith of settlement was during the Chalcolithic and Early Bronze periods, namely from 6,000 to 4,000 BP. Avner attributes this flourishing of the desert to a more humid climate. He further suggests that the source of the rain was monsoonal. Yet this explanation makes it difficult for him to find a climatic reason for the fact that this valley continued to flourish during the early Bronze; and he explains this anomaly anthropogenically, the adaptation of these societies to the desert conditions. This explanation may not be necessary, however, if one considers the Early Bronze Period to have been cold and humid as is suggested by Issar and Bar-Mathews *et al.* herein. While during the lower Holocene this area was more humid due to the sub tropical rain regime, it probably profited during the Middle and Upper Holocene from North Westerly rain bearing storms.

The upper part of the Holocene's paleoclimatic time series is given by Mira Bar-Matthews, Avner Ayalon and Aharon Kaufman These authors have carried out a detailed study of the isotopic composition ($d^{18}O$ and $d^{13}C$) of speleothems from a cave in the vicinity of Jerusalem, in the mountainous part of central Israel. The age determinations were by the $^{230}Th/^{234}U$ method. The isotopic record, which is traceable for the last 58,000 years, shows a pronounced difference between the values characterizing the speleothems which were formed before 6,500 BP, and those which were formed later, including the contemporary deposits. This is probably because of altogether different climatic regimes. This is an important observation with regard to the limits on the dimension of time, for which a simulation of the GCM scenarios with data compiled from proxy data can be carried out. With regard to the climate changes during the last 6,500 years, the authors have calculated the paleorainfall values by correlating the paleo $d^{18}O$ records, with the contemporary ratios of $d^{18}O$/rainfall. Based on the $d^{18}O$ and $d^{13}C$ values and calculated paleorainfall they divide the record into 4 stages. Stage 1, which lasted from 6,500 to 5,400 BP, covers the Chalcolithic period which was very wet. During the period extending from 5,600 to 3,050 BP, the Bronze Age, the climate was in general humid, interrupted by four short dry spells. One was between 5,200 and 5,000 BP, which corresponds with the change from the Chalcolithic period to that of the Bronze; and another at *ca.* 4,000 BP, which corresponds with the shift from Early to Middle Bronze. This change had a profound impact on the natural environmental and socio-economic systems in the Middle East. Stage 3, lasting from 3,050 to 1,050 BP, was transitional to drier and more stable conditions. Stage 4, from 1,050 to the present, was characterized by fluctuations in rainfall. The high values between 400 and 500 BP may be connected with the Little Ice Age, while the increase in $d^{13}C$ values, which started *ca.* 700 BP, may indicate a process of deforestation and increased grazing during the Turkish period.

Correlating the data presented by Bar-Matthews *et al.* with the other time series from Israel presented at this workshop one can conclude that in general there is agreement with regard to the extreme changes which had a decisive effect on the environmental and socio-economic systems. For example, the humid periods of the Early Bronze, Middle Iron Age, Early Roman, and Little Ice Age correlate with high Dead Sea levels and low Mediterranean sea levels as well as the expansion of settlements in the arid Negev, while the opposite occurs during the aridity of the several transitions from the Chalcolithic to Early Bronze, Early to Middle Bronze, Byzantine to Arab-Moslem and at the end of the Little Ice Age. The Chalcolithic humid period was not observed as a high Dead Sea level by Frumkin *et al.* This may be due to its erosion by a later and higher Early Bronze shore line.

Similar conclusions with regard to the middle part of the Holocene were reached by Arlene Miller Rosen in Southeastern Anatolya. Her investigation was based on geomorphological sequences in the Urfa Plain and was correlated with time series of proxy data from Anatolya and neighboring countries. The main conclusions from this study are that during the Late Pleistocene period the climate was humid, ranging from very wet in the Middle Paleolithic to moist in the Upper Paleolithic. Although scarce, the evidence from the late Epipaleolithic and earliest Neolithic shows this period was cold and dry. This caused the truncation of Late Pleistocene sediments and their incision. During the early 6th millennium (BC calibrated) (terminal Pottery Neolithic), there is some evidence of more humidity and hence of a renewal of alluvial activity. A major phase of alluvation occurs during the Mid-Holocene from the Middle Chalcolithic through the early Bronze Age. This phase ends at the beginning of the 2nd millennium BC, more or less at the beginning of the Middle Bronze period. During this later period, water tables dropped and the streams began to incise their beds.

Walter Dragoni surveys the historical, archeological and climatic data for the last 3,000 years, for the Italian peninsula south of 43°N. He arrives at the conclusion that, during these years, there were climatic oscillations from warm equated with dry to cool equated with wet, each successive period lasting a few hundred years. During the warm-dry the level of the lakes was at low ebb, and *vice versa* during the wet-humid.

Towards the end of the 19th century, a warming up of the climate can be observed together with a reduction in precipitation. This trend continues to the present time. Extrapolating this trend to the next century shows a decrease in the water yield of up 10–15%, compared to the present.

Leszek Starkel, on the other hand, considers the varying frequency of heavy rainfalls and floods across part of the North European Plain. Such events show a strong tendency to cluster as a reflection of changes in the atmospheric circulation. He establishes concurrence between climatic and geomorphic extreme events, which occurred during the 20th century in the Polish Carpathians and Darjeeling Himalaya. In the second stage he refers to sedimentary and geomorphic features dating from the Little Ice Age. For these he finds synchrony throughout Europe from just before 1600 AD to about 1700 AD, and between 1800 and 1820 AD.

In the Tiber Valley near Rome as in the valleys of central Europe, intervals of frequent floods occurred from 2,050 to 1,800 BP, and from 1,450 to 1,250 BP. Starkel finds a rather good correlation with periods of high floods in the Midwest North America as well as in the southwest desert region of the U.S.A. Meanwhile the Nile in Egypt and the Hoang-ho river of Eastern China have another sequence of high floods due to the fact that they belong to another circulation system, namely the monsoonal. Considering causation, Starkel suggests fluctuations in volcanic and in solar activity as primary. Major intensities may occur when these two causes are superimposed.

Vladimir Kovalevsky summarizes Russian data from the past together with future scenarios obtained by hydro-climatic modeling. An important conclusion is that the global warming by 1°C is liable to bring about an increase of winter and summer temperatures of 2.0–4.0°C in the North of Russia and 0.5–1.5°C in the middle and southern latitudes. The precipitation in the north will be 50–100 mm/y more. In mid latitudes there will be no significant changes, while in the south it will be 25–50 mm/y less. A global rise of temperature by 3.5°C, a prediction for the year 2025, would cause in the polar regions a rise in July temperatures by 12–15°C and January by 15–20°C. In middle latitudes the increase will be 2–5°C in July and 10–15°C in January. The precipitation in the north will increase in 200–600 mm/y and by 50–200 mm/y in the south. But due to higher potential evaporation the hydrological balance in the south will be negative, bringing aridization.

Anne Marie Lézine examines and interprets the palynological time series existing for West Africa. The climate of this region is influenced by Mobile Polar Highs generating anticyclonic cells. These supply the low layer of the fluxes moving towards the meteorological equator as marine trade winds of north south direction and continental trade winds of east west direction. While the marine trade winds contribute to the southward advection of cool water in the Canaries Current, the continental trade wind is responsible for dust and pollen transport to the sea. Over the Meteorological Equator a complex structure exists. In its lower atmosphere the dry continental trade winds ride above the humid monsoon, located near the surface causing westward moving isolated rainstorms. In the middle layers of the atmosphere, the upwelling conditions connected with the Inter tropical Convergence Zone (ITCZ) is the mechanism which is responsible for the abundant and regular rains. During the year the ITCZ remains more or less in the vicinity of the geographical equator, while during summer the meteorological equator moves northward up to 25°N. As a result precipitation moves northward but decreases from north to south causing the zoning of the vegetation according to latitude.

Cores in the Gulf of Guinea and off the west coast of Africa show two well-defined pollen maxima at *ca.* 15,000 and *ca.* 10,300 (Younger Dryas) and a minor peak at 7,000 yrs BP. The first two peaks are characterized by Saharan type pollen, and high amounts of dust derived deposits. In one of the cores the increase in the percentage of pollen of steppic origin also contain assemblages of foraminifera pointing to lower sea temperatures. The pollen in the layers dated from the Holocene come from the nearest tropical forests. Similar trends were observed in a core. So it can be concluded that, during the cold periods of the Last Interglacial, the continent went through a dry period, dominated by northeast trades. During the Holocene these severe conditions did not recur except at *ca.* 7,000 BP.

The scarcity of data makes it difficult to reconstruct the rainfall patterns determined by climatic variations during the Holocene. It seems, however, that during the early Holocene the 400 mm isohyet was placed at about 20°N. From 6,000 BP to 4,500 BP this annual total decreased to 300 at Oyo Suan. Subsequent diminution brought it there to the present level of 5 mm.

Southward in the Sahel humid ecosystems and an abundance of lakes were characteristic of the environment between 9,000 and 8,000 BP. Around 7,500 BP a well marked seasonality of precipitation is evidenced by the pollen assemblage. Increased rates of precipitation and high lake levels were manifested between 4,500 and 2,000 BP. Afterwards conditions in the Sahel changed to the present situation. The equatorial rain forest did not go through any marked changes during the Holocene, except for a short period of dryness at *ca.* 3,000 BP.

Hugues Faure and Liliane Faure-Denard survey the climate changes which the Sahara desert underwent during the Quaternary. Again the main alternations were between 'wet' and 'dry'. During the wet intervals, corresponding to interglacial regimes, inland sand dunes were inactive, flattened, and covered with Sudanian to Sahelian vegetation. The main evidence for wet phases come from laminated diatomites and lacustrine limestone dated, notably at *ca.* 30,000 to 20,000 and *ca.* 9,000 to 6,000 years BP. During the wet episodes, the Sahara was a terrain over which vegetation, fauna and soils were widely expanding with humans having left their mark everywhere. At least five to ten times more carbon than at present was stored in the total mass. During the dry phases eolian conditions of sand and dust movement prevailed, producing areas covered by sands and areas of deflation. During the dry periods, groundwater levels were probably very low. The lacustrine deposits preserved even in the most arid sectors testify to Holocene summer monsoon rains regularly reaching north of the Tropic of Cancer, at around 8,500 radiocarbon years BP.

S.N. Rajaguru and Sheila Mishra in their paper on Upland Western Maharashtra present new data showing that the upland rivers responded to Late Pleistocene aridity by aggradation. Moreover, minor episodes of climate change during the Lower Holocene had their impact on the morphology, with aridity marked by incision.

Mike Bonell conducts a review of the literature about the coupling of atmospheric and hydrological models within the GCM framework in order to get outputs accurate enough to be used in water resource planning. He concludes that considerably more time and effort is needed to reach this goal. The data presented at the present workshop encourages him in urging policymakers to make greater use of analogs of Holocene climate variability within the context of water management.

As the editors of these proceedings we have been struck by the way that they have highlighted the frequency and intensity of climatic change throughout the Holocene. With continual changes in temperature and rainfall in whatever directions the impact of man on these processes must long have been minimal. At the same time these fluctuations have had a profound influence on human affairs as a number of the papers demonstrate. In this regard the past has much to teach us about how to manage the future in this sphere. This may apply at every level. The paleoclimate findings may help in calibrating or otherwise refining the atmospheric and hydrological models and the coupling between them. At the same time measures for mitigating the future impact of climate change can be informed by experience by of adaptation or the lack of it by societies in the past.

Let us conclude by expressing our warm thanks for the help given with the preparation of the conference and the proceedings by Dr. Zvi Shilony of the Ben Gurion Research Center who was the chief organizer and by Mrs. Dorit Elimelech of the J. Blaustein Institute for Desert Research; and, on the Oxford side, by Mrs. Jill Wells and Mrs. Anne Maclachlan.

Arie Issar & Neville Brown

Chapter 1

Application of a Global Volcanicity Time-Series on High-Resolution Paleoclimatic Modeling of the Eastern Mediterranean

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1.1. Introduction

In order to understand adaptation to climate change it is first necessary to set that change within human temporal and geographic dimensions. As Halstead and O'Shea (1989:1) phrase it: "human communities have developed an impressive array of cultural mechanisms for buffering variability. The diversity of these mechanisms, however, should not mask the fact that an effective strategy must match, in both capacity and scale, the variability with which it is to cope." Winterhalder (1980) suggests that a population's adaptive flexibility will not be fully realized if the environment is dynamic in ways not anticipated by analysts. He adds (1980:139) that the "assumption that environments are stable, or that change is either very gradual or abruptly cataclysmic, leads to the reverse of [this] problem: the failure to examine environmental sources of causation when rapid changes are recorded in historical or archaeological records." These passages suggest that if meaningful results are to be derived from studies of the interaction between land-use strategies and climate, human responses to the specifics of climate change (*e.g.* responses to changes in the reliability, magnitude, seasonality, or frequency of rainfall) rather than to climate change in general should be addressed. This concern relates directly to both the temporal and geographic resolution of paleoclimatic models regardless of the cultural group or time period of interest.

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Until recently, the results of paleoclimatic analyses lacked the specificity needed to evaluate adequately these dynamics. It now appears possible, however, to produce useful site-specific simulations of past climates with a temporal resolution on human scales, although not yet on that envisioned by Mehringer and Wigand (1990:294) in their suggestion that "a span of about a generation, or 20 years, is an appropriate period to evaluate potential human response to environmental change." Despite the fact that its methodology has only been developed in the past few years, archaeoclimatic modeling offers considerable promise as a means of gaining insight into the specifics of climate change. To date, the results of these simulations have compared favorably with paleoclimatic reconstructions based on other modeling techniques and analyses of paleoenvironmental proxy data. High-resolution archaeoclimatic modeling was used to produce the site-specific reconstructions which are presented below. How useful these simulations will ultimately be can be assessed only in the light of comparison with cultural and paleoenvironmental field data and of further improvements in the modeling and interpretive methodology.

1.2. Quantitative Paleoclimatic Modeling

1.2.1. MODELING TECHNIQUES

Kutzbach (1985) offers an interesting historical prologue to his discussion of modern techniques of paleoclimatic modeling. Among others, he cites the work of Halley (1693, 1715), Lyell (1837), Croll (1864), and Milankovitch (1920, 1941) as instrumental in setting the stage for current methods. Among these authors, Milankovitch is singled out by Kutzbach (1985:165) as having 'invented' the climate model as an analytical tool through his development of what was essentially an energy balance model relying heavily on changes in the incidence of solar radiation caused by shifts in the Earth's orbital geometry. Since the advent of the widespread use of computers, the approaches which have been employed to model the climate of the Earth quantitatively can essentially be grouped into three general categories: statistical-dynamical, explicit dynamical, and archaeoclimatic modeling. Although these methods all find their antecedents in the early research cited by Kutzbach, each was designed to serve other purposes; and each produces outputs of distinctive geographical and temporal applicability.

Of the three, the first two are broadly similar, except for the time scales within which they customarily operate. In fact, they have sometimes been used in conjunction. Statistical-dynamical models, which include thermodynamic or energy balance models, involve equations which are often formulated in terms of averages for days, months, years or longer intervals. Explicit dynamical models, which include General Circulation Models (GCMs), deal expressly with day-to-day synoptic-scale weather systems and their associated patterns of precipitation; and so require time steps of the order of minutes to hours for the atmospheric portion of the climate mechanism (Kutzbach, 1985:171). The former group have often focused on those parts of the climate system which are slow to respond to macroscale changes in thermodynamic boundary conditions, such as the deep oceans and ice sheets; and have been applied to questions about the evolution of climate on long time scales. The large computational costs of GCMs, on the other hand, have largely limited their application in this field to the production of what Kutzbach (1985:172) refers to as "snapshot views of the climate at specific times in Earth's history." Processes that depend on the particularities of atmospheric flow (such as precipitation) are included in the GCM simulations of paleoclimates.

Since the first paleoclimatic experiment using a GCM with a global domain was conducted by Williams *et al.* (1974) a little over twenty years ago, attempts have been made thus to simulate pre-Quaternary through Holocene climates (Street-Perrott, 1991:74). However, GCMs have very largely been utilized to study alterations in the atmospheric circulation over the last 18,000 years. These have included Kutzbach and Guetter's (1986) early experiments with a set of such interactive models, and those reported by various members of the Cooperative Holocene Mapping (COHMAP) group. The latter have included, for instance, inputs from the COHMAP group as a whole (COHMAP Members, 1988), Kutzbach and Street-Perrott (1985), Kutzbach *et al.* (1991), Harrison *et al.* (1991), Kutzbach and Webb (1991), and Wright *et al.* (1993), to name just a few. The output of these particular experiments has been in the form of paleoclimatic snapshots of the Northern Hemisphere taken in January and July at 3,000 year intervals. COHMAP's goal has been to arrive at "an improved understanding of the physics of the climate system, particularly the response of tropical monsoons and mid-latitude climates to orbitally induced changes in solar radiation and to changing glacial-age boundary conditions, such as ice-sheet size" (COHMAP Members, 1988:1043).

Most of the simulation experiments of the COHMAP group have been conducted using one of the versions of the Community Climate Model (CCM) at the National Center for Atmospheric Research (NCAR) in Boulder, Colorado. The CCM at NCAR uses the non-linear equations of fluid motion and the principles of mass and energy conservation to describe atmospheric dynamics mathematically, including radiative and convective processes, condensation, and evaporation. A horizontal resolution of roughly 4.5° of latitude and 7.5° of longitude is provided by the model (Kutzbach and Ruddiman, 1993:13). Parameters including orbitally-determined insolation, mountain and ice-sheet orography, atmospheric trace-gas concentrations, sea-surface temperatures, sea-ice limits, snow cover, land-surface albedo, and effective soil moisture serve as input boundary conditions for the model. An array of surface and upper air parameters are computed as output from the CCM. These include atmospheric energy budgets as expressed in kinetic and potential energy; zonal, stationary-eddy, and transient eddy components; geopotential height fields; and sea-level pressure. Temperature, wind, and moisture balances over land are of the greatest paleoclimatic interest because they can be compared with surface geologic and biologic evidence (Kutzbach and Ruddiman, 1993:14), although not always directly.

The methodological basis of these, and indeed most, General Circulation Models is essentially microphysical in nature. This means that the equations of motion are used primarily with reference to forces and influences on individual parcels of air, including the effects of as many surface boundary conditions as are computationally (or economically) feasible. In general, this approach starts with these small segments of the atmosphere and through iteration works up to global weather patterns which are then averaged to represent the climate. The iteration is usually with rather short time steps. So climatic simulations may require millions of calculations (Bryson, 1993). This factor alone can cause significant error, however, the strongest criticisms of GCMs relate to their coarse spatial resolution (a *ca.* 5° grid) and modified treatment of topographic features. As regards the first concern, Street-Perrott (1991:74) notes that key climatic changes (such as the strengthened Northern Hemisphere monsoons in about 9,000 BP) depend on only about a 5° shift in the positions of major circulation features which cannot be adequately delimited with current GCM methodology. The spherical harmonics used to simulate the Earth's topography in these models create a situation in which certain of the major features that produce much of the spatial variability of climates are altogether missing, while others are misplaced (Thompson *et al.*, 1993). In general, GCMs do have sufficient resolution to describe variations in large-scale atmospheric circulation patterns. But they cannot accurately simulate smaller-scale circulation features nor achieve the interactions with topography that would likely explain much of the spatial variability found in the climates of many areas (Mock and Bartlein, 1995).

1.2.2. ARCHAEOCLIMATIC MODELING

A third method of quantitative paleoclimatic modeling is comparatively simpler than general circulation modeling and may not require iteration. This is the macrophysical approach, termed 'archaeoclimatic modeling' by Bryson and Bryson (1997). Models of this sort are also complicated but, being hierarchical, can be run on a microcomputer. Thus computational costs are minimal. This methodology can be applied in either hemisphere; the modeling output is site-specific; and its temporal resolution ranges from 200 to 500 year intervals for the Holocene and Late Pleistocene. Unlike GCMs, archaeoclimatic models were developed to provide paleoclimatic reconstructions of use to scientists whose interests focus on interpreting the relationship between climatic change and the ecological and/or cultural dynamics observed in the field.

Perhaps most importantly, however, the use of these models requires one to accept a fundamentally different definition of climate than has been applied in the development of GCMs. An experienced meteorologist may readily, without looking at the detailed data, determine whether the atmospheric circulation on a weather map represents a summer pattern or a winter one. Usually, the identification can be even closer. The reason the array of weather patterns characteristic of one season differs from the array of another is simply that the climate differs from season to season. Yet this statement does not make sense if the climate is defined as the summation of the 'average' weather over some period. It does make sense, however, if the thermodynamic-hydrodynamic status of the earth-atmosphere-hydrosphere-cryosphere system *determines* the array of possible (and necessary) weather patterns. The fundamental premise of archaeoclimatic modeling is that this status, which changes with time and season, *along with* the associated weather patterns, constitutes the climate. In other words, climate is the product of the global boundary conditions which determine the concurrent array of weather complexes, which differ as the climate differs from one time to another (Bryson, 1997). This is why the climatic state thought of as summer has an array of associated weather patterns different from those in the climatic state called winter.

Modeling the climate on this basis adopts what has been called the 'macrophysical' approach in that relationships of a large scale nature are used, such as the Rossby long-wave equations, the thermal wind relationship, and the Z-criterion derived from the work of Smagorinsky (1963). These are examples which properly apply on the scale of a few thousand miles or kilometers and of hours to days (Bryson, 1993:342). The archaeoclimatic reconstructions presented below were done with such a model, which is based on a series of hierarchical steps, each of which depends on certain underlying assumptions. These steps are summarized in Figure 1.1. This process first required a model of global glacial volume, from which the glaciated area could be calculated and the ice albedo effect estimated (Bryson and Goodman, 1986). By using next the Milankovitch (1941) variations in solar irradiance as calculated by Hopkins (1985), and allowing for their modulation by volcanic aerosols (Bryson, 1988), it was possible to generate a hemispheric temperature model. This section of the overall model is essentially a 'heat budget model' but one with explicit treatment of the volcanic modulation.

Deriving regional models from hemispheric or global models requires that some broad-scale relationships be established. For example, because equatorial temperatures change little even from glacial to non-glacial times, the average hemispheric temperature and the equator-to-pole (or meridional) temperature gradient must both depend primarily on high-latitude temperatures (Bryson, 1993:342). Indeed, it can be shown that the meridional temperature gradient is a relatively simple function of the hemispheric mean temperature. It follows that, if the hemispheric mean temperature can be modeled through time, so can the meridional temperature gradient. Further, given that the westerlies must become stronger with height up to about 10 km because the earth is colder at the poles than at the equator, and that the rate of increase is proportional to the magnitude of the meridional temperature gradient, it is possible to use this gradient as a proxy for the strength of the westerlies.

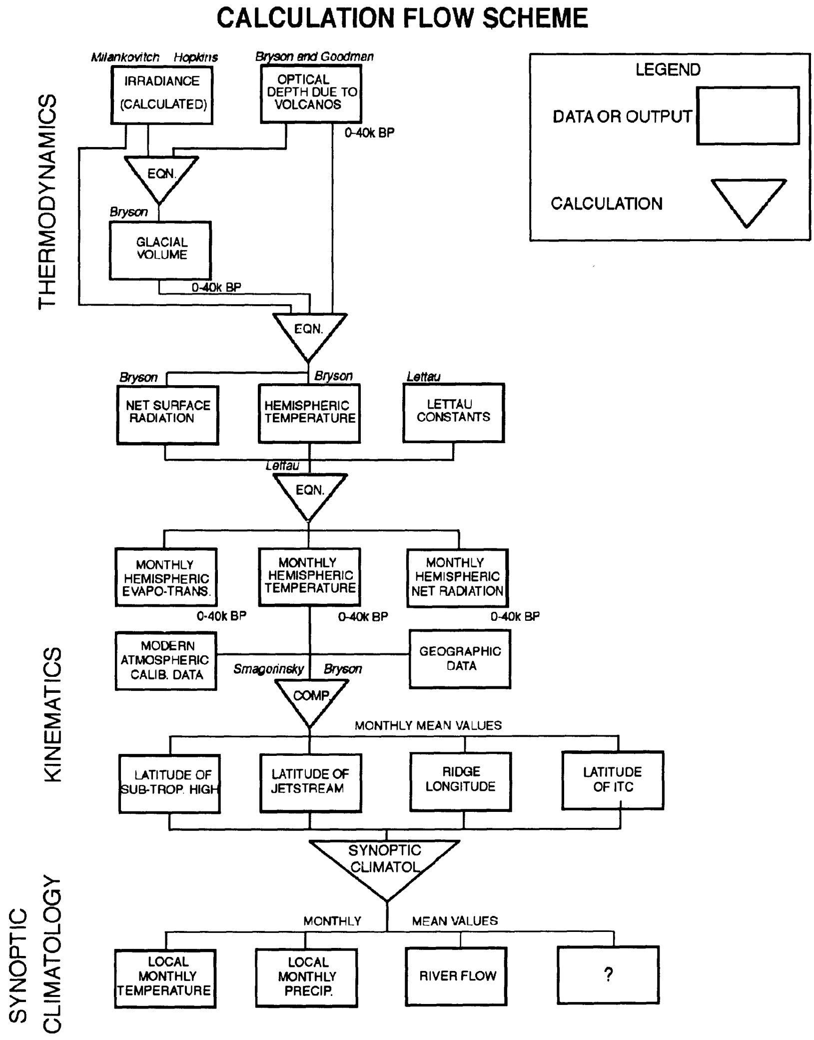

Figure 1.1. Schematic structure of the hierarchical macrophysical model.

In developing this modeling approach, Bryson (1989) applied several assumptions about the relationship between the zonal component of the wind at 500 mb (*i.e.* at an elevation of *ca.* 5500 m), the meridional temperature gradient, the meridional seasonal component, and the inter-seasonal contrast to build a simple model of the direction over time of the northwest India monsoon wind. He then related this to area rainfall. This consideration of the monsoon was then extended to North Africa, using the Z-Criterion (Smagorinsky, 1963; Flohn, 1965) to calculate the latitude of the subtropical anticyclone and that of the jet axis at 500 mb (Bryson, 1992). Empirical values of the appropriate gradients were derived from the inter-monthly variation, including the relation between the latitude of the subtropical anticyclone and that of the Intertropical Convergence (ITC). Ilesanmi's model (1971) relating the position of the ITC to rainfall south of it could then be used to estimate rainfall in the Sahel at two or five century intervals during the late Pleistocene and Holocene. The match between the model results and the available field data proved to be quite satisfactory.

In general, if the latitude of the jetstreams and the locations of the subtropical anticyclones can be determined over time, then these and other major atmospheric circulation features (formerly called 'centers of action') can be used to model local rainfall and precipitation over the same period. This is accomplished through application of the techniques of synoptic climatology, by which the behavior of a climatic element is explained in terms of atmospheric circulation patterns, particularly the positions of the major features. Indeed, in many ways archaeoclimatic modeling can be thought of as synoptic paleoclimatology. This is based on the reasonable premise that, for any particular place, the *relationship* between the monthly positions of the 'centers of action' and monthly precipitation (or temperature) has remained essentially constant through the very late Pleistocene and Holocene. In other words, it is assumed that the physics of the situation has remained the same over this period. This relationship can be determined through modern synoptic climatology and calibrated by the multiple regression, not necessarily linear, of the current (*i.e.* observed) precipitation against the current locations of the pertinent circulation features. It then becomes possible to calculate past monthly precipitation from the modeled past positions of the centers of action.

Central to the methodology of archaeoclimatic modeling is the use of a volcanicity index, based on radiocarbon-dated volcanic eruptions, from which the optical aerosol depth is calculated. In short, a serpentine curve is fitted to the frequency distribution of radiocarbon dates (Figure 1.2), allowing calculation of an expected frequency at any time interval. Departures from the expected values can then be expressed in terms of a volcanicity index (Figure 1.3) which has been shown to be closely correlated with optical aerosol depth (Goodman, 1984; Bryson, 1988). Because the temporal structure of these models derives from the volcanic database, the time depth of this archaeoclimatic modeling theoretically extends to the limits of the radiocarbon method, about 40,000 years. As a result of the high frequency terms which are introduced by the volcanic aerosol modulation of solar radiation, however, the temporal resolution of the modeling is very high. It is currently possible to produce archaeoclimatic models providing monthly values of temperature or precipitation for 200 year intervals back to 14,000 BP and 500 year intervals from there back to 40,000 BP The limiting factor is the distribution of dated volcanic eruptions, particularly for the early part of the sequence, due to the limited preservation of datable materials.

The modeling output may be site-specific because the means by which the synoptic relationships between precipitation (or temperature) and the major circulation

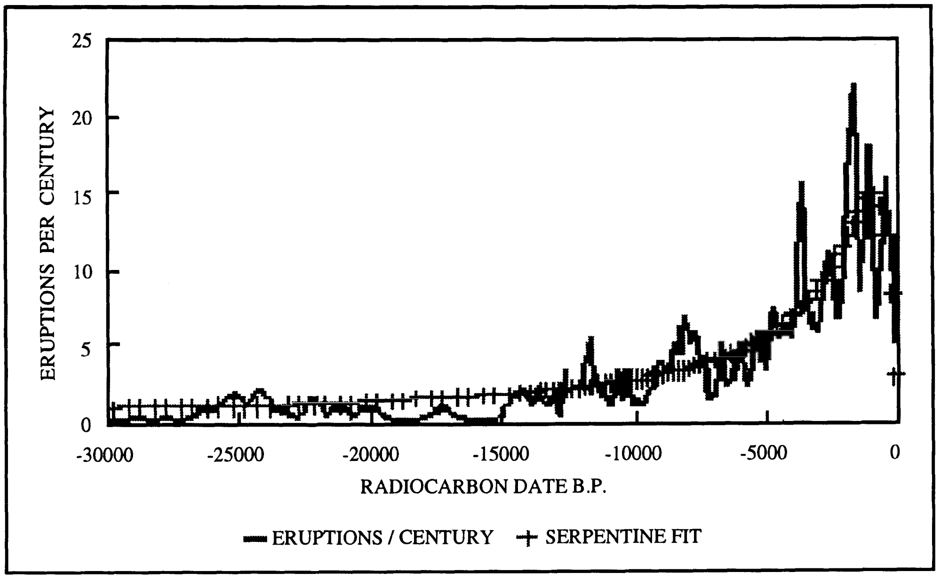

Figure 1.2. The fit of a serpentine curve to the frequency distribution of radiocarbon-dated volcanic events in 200 to 500 year intervals. The use of this curve helps to compensate for that portion of the distribution that is solely related to the poor preservation of very old datable material as well as that resulting from the low number of dates run on materials from the historic era. When this methodology was originally reported (*e.g.* Bryson, 1988), a power law curve was used instead of the serpentine curve which appears here. The current approach appears to provide a better representation of the frequency distribution.

features are calibrated can utilize modern climatic data from a site of interest. As a further result, this modeling methodology is theoretically applicable anywhere on the globe, although there are certain practical limitations to that coverage. In cases where no recording station is located in close proximity to the site, it is sometimes possible to simulate local monthly precipitation and mean temperatures by following an adaptation of V. Mitchell's (1969) methodology as applied by Bryson and Bryson (1997).

The surface boundary conditions chosen by various researchers using GCMs represent their attempts to quantify explicitly some of the mechanisms, other than the Earth's orbital variations, which have affected atmospheric circulation on a global or hemispheric scale. However, this dimension does not comprehend short term variance in the paleoclimatic record. Nor does it resolve at a regional level the spatial heterogeneity of climates resulting from how the influence of large-scale circulation controls is locally mediated by topography (Whitlock and Bartlein, 1993; Whitlock *et al.*, 1995). In contrast to the GCMs, archaeoclimatic modeling works from the premise that, as a first approximation, the volcanic aerosol modulation of insolation embraces enough of the high frequency variance present in the climatic system for short-term paleoclimatic fluctuations to be satisfactorily represented in

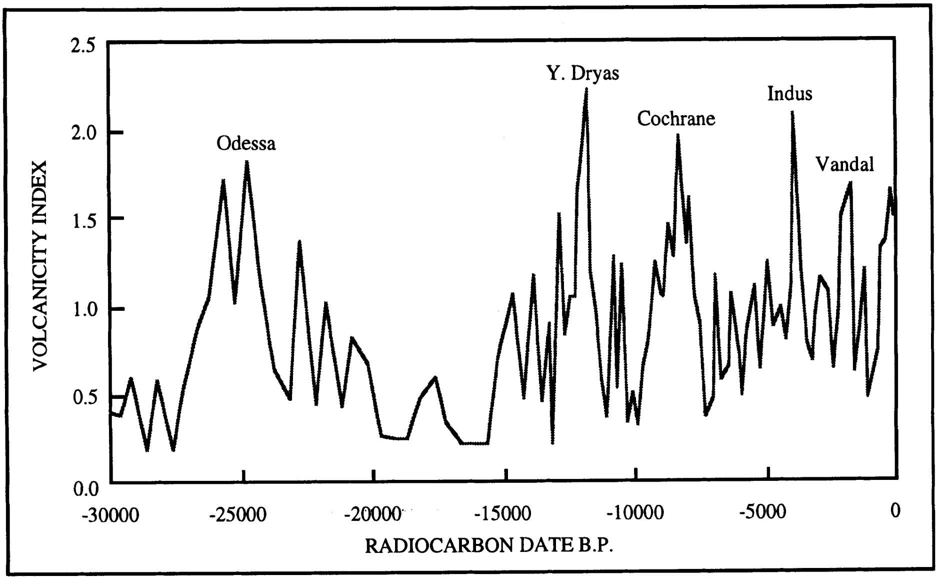

Figure 1.3. This graph presents the resultant volcanicity index used to calculate optical aerosol depth following the methodology described by Goodman (1984) and Bryson (1988). Named climatic episodes are labeled above the curve.

the models. This is considered to be the case particularly because the methodology also entails the implicit inclusion of internal adjustments between different climatic subsystems through its dependence on the synoptic relationship between circulation features and local conditions, a relationship calibrated by local weather data. In essence this results from the simple truth that the relationship of local conditions to the circulation features is conditioned by the specifics of site topography and regional geography. Often the linkage is non-linear and must be dealt with as such during calibration. Still, the effects of local conditions are considered in such a fundamental way that they do not usually need to be specified further. This is a big advantage of this type of modeling.

Additional descriptions of the methodology and underlying assumptions of archaeoclimatic modeling are presented in Bryson and Bryson (1996, 1997). In all, archaeoclimatic models have so far been constructed for over 300 sites worldwide.

1.3. Archaeoclimatic Models of Northern Africa

1.3.1. THE SAHARAN REGION

The northern part of Africa and the eastern Mediterranean currently experience winter rains and summer drought, with the rains diminishing from west to east and inland towards the heart of the Sahara. The winter rains are associated with the southward extension of the circumpolar westerlies and the jetstream, and the passage of cyclonic storms within the westerlies during that season. In summer, as this circumpolar vortex shrinks, the subtropical anticyclone in the North Atlantic expands and extends eastward, bringing the Mediterranean coast under divergent flow from the northeast quadrant of the anticyclone (LaFontaine *et al.*, 1990).

South of the central Sahara, summer rains penetrate inland from the Gulf of Guinea. This is associated with the invasion of moist maritime air behind the Intertropical Convergence (ITC or ITCZ). Ilesanmi (1971) has shown that these rains actually increase southward for some distance behind the ITC but then decrease until close to the coast where monsoonal rains cannot be distinguished from those associated with coastal effects. How far these monsoonal rains penetrate into the interior is related to the latitude attained by the subtropical anticyclone in its annual displacement. Thus both the winter rains of the north and the monsoonal rain of the south can be related to the positions of the subtropical anticyclone, the jetstream, and the ITC. These fluctuate throughout the year. Moreover, their seasonal movements changed considerably over the late Pleistocene and Holocene (Bryson, 1992).

Following this line of reasoning, Bryson (1992) developed a macrophysical model of rainfall in the Sahel for the band between 17° and 21°N. He compared this simulation with the frequency distribution of radiocarbon dates of named cultures in the same zone, expressed as an 'Occupation Index', which simply sums the raw counts falling in the same 200 year intervals used in the precipitation simulation. This comparison is reproduced in Figure 1.4, with the Occupation Index multiplied by a constant scaling factor for ease of presentation.

Although the two curves do not correspond perfectly, as has to be expected for a variety of reasons (*cf.* Bryson, 1985), there is a sufficiently good match between the *patterns* of both curves to suggest that the precipitation model is valid in this sort of areal application. Indeed, similar results can be derived by a comparison of the simulated rainfall with the set of dates on indicators of moisture in the Sahel (biological, cultural, and geological) assembled by Petit-Maire *et al.* (1993). This raises the important point that it is the pattern of modeled paleoclimatic change (rather than the absolute values of, for instance, precipitation) which appears most useful in comparisons with environmental proxy data.

1.3.2. A RECONSTRUCTION OF THE FLOW OF THE NILE

The latitude to which the monsoon (and the ITC) penetrate into west Africa is not the same at all longitudes, probably because of the shape of the continent. If one assumes that this effect has been the same as long as the shape of the continent has, past rainfall at various places can be estimated by applying Ilesanmi's (1971) methodology. In East Africa the relationship between the position of the ITC and rainfall is more complex than in West Africa, but not intractably so. In each case, how-

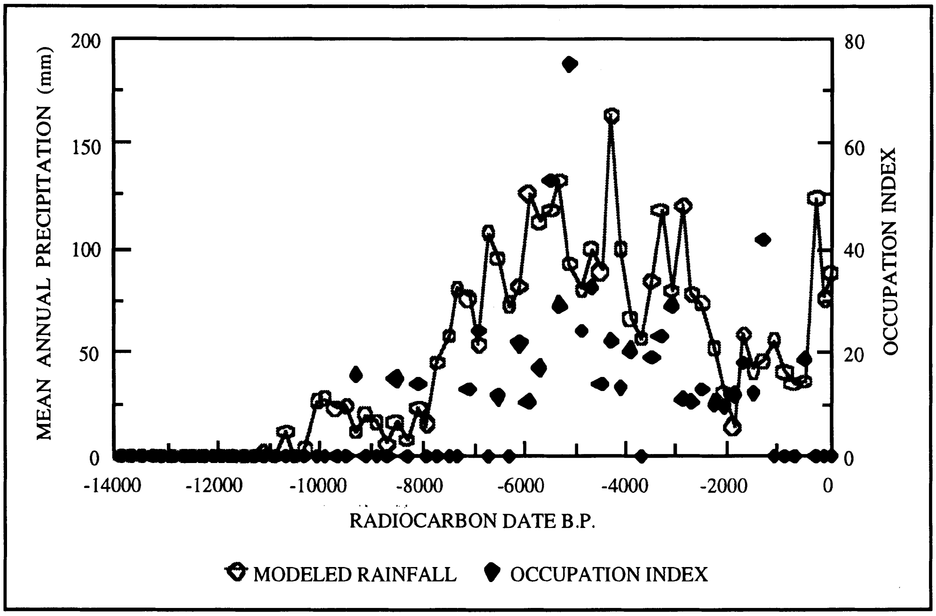

Figure 1.4. In this graph, modeled rainfall in the 17–21°N band across north Africa (circles) is compared with an Occupation Index (diamonds) reflecting the number of cultural radiocarbon dates reported from this region (adapted from Bryson, 1992).

ever, the rainfall intensity is different when the ITC is advancing than when it is retreating (Bryson, 1992).

The bulk of the Nile discharge at peak flood (*ca.* 95%) comes from the Blue Nile, which carries precipitation from northern Ethiopia, while most of the remainder (*ca.* 5%) comes from Lake Victoria via the White Nile (Waterbury, 1979). The other main Nile tributary, the Atbara, can be assumed to parallel in behavior the Blue Nile because of the proximity of their sources. To estimate the peak discharge, which is a good indicator of how much flooding of the arable area would have occurred, monthly rainfall amounts on the headwaters of the Blue and White Nile were modeled and summed to produce annual means (Figure 1.5) using fundamentally the same approach as that applied by Bryson (1992) in modeling rainfall in the Sahel. This was possible because the processes are similar. Monthly rainfall at the Nile headwaters is demonstrably related to the position of the ITC, though not as simply as in West Africa nor is the relationship quite the same for northern Ethiopia as for the Lake Victoria region. Scaling factors have been used to produce estimates of Nile discharge at Aswan in millions of cubic meters/day. These can be compared with Hassan's (1985) reconstruction of levels of Lake Moeris, a Nile overflow basin in the Fayum (Figure 1.6) as well as with cultural evidence presented by Hassan and others. The match between the modeled output and Hassan's field data appears to be quite satisfactory for the last 9,000 years.

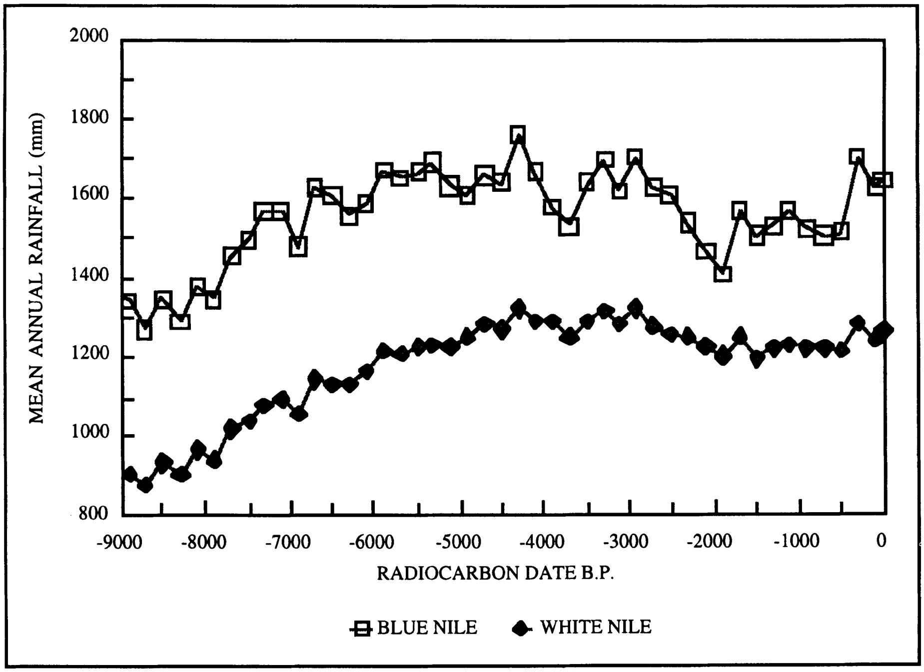

Figure 1.5. Modeled precipitation on the headwaters of the Blue Nile and the White Nile.

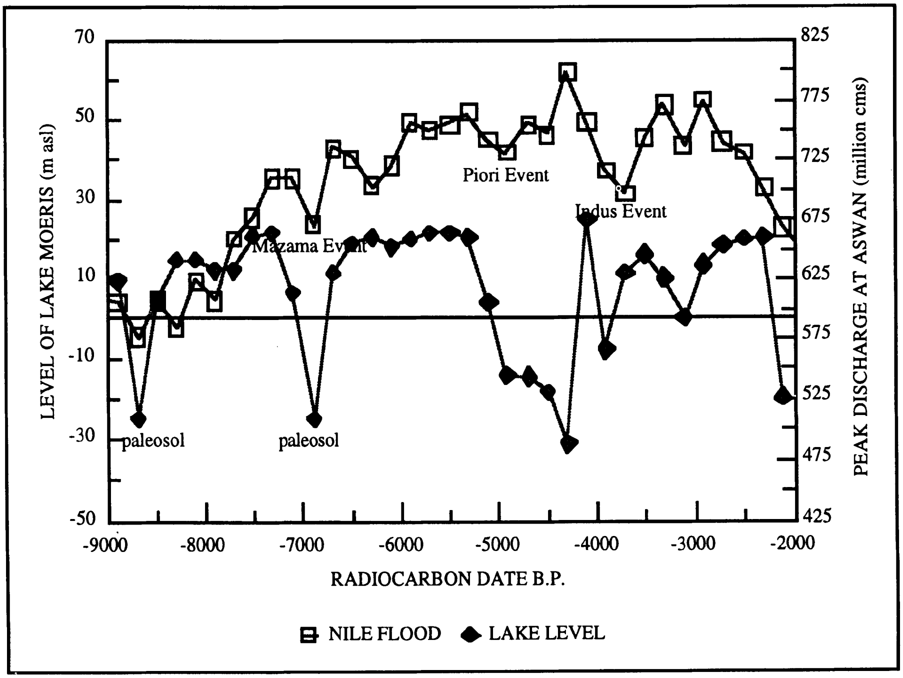

Figure 1.6. Comparison of modeled Nile flow with reconstructed levels of Lake Moeris (adapted from Hassan, 1985).

1.3.3. THE LEVANT

Analysis of the modern climate of Syria shows that the monthly distribution of precipitation is closely related to the alignment of the Mediterranean branch of the jetstream. Monthly precipitation is relatively uniform in quantity north of the jet and absent south of it. Because past monthly jetstream locations were previously modeled by Bryson (1992), the simulation of monthly precipitation at specific sites is relatively straightforward. The results of such a simulation for Kameshli, Syria—close to the archaeological site of Tell Leilan discussed by Weiss *et al.* (1993)—are presented in Figure 1.7 along with the modeled Nile flood. One benefit of this comparison is food for thought regarding the ebb and flow of power between the civilizations of the Nile and Mesopotamia. The severe drought suffered by occupants of Tell Leilan is noted by Weiss *et al.* and is labeled the 'Tell Leilan Event' in Figure 1.7 (from Hassan, 1985).

1.3.4. THE DEAD SEA

Although the Dead Sea lies roughly equidistant between Egypt and Syria, its climate is more closely akin to the latter because of its position relative to seasonal movements of the jetstream and long distance from the area of monsoon rains. It

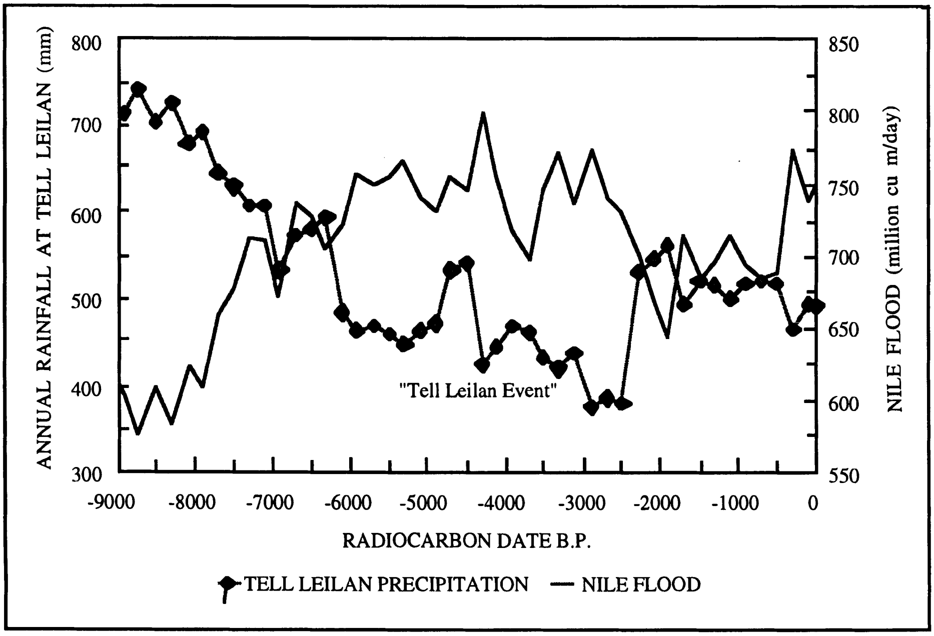

Figure 1.7. In this figure the sequence of increases and decreases in the Nile flood are contrasted with the modeled precipitation history of Tell Leilan (modern day Kameshli, Syria). The major drought described by Weiss *et al.* (1993) is noted as the 'Tell Leilan Event'.

should be possible to model former levels of the Dead Sea using the methods described here combined with hydrographic data. However, as a first approximation for present purposes, the flow of the Jordan River, assumed always to be the principal supplier of discharge into the basin, is simulated by modeling the precipitation history of Jerusalem. In Figure 1.8, this simulation is compared with the summary of Mount Sedom cave passage width ratios presented in Frumkin *et al.* (1994:317). The triangles in the figure represent the distribution of actual passage width ratio measurements dated by radiocarbon assays of wood and other plant material found in the caves. The passage width curve itself, however, presents the authors' interpretation (1994:323) of the actual data.

It is clear upon inspection of Figure 1.8 that the reconstruction of Frumkin *et al.* and the simulation for Jerusalem correspond reasonably well only for the late Holocene—*i.e.* since 3,500 BP. The same is true if the modeled precipitation history is compared with the figure depicting fluctuations of Dead Sea levels and sediment types in the south basin presented by the same authors (1994:324). This result could be due to a variety of different factors acting either singly or in consort. It is most probable that controls over Dead Sea levels are considerably more complex than can be accounted for in this relatively simplistic comparison and, at any rate, the Mt. Sedom data represent extreme events rather than annual averages and Jerusalem is not particularly close to the headwaters of the Jordan River.

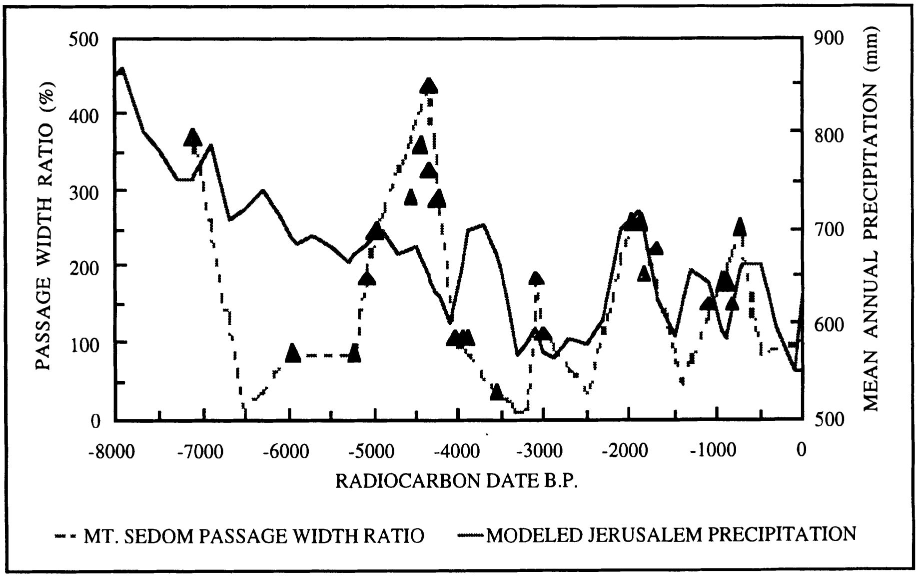

Figure 1.8. Compared here are the modeled precipitation history of Jerusalem and the Mount Sedom passage width ratios compiled by Frumkin *et al.* (1994).

1.3.5. THE RED SEA LITTORAL

A final set of archaeoclimatic models is offered as a means of demonstrating the need for paleoclimatic models with high geographic resolution. During the first millennium BCE, there were a number of prosperous states which developed around the trading corridor of the Red Sea. Among these were the 'incense kingdoms' of Axum, in Ethiopia, and Sabaea, in Arabia. The modeled precipitation histories for Massawa and Aden presented in Figure 1.9 should be representative of changing climatic conditions in Axum and Sabaea respectively. It is clear that although these two sites lie in relatively close proximity to one another, their modern climates are quite different and their paleoclimates appear to have followed very different courses as well.

Once again, if meaningful conclusions are to be drawn from studies of the interaction of culture and climate, it is necessary to consider the consequences of climatic change on a regional or local level. Perhaps the large dam built near the Sabaean capital of Ma'rib around 600 BCE was designed to mitigate the effects of a rapidly declining precipitation regime in an already arid region. Whatever the case may be, there are many such questions that cannot be adequately addressed without

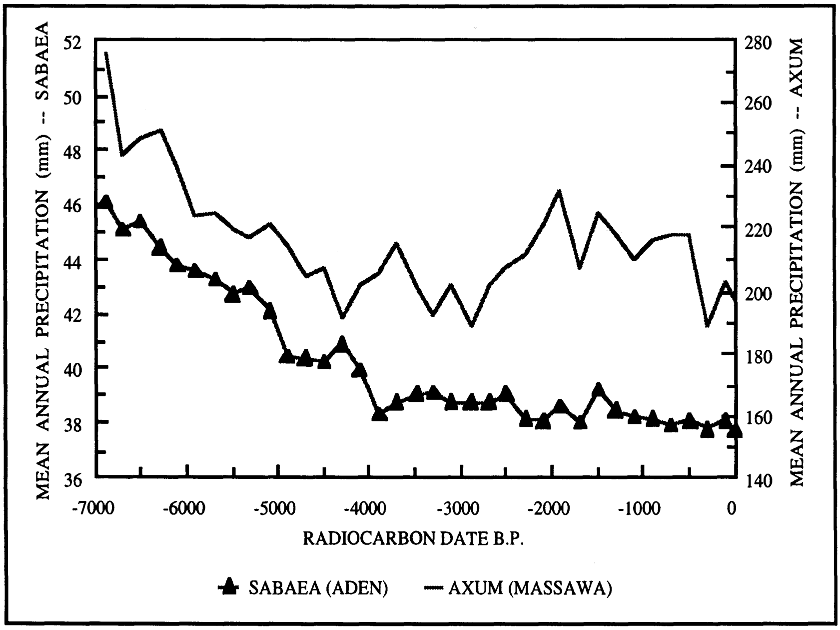

Figure 1.9. The modeled precipitation histories for the 'Incense States' of Axum (modeled based on data from modern day Aden) and Sabaea (modeled based on data from Massawa) are contrasted in this graph. Despite the fact that the sites are in relatively close proximity to one another, these two models display significantly different patterns.

paleoclimatic studies that provide sufficiently high spatial and temporal resolution to allow assessment of the regional effects of paleoenvironmental variability.

1.4. Summary and Concluding Remarks

As discussed in brief above, archaeoclimatic modeling of paleoclimates is based on a series of hierarchical steps, each of which relies on certain underlying assumptions. These steps are summarized in Figure 1.1 and the assumptions are detailed in Bryson and Bryson (1996) and in Bryson *et al.* (1996). In general, incoming solar radiation values are derived from Hopkin's (1985) calculations of the Milankovitch periodicities based on the earlier work of Berger (1978). Loadings of volcanic aerosols, calculated on the basis of an index of radiocarbon-dated volcanic eruptions, are assumed to be the major factor affecting atmospheric optical depth and hence insolation. Together, diachronic changes in insolation and optical depth are used to calculate glacial volume and, from that, glacial area and albedo (*i.e.* reflectivity) figures. These findings are then combined at the hemispheric level to model seasonal histories of mean temperature for the radiocarbon period of record.

From these data, monthly values of hemispheric temperatures, net radiation, and evaporation can be calculated through the application of the climatonomic protocols developed by Lettau (*e.g.* 1969). A relatively simple calculation can convert the monthly hemispheric temperatures into meridional (equator to pole) temperature gradients which can then be used to ascertain the locations over time of major atmospheric circulation features. Synoptic climatology is used to determine the relationship between modern monthly precipitation (or temperature) and the modern locations of pertinent circulation features. This relationship is calibrated through a series of statistical steps and then applied to the modeled past monthly positions of the relevant circulation features in order to calculate past precipitation (or temperature) at specific intervals.

Archaeoclimatic models of the Nile and of several sites in the eastern Mediterranean were presented and compared for the sole purpose of generating new questions about the role of climatic change in the history of the region. Although only modeled precipitation histories were offered here, archaeoclimatic modeling of temperatures at specific sites follows the same methodology as that applied to precipitation and produces output with the same temporal resolution. Typically, however, a different set of synoptic features of the atmosphere is considered in developing simulated precipitation histories. In all, the comparison of such simulations with the paleoenvironmental and cultural proxy records made to date suggest that archaeoclimatic modeling offers considerable promise for providing hypotheses about the nature of the climate changes which past cultures experienced and to which they had to adapt.

References

Bryson, R.A. (1985) On climatic analogs in paleoclimatic reconstruction. *Quaternary Research* **23**, 275–286.

——. (1988) Late Quaternary volcanic modulation of Milankovitch climate forcing. *Theoretical and Applied Climatology* **39**, 115–125.

——. (1989) Modeling the NW India monsoon for the last 40000 years. *Climate Dynamics* **3**, 169–177.

——. (1992) A macrophysical model of the holocene intertropical convergence and jetstream positions and rainfall for the Saharan region. *Meteorology and Atmospheric Physics* **47**, 247–258.

——. (1993) Simulating past and forecasting future climates. *Environmental Conservation* **20(4)**, 339–346.

——. (1997) The paradigm of climatology: an essay. *Bulletin of the American Meteorological Society* **78(3)**, 449–455.

Bryson, R.A. and Bryson, R.U. (1996) High resolution simulations of regional holocene climate: North Africa and the near east. *In* H.N. Dalfes, G. Kukla, and H. Weiss (eds.), *Climatic Change in the Third Millenium BC,* NATO ASI Series, Subseries I 'Global Environmental Change', pp. 565–593.

Bryson, R.A., and Goodman, B.M. (1986) Milankovitch and global ice volume simulation. *Theoretical and Applied Climatology* **37**, 22–28.

Bryson, R.U. and Bryson, R.A. (1997) A comparison of cultural evidence and simulated holocene climates of the Pacific northwest: an experiment in archaeoclimatology. *In* A.E. Oetting (ed.), *Occasional Papers of the Association of Oregon Archaeologists, Vol. 6,* Association of Oregon Archaeologists, Eugene, OR. (in press).

Bryson, R.U., Bryson, R.A., and Ramirez, L.M. (1996) Paleoclimates of the Northwestern Great Basin: Northeast California and Western Nevada. *In* M. Delacorte (ed.), *Archaeological Investigations along the Tuscarora Pipeline Corridor, Volume 1 Environmental Overview,* Far Western Anthropological Research Group, Davis, CA., pp. 26–48.

COHMAP Members (1988) Climatic changes of the last 18,000 years: observations and model simulations. *Science* **241**, 1043–1052.

Croll, J. (1864) On the physical cause of the change of climate during geological epochs. *Philosophy Magazine* **4(28)**, 121–137.

Flohn, H. (1965) Probleme der theoretischen Klimatologie. *Naturwissenschaftliche Rundschau* **10**, 385–392.

Frumkin, A., Carmi, I., Zak, I. and Magaritz, M. (1994) Middle holocene environmental change determined from the salt caves of Mount Sedom, Israel. *In* O. Bar-Yosef and R.S. Kra (eds.), *Late Quaternary Chronology and Paleoclimates of the Eastern Mediterranean,* Radiocarbon, Department of Geosciences, University of Arizona, Tuscon, pp. 315–332.

Goodman, B.M. (1984) The Climatic Impact of Volcanic Activity. Unpublished Ph.D. dissertation, Department of Meteorology, University of Wisconsin-Madison.

Halley, E. (1693) A discourse concerning the proportional heat of the sun in all latitudes, with the method of collecting the same. *Philosophical Transactions of the Royal Society of London* **17**, 878–885.

Halley, E. (1715) On the causes of the saltiness of the oceans, and of the several lakes that emit no rivers. *Philosophical Transactions of the Royal Society of London* **29**, 296–300.

Halstead, P. and O'Shea, J., eds. (1989) *Bad Year Economics: Cultural Responses to Risk and Uncertainty.* Cambridge University Press, Cambridge, U.K.

Harrison, S.P., Kutzbach, J.E. and Behling, P.J. (1991) General circulation models, palaeoclimatic data and last interglacial climates. *Quaternary International* 10(12), 231–242.

Hassan, F. (1985) Holocene lakes and prehistoric settlements of the western Faiyum, Egypt. *Journal of Archaeological Research* 13, 483–501.

Hopkins, E.J. (1985) Analysis of Astronomically-Induced Monthly and Zonally Averaged Extra-Atmospheric Irradiance variations of the Earth over the Last 500,000 Years. Unpublished Ph.D. dissertation, Department of Meteorology, University of Wisconsin, Madison.

Ilesanmi, O.O. (1971) An empirical formulation of an ITD rainfall model for the tropics: a case study of Nigeria. *Journal of Applied Meteorology* 10, 882–891.

Kutzbach, J.E. (1985) Modeling of paleoclimates. *Advances in Geophysics* **28A**, 159–196.

Kutzbach, J.E., and Guetter, P.J. (1986) The influence of changing orbital parameters and surface boundary conditions on climate simulations for the past 18,000 years. *Journal of the Atmospheric Sciences* **43**, 1726–1759.

Kutzbach, J.E., and Ruddiman, W.F. (1993) Model description, external forcing, and surface boundary conditions. *In* H.E. Wright *et al.* (eds.), *Global Climates Since the Last Glacial Maximum*, pp. 12–23.

Kutzbach, J.E., and Street-Perrott, F.A. (1985) Milankovitch forcing of fluctuations in the level of tropical lakes from 18 to 0 kyr BP. *Nature* 317(6033), 130–134.

Kutzbach, J.E. and Webb, T., III (1993) Conceptual basis for understanding late-quaternary climates. *In* H.E. Wright *et al.* (eds.), pp. 5–11.

Kutzbach, J.E., Gallimore, R.G., and Guetter, P.J. (1991) Sensitivity experiments on the effect of orbitally-caused insolation changes on the interglacial climate of high northern latitudes. *Quaternary International* 10–12, 9–28.

LaFontaine, C.V., Bryson, R.A., and Wendland, W.M. (1990) Airstream regions of North Africa and the Mediterranean. *Journal of Climate* 3(3), 366–372.

Lettau, H.H. (1969) Evapotranspiration climatonomy, I. A new approach to numerical prediction of monthly evapotranspiration, runoff and soil moisture storage. *Monthly Weather Review* 97, 691–699.

Lyell, C. (1837) *Principles of Geology*. Volume 1. James Kay Jr. & Brother, Philadelphia, Pennsylvania.

Mehringer, P.J., Jr. and Wigand, P.E. (1986) Western Juniper in the Holocene. *In* R.L. Everett (ed.), *Proceedings of the Pinyon-Juniper Conference*, USFS Intermountain Research Station, General Technical Report INT-215, pp. 109–119.

Milankovitch, M.M. (1920) *Théorie Mathématique des Phénomènes Thermiques Produits par la Radiation Solaire*. Gauthier-Villars, Paris.

Milankovitch, M.M. (1941) Kanon der Erbdestrahlung und seine Anwendung auf das Eiszeit-problem. *Königlich Serbische Akademie Édition Spécializat* 133, 1–633. Beograde, Yugoslavia. (English translation published by Israel Program for Scientific Translations, Jerusalem. (U.S. Department of Commerce, Washington, D.C., 1969).

Mitchell, V.L. (1969) The Regionalization of Climate in Montane Areas. Unpublished Ph.D. dissertation, Department of Meteorology, University of Wisconsin, Madison.

Mock, C.J. and Bartlein, P.J. (1995) Spatial variability of late-quaternary paleoclimates in the western United States. *Quaternary Research* **44**, 425–433.

Petit-Maire, N., Page, N., and Marchand, J. (1993) *The Sahara in the Holocene: Map 1/5,000,000*. UNESCO, CGMW. Paris (FR).

Smagorinsky, J. (1963) General circulation experiments with the primitive equations, I: The basic experiment. *Monthly Weather Review* 91, 99–164.

Street-Perrott, F.A. (1991) General Circulation (GCM) Modelling of palaeoclimates: A critique. *The Holocene* 1(1), 74–80.

Thompson, R.S., Whitlock, C., Bartlein, P.J., Harris, S.P., and Spaulding, W.G. (1993) Climatic changes in the western United States since 18,000 yr B.P. *In* H.E. Wright *et al.* (eds.), pp. 468–513.

Waterbury, J. (1979) *Hydropolitics of the Nile Valley*. Syracuse University Press, Syracuse, New York.

Weiss, H., Courty, M.A., Wetterstrom, W., Guichard, F., Senior, L., Meadow, R., and Curnow, A. (1993) The genesis and collapse of a third millennium North Mesopotamian civilization. *Science* **261**, 995–1004.

Whitlock, C. and Bartlein, P.J. (1993) Spatial variations of holocene climatic change in the northern Rocky Mountain region. *Quaternary Research* **39**, 231–238.

Whitlock, C., Bartlein, P.J., and Van Norman, K.J. (1995) Stability of holocene climate regimes in the Yellowstone region. *Quaternary Research* **43**, 433–436.

Winterhalder, B.P. (1980) Environmental analysis in human evolution and adaptation research. *Human Ecology* **8**, 135–170.

Wright, H.E., Jr., J.E. Kutzbach, T. Webb, III, W.F. Ruddiman, F.A. Street-Perott, and P.J. Bartlein, eds. (1993) *Global Climates Since the Last Glacial Maximum*. University of Minnesota Press, Minneapolis.

Chapter 2

Abrupt Climate Change around 2,650 BP in North-West Europe: Evidence for Climatic Teleconnections and a Tentative Explanation

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2.1. Introduction

Natural variations in atmospheric ^{14}C, which are expressed as wiggles in the radiocarbon calibration curve (Stuiver *et al.*, 1993) severely limit the possibilities for fine-resolution dating of changes in vegetation and climate as recorded in lake deposits and raised bogs (*e.g.* van Geel, 1978; Magny, 1993a; Barber *et al.*, 1994). Van Geel and Mook (1989) stressed the importance of ^{14}C Wiggle-Match Dating (WMD) of organic deposits, given that WMD can reveal relationships between ^{14}C variations and short-term climatic fluctuations caused by solar and/or geomagnetic variations. Kilian *et al.* (1995) have shown that, by using this strategy, raised-bog deposits in particular can be dated more precisely. Hence the raised-bog archive can be compared effectively with other proxy data archives, the more so because WMD has shown that an unexpected ^{14}C reservoir effect plays a role in dating raised-bog deposits (individual conventional radiocarbon dates appear to be 100 to 250 years too old). Wiggle-matching is an elegant way of identifying this effect, and of estimating its magnitude.

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Moreover, Kilian *et al.* (1995) showed that the sharp rise of $\Delta^{14}C$ between 850 and 760 calendar years BC ($\Delta^{14}C$ represents the changing radiocarbon content of the atmosphere as calculated from the dendrochronological ^{14}C-calibration data; see Figure 2.1), appeared synchronous with the transition from the often highly decomposed 'Older *Sphagnum* Peat' to the less decomposed 'Younger *Sphagnum* Peat' at the Sub-boreal/Sub-atlantic transition in NW European raised bogs. This change in the decomposition and also species composition of raised bogs represents one of the most clearly-defined climate shifts during the Holocene; and was used by Blytt and Sernander in their classical division of this epoch. The Sub-boreal was interpreted as representing a dry and often warm period and the Sub-atlantic as a humid and, especially at the onset, a cold episode.

Van Geel *et al.* (1996) have illustrated the succession of peat forming mosses in a core from the raised bog Engbertsdijksveen and have detailed a correspondence between the changing moss composition, the fluctuations in the pollen curve of *Corylus avellana* (Hazel) and the rise in the ^{14}C content of the atmosphere around 800 cal BC. Van Geel *et al.* (1996) have also combined palaeoecological evidence with archaeological information for the impact of that climate change on human populations in the Netherlands. They postulated that a climate transition around 2,650 BP (equivalent to *ca.* 800 calendar years BC; see Fig. 2.1) was the most plausible reason for: (1) the abandonment of Late Bronze Age settlements in marginal areas in the northern Netherlands, and, (2) the colonization of coastal salt marsh areas in Friesland and Groningen by the displaced populations. They also considered the evidence for a synchronous climate change from elsewhere in Europe or further afield. In the present paper a shortened version of the evidence for climate change around 800 cal BC will be given and, supplementing the publication by van Geel *et al.* (1996), a tentative palaeoclimatological explanation for the recorded changes will be presented.

2.2. Paleoecological and Archaeological Evidence for Climate Change around 2,650 BP in The Netherlands

2.2.1. THE RAISED-BOG DEPOSIT ENGBERTSDIJKSVEEN-I (EASTERN NETHERLANDS)

A comprehensive set of palaeoecological data from a study of the site Engbertsdijksveen-I was presented by van Geel (1978). Selected data from this site (covering the changing representation of peat-forming mosses (*Sphagnum* species), the pollen curve of *Corylus*, and stable isotope records) have been presented more recently by van Geel *et al.* (1996). Also WMD has been used to relate the data for the period 1,000 to 500 cal BC to the fluctuations of $\Delta^{14}C$. The most important climate shift took place around 800 cal BC (*ca.* 2,650 BP): *Sphagnum rubellum* shows a decline while *S. papillosum* and subsequently *S. imbricatum* become important peat formers.

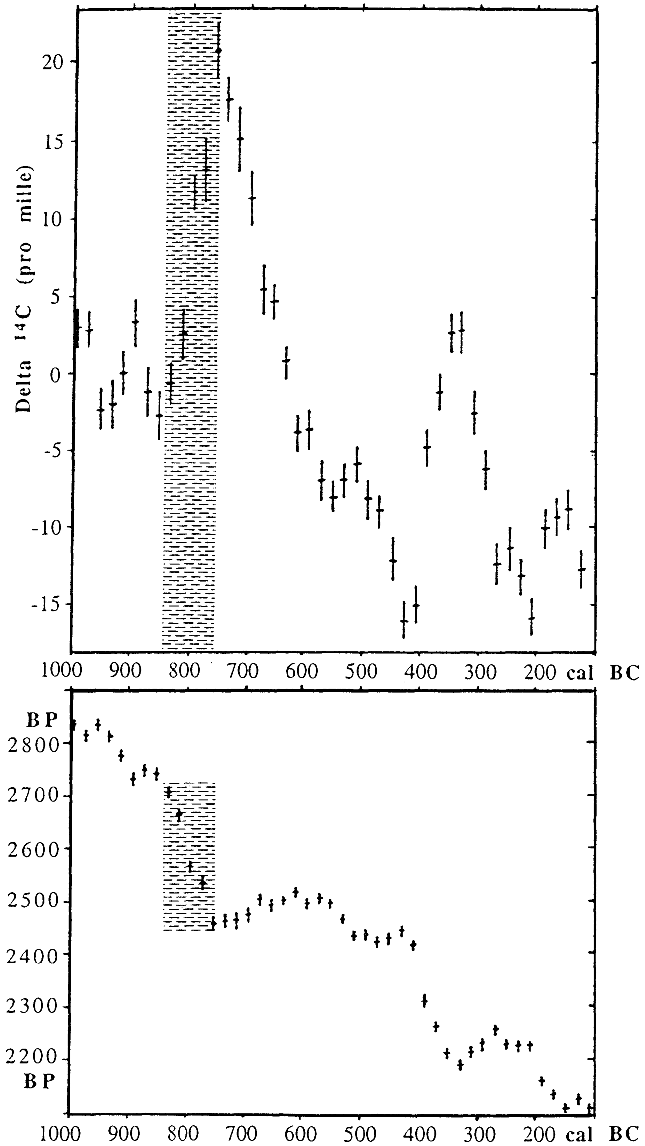


Figure 2.1. The ^{14}C-calibration curve (below) with corresponding fluctuations in Δ^{14}C (above) for 1,000 to 100 calendar years BC. Based on Stuiver *et al.* (1993). Radiocarbon dates for the palaeo-ecological, geological and archaeological evidence for worldwide climate change locate a major transition within the period shown by horizontal hatching. This period is around 2,650 BP (between ca 850 and 760 cal BC). It is characterized by a considerable rise in the ^{14}C-content of the atmosphere (Δ^{14}C). This rise of Δ^{14}C is shown in the upper part of Fig. 2.1. A second considerable rise is observable shortly after 400 cal BC. Apparently, however, this event did not cause climate change, presumably because thresholds were not passed again.

S. imbricatum in particular has a preference for oceanic conditions with a high atmospheric humidity. From the data of the site Engbertsdijksveen-I, it is evident that an abrupt climatic deterioration coincided with the sharp rise in $\Delta^{14}C$ between *ca.* 850 and 760 cal BC.

2.2.2. THE END OF LATE BRONZE AGE SETTLEMENTS IN MARGINAL AREAS IN THE NETHERLANDS; THE BEGINNING OF HABITATION IN THE COASTAL MARSHES OF THE NORTHERN NETHERLANDS

2.2.2.1. *West-Friesland*

Tidal activity ceased about 3,500 BP in this NE part of Noord-Holland as a consequence of the closure of a tidal inlet. Meanwhile West-Friesland became attractive for farmers who colonized the area around 3,350 BP. Extensive archaeological investigations have been undertaken on settlement sites, with detailed archaeobotanical and archaeozoological studies being carried out in order to gain information on husbandry and environment. During the later habitation period, for which 13 radiocarbon dates are available (ranging from 2,760–2,620 BP), people adapted to the increasing wetness of the region by building their houses on dwelling mounds (*terpen* in Dutch). However, the settlement areas eventually became so wet that no further adaptations were possible. So the area was abandoned shortly after 2,620 BP and not reoccupied until medieval times.

There is archaeological evidence for a rise of the water table during the later habitation period of about 45 cm. The ^{14}C dates for an accelerated rise of water table fall between 850 and 800 cal BC. Apart from the archaeological evidence (van Geel *et al.*, 1996), reference is also made to ample palaeoecological evidence for an accelerated water table rise in West-Friesland during this period. It is realized, too, that the period of the *terpen*-phase was contemporaneous with the beginning of a rapid increase in ^{14}C content of the atmosphere. Moreover the indications of increasing wetness and the final abandonment of the area match the abrupt climatic change as recorded in raised bog deposits.

2.2.2.2. *Friesland, Eastern Part*

The former raised bog Fochtelooërveen is situated in the eastern part of the province of Friesland (N. Netherlands), on a relatively low part of the sandy pleistocene area. The western part of the former bog complex has been used for fuel completely, and it is in this formerly peat-covered area that archaeological evidence for settlement sites is present. According to the data, the habitation ended during the transition from the Bronze Age to the Iron Age (Fokkens, 1991). Klaver (1981) studied a peat column from the nearby Fochtelooër raised bog deposit. The samples were taken and stored in metal boxes. So in 1995 extra material could be taken from the column to apply ^{14}C AMS (Accelerator Mass Spectrometry) wiggle-match dating. The dating results of a series of five contiguous charcoal and peat samples on top of the sandy subsoil (van Geel *et al.*, submitted) show that peat growth started here (like in the area of West-Friesland, see above) around 2,690 BP. Also in the area of the Fochtelooërveen a sharp rise of the water table, as a consequence of climate change, caused a loss of formerly cultivated land.

2.2.2.3. *Zwolle-Ittersumerbroek*

At Zwolle-Ittersumerbroek in Overijssel, near the river Ijssel, a Late Neolitic/Bronze Age/Early Iron Age settlement area was excavated on a relatively low coversand plateau (Waterbolk, 1995a,b). Here, too, the end of local settlements was caused by a rise of the water table with related peat growth and the deposition of clays and sands by the river Ijssel. The youngest radiocarbon dates of the settlement (2,670 ± 35 BP, 2,600 ± 30 BP and 2,540 ± 30 BP) indicate that the area became uninhabitable (apparently as a consequence of impeded drainage) during the period of sharp increase of $\Delta^{14}C$ (compare Figure 2.1) and abrupt climate change.

2.2.2.4. *Colonization of the Salt Marshes*

The loss of cultivated land in Pleistocene sandy areas (those which were already marginal from a hydrological point of view) caused depopulation. A causal relationship between this depopulation and the colonization of the salt marshes in the northern Netherlands was already mentioned by Waterbolk (1959, 1966). Arguments for the migration to the salt marsh areas were based on archaeological evidence: pottery of the Ruinen-Wommels type was found in both areas. Radiocarbon dates of the earliest settlements in the salt marsh area are particularly important. The start of the settlement at Middelstum (Boersma, 1983) is dated 2,555 ± 35 BP, which indicates that the earliest colonization occurred during the period of climate change, when $\Delta^{14}C$ showed a steep rise.

The colonization of the salt marsh area was not only related to the above-mentioned environmental changes in the adjacent Pleistocene areas. Earlier migration had not been possible because the salt marshes emerged for the first time around 2,650 BP. Van Geel *et al.* (1996) postulated that climate change, and the contemporaneous slowing in sea level rise, were related to thermal contraction of the upper layer of the ocean (for the phenomenon of thermal expansion, see Mörner, 1995 and Wigley and Raper, 1993) and/or of reduced velocity and pressure on the coast by the Gulf Stream. Moreover, after this climate transition in the temperate zones more water will have accumulated in glaciers, as well as in the soil and in fens and bogs.

2.3. Evidence for Climate Change Elsewhere in Europe and on Other Continents

So decisive a climatic change cannot have been restricted to the Netherlands and adjacent parts of Europe. Indeed, major changes in the radiocarbon content of the atmosphere occurred worldwide and the sharp rise of the ^{14}C-content of the atmosphere between 850 and 760 calendar years BC appears synchronous with abrupt climate transitions elsewhere (van Geel *et al.*, 1996; submitted). In consequence, sudden and important environmental changes took place, which affected plants, animals and also people. These were particularly pronounced in areas where thresholds were passed, those areas having already been marginal—*e.g.* from a hydrological point of view. Apart from alterations in vegetation composition, the effects of climate deterioration were also recorded in tree rings, and in a range of geomorphological activity including glacier advances, avalanches, landslides in mountainous areas and rising lake levels. Radiocarbon dates on these various lines of evidence seem to span several centuries, but this is probably a consequence of the rapidly increasing ^{14}C-content of the atmosphere during the period concerned. On the calendar time scale the period of radiocarbon ages between *ca.* 2,750 BP and 2,450 BP lasted only for *ca.* 90 years (*viz.*, from *ca.* 850 to *ca.* 760 calendar years BC; see Figure 2.1).

2.3.1. EVIDENCE FROM EUROPE

Leuschner (1992) has interpreted germination, dying off and growth reduction phases in bog oaks from Ostfriesland (N.W. Germany) as due to wetter and cooler conditions. From 855 to 835 calendar years BC, a clear reduction in growth is apparent.

According to Godwin (1975), the more or less dry surfaces of raised bogs and blanket bogs became rapidly waterlogged at the Sub-boreal/Sub-atlantic transition as a consequence of climatic deterioration. From peat-stratigraphical data in raised bogs Barber (1982) concluded that a catastrophic decline to a cooler and/or wetter climate occurred around 2,850–2,550 BP. Charman (1990, 1995) interpreted the stratigraphic change of blanket mires at *ca.* 2,700 BP in northern Scotland as expressive of increased surface wetness in response to climate change.

A Bronze Age settlement phase on Stannon Down, in Cornwall, England (Mercer, 1970) was soon succeeded by the formation of peat on the hut floors and field surfaces. Although radiocarbon dates were not available, the author interpreted the abandonment of the settlement as the result of the changing climate at the Sub-boreal/Sub-atlantic transition.

In Carbury Bog, Ireland, a sharp transition from 'dry' peat, rich in Ericaceae, to *S. imbricatum*-peat was dated by ^{14}C AMS wiggle-matching to around 800 cal BC (van Geel *et al.*, submitted).

According to Ballantyne (1993) radiocarbon dating of organic matter buried by solifluction lobes in the Scottish mountains indicate an accelerated downslope movement around 2,500 BP. Similar evidence is given by Matthews *et al.* (1993).

In Bridge *et al.*'s (1990) records of fossil *Pinus* from Scottish blanket bogs, there are no data for the period between 2,670 and *ca.* 1,800 BP. Apparently the tree line was considerably lower then.

Similar results were obtained by Eronen and Huttunen (1993) in Fennoscandia where there are few dates on subfossil *Pinus* between 2,600 and 2,000 BP. Records of glacier activity also point to a climate deterioration after 3,000 BP (Karlén, 1993).

Blikra and Nemec (1993) recorded snow avalanching in Western Norway around 2,600 BP, while Nesje (1993) found evidence for an expansion of glaciers.

There is archaeological and palynological evidence that the Halne area (Hardangervidda Plateau, Southern Norway) was used for grazing by domesticated animals between 4,900 and 2,800 BP (Moe *et al.*, 1988). Yet for the period from 2,800 to 2,200 BP there are no indications of such grazing, in our opinion most probably as a consequence of a change of climate.

According to Berglund (1991) groundwater levels in southern Sweden rose around 2,750 BP, which caused an extension of fenlands. At Lake Bjäresjö, where a peat deposit began to form in the early Holocene an abrupt rise in the water table during the Late Holocene is reflected in a transition from peat to limnic deposits (Gaillard and Berglund, 1988). The top of the peat was dated 2,680 ± 50 BP and 2,690 ± 60 BP (evidently contemporaneous with the sharp rise in $\Delta^{14}C$, compare Figure 2.1).

In the Alps there is evidence for a lowering of the upper forest limit around 2,800 BP (Burga, 1993). Indications of an increase of solifluction in the Alps are described by Gamper (1993) and Veit (1993). The largest glacier in the Italian Alps is Ghiacciaio dei Forni. Orombelli and Pelfini (1985) dated its maximal Holocene extension as 2670 ± 130 BP.

Magny (1993a,b) presented records of Holocene lake-level fluctuations and glacier movements from the Jura and French Subalpine regions. He found chronological correlations between these fluctuations and the atmospheric ^{14}C record (lake transgressions and glacier advances were in phase with ^{14}C maxima) and he suggested that the short-term ^{14}C variations are an empirical indicator of Holocene palaeoclimates. At 800 cal BC a major climatic deterioration was recorded in the lakes. Dendrochronological dates from lake-shore settlement sites in the area north of the Alps indicate that no sites were in use and no new sites were occupied for several centuries after 850 cal BC (Becker *et al.*, 1985).

According to Piotrowski (1995) the fortified lake-margin settlement Biskupin (Central Poland) was abandoned due to an abrupt climate change (a progressive cooling and a significant increase in humidity), which resulted in a rapidly rising lake level and hence social and economic collapse. Niewiarowski *et al.* (1992) dated the base of the lake sediment on top of the old soil surface in the settlement at 2,540 ± 100 BP. This fits the sharply rising $\Delta^{14}C$ values between *ca.* 850 and 760 cal BC though the age of the sediment may also fall within the ^{14}C-plateau (*ca.* 760 to 420 cal BC). According to dendrochronological data (Wazny, 1994) Biskupin was in use between 747 and 722 calendar years BC, and growth of tree rings was markedly reduced between 750 and 722 calendar years BC. The abandonment of Biskupin appears to have occurred at about the same time as the end of settlement sites in marginal areas in the Netherlands and in the Alpine region.

2.3.2. EVIDENCE FROM NORTH AMERICA

Bhiry and Filion (1996) studied a Holocene peat succession in a dune-swale environment of southern Québec. Macrofossil data indicated that: (1) xerophilous and mesophilous trees (*Tsuga canadensis, Pinus strobus, Betula alleghaniensis*) declined on the dune ridges after *ca.* 3,000 BP, (2) *Larix laricina* dominated the site after *ca.* 3,000 BP until 1,900 BP. Based on radiocarbon dates and the study of tree rings, it was concluded that the trees reached old age only until *ca.* 2,600 BP, (3) *Sphagnum recurvum* increased abruptly soon after 3,000 BP and has remained abundant until the present. The *Larix* maximum in old trees (*ca.* 2,600 BP) and the following sharp decline corresponds with the period of sharply rising $\Delta^{14}C$ (Figure 2.1). Hence in southern Québec also marked changes of vegetation and climate appear to have been synchronous with that rise.

Moreover, in northern Québec Payette and Gagnon (1985) showed that the deforestation in the forest tundra became increasingly marked after 3,000 BP. This was confirmed by Gajewski *et al.* (1993) from pollen data. Fire-mediated aeolian activity also declined between 2,800 and 2,500 BP as a result of cooler and wetter conditions (Filion *et al.*, 1991). Gelifluction increased worldwide at the same time (Morin and Payette, 1988; Filion *et al.*, 1991).

Smith (1993) described an increase in solifluction activity associated with excessive waterlogging (cold and humid 'Neoglacial' after 2,800 BP) in N. America. Jirikowic *et al.* (1993) and Davis *et al.* (1992) also found evidence for a rapid climate change (the start of a cold and wet period in Nevada and California) coinciding with the maximum in $\Delta^{14}C$ around 750 cal BC, and they suggested a link between these phenomena.

Meyer *et al.* (1992) interpreted a minimum in fire-related debris flows and concurrent stream-terrace formation in Yellowstone Park as the effect of relatively cool and wet conditions at *ca.* 2,700 BP.

2.3.3. EVIDENCE FROM SOUTH AMERICA

Melief (1985) and Salomons (1986) recorded a downward shift of the upper forest limit in the Colombian Andes as a consequence of a change to cool and wet conditions at *ca.* 2,700 BP. Also Kuhry (1988) found evidence for a shift to colder climate at about that time in Colombia.

From palynological evidence Heusser (1990) concluded that the climate became cooler and more humid in subtropical Chile. Lumley (1993; see also Ritchie, 1995) recorded a catastrophic decline of the tree *Pilgerodendron uviferum* at *ca.* 2,600 BP on the Taitao Peninsula in southern Chile though she interpreted this as resulting from an attack by a plant pathogen just after volcanic ash deposition. In a palynological study of a site in the South Brazilian highlands, Behling (1997) recorded an increase of *Araucaria* around 2,800 BP, indicating a cooler and wetter climate. In southernmost subarctic Argentina, an increase of *Astelia pumila* in a cushion bog, dated 2,630 ± 90 BP, is indicative of a sudden decline of temperature and the onset of a wetter climate (Heusser, 1995).

2.3.4. EVIDENCE FROM NEW ZEALAND

McGlone and Moar (1977) report a decline of the frost and drought sensitive small tree *Ascarina lucida* around 2,600 BP on southern North Island and on South Island.

2.3.5. EVIDENCE FROM JAPAN

Tsukada (1967) dated 'recurrence surfaces' (transitions from strongly decomposed peat, formed in relatively dry local conditions, to less decomposed peat formed in wet conditions) in Japanese raised bogs at around 2,650 BP.

2.3.6. EVIDENCE FROM THE CARIBBEAN

Curtis and Hodell (1993) studied isotopes and trace elements in the sediments of a freshwater lake in Haiti. They concluded that the climate was very dry during the interval between *ca.* 2,500 and 1,500 BP.

2.3.7. EVIDENCE FROM TROPICAL AFRICA

From palynological evidence in Cameroun, Reynaud-Farrera *et al.* (1996) recorded a drastic change of the vegetation cover as a consequence of dryness after *ca.* 2,730 BP. This change to relatively dry conditions was also observed elsewhere in the central African rainforest belt (Elenga *et al.*, 1994; Giresse *et al.*, 1994). Shortly afterwards farmers migrated into the area, availing themselves of what was from the human standpoint a regional climatic improvement.

2.4. A Major Fluctuation of $\Delta^{14}C$ and Evidence for an Abrupt, Worldwide Climate Change: Discussion and Possible Explanation for the Mechanism behind the Inferred Climate Change

From radiocarbon measurements of dendrochronologically dated wood, changes in atmospheric ^{14}C-content during the Holocene have been calculated and published as $\Delta^{14}C$-data. There were numerous minor fluctuations and several major changes of $\Delta^{14}C$. One of the most pronounced short-lived increases in the ^{14}C-content of the atmosphere occurred between 850 and 760 calendar years BC as a consequence of which about 300 ^{14}C-'years' passed in less than 100 calendar years.

During that period (*ca.* 2,750 BP–*ca.* 2,450 BP) radiocarbon-'years' were of a very short duration, whereas (as a consequence of a declining ^{14}C-content of the atmosphere), during the following 300 calendar years the ^{14}C-'age' remained at the level of about 2,450 BP (*viz.*, from *ca.* 760 to *ca.* 450 cal BC). This is the so-called 'Hallstatt-plateau' in the calibration curve (see Figure 2.1). Wiggle-matching of series of ^{14}C-measurements in N.W. European raised bog deposits has shown that the sharp rise of the ^{14}C-content of the atmosphere at the Sub-boreal/Sub-atlantic transition corresponds with the interface between the 'Older' and the 'Younger *Sphagnum* Peat'. This stratigraphic change is often clearly recognizable in peat profiles in NW-Europe, and it is generally accepted that it marks the transition from a relatively continental (dry and mainly warm) to a more oceanic climate régime (cooler and wetter).

From archaeological and palaeoecological studies in the Netherlands, it is now evident that a sudden rise in ground water tables made marginal areas unsuitable for Late Bronze Age farming communities. The later radiocarbon dates from drowned settlement sites in West-Friesland, those from incipient peat growth in the eastern part of the province of Friesland as well as from a site near the river Ijssel, are all contemporaneous with the interface between Older and Younger *Sphagnum* Peat in raised bogs and thus with the sharp change in the ^{14}C-calibration curve between *ca.* 850 and 760 calendar years BC. In West-Friesland and other marginal areas depopulation took place around 2,650 BP while new settlements were established in newly exposed salt marsh areas along the northern coast. Environmental stress (a lack of arable and pasture land, harvest problems, inundations, extensions of fenland) was probably an important factor in the movement of communities to the new coastal sites.

We suspect that a climate transition around 800 cal BC might have made a considerable impact on prehistoric people in other regions. In marginal areas in general (*e.g.* regions with a relatively high ground water table or mountainous areas near the climatic limits for cultivation of crops) climate alteration may have had serious negative effects on prehistoric peoples. However, other regions became more suitable for habitation as a consequence of the same transition (*e.g.* new savanna areas in Cameroun, and newly exposed salt marshes in the northern Netherlands).

This notably abrupt climate change around 2,650 BP corresponds to one of a number of Holocene cold events that have been identified by Harvey (1980). He analyzed published proxy climate evidence and found indications for at least three Holocene occurrences of simultaneous cooling in Europe, North America and the Southern Hemisphere, *viz.*, *ca.* 4,700–4,500 BP, *ca.* 2,700–2,300 BP and the Little Ice Age.

Recently, O'Brien *et al.* (1995) drew a close correlation with the above and the glaciochemical time series derived from the GISP2 Greenland Ice core. The concentrations of sea salt and terrestrial dust were relatively high in the ice core during the intervals 6,100–5,000, 3,100–2,400 and 600–0 calendar years BP. In addition, similar increases were found for phases at more than 11,300 cal yr BP (the Younger Dryas) and between 8,000–7,800 cal yr BP. According to the authors, the increases in sea salt and dust are related to relatively cold climate conditions. O'Brien *et al.* further infer that the cool phases are characterized in the North Atlantic region by an intensified polar circulation of the atmosphere, bringing winterlike, stormy weather more often.

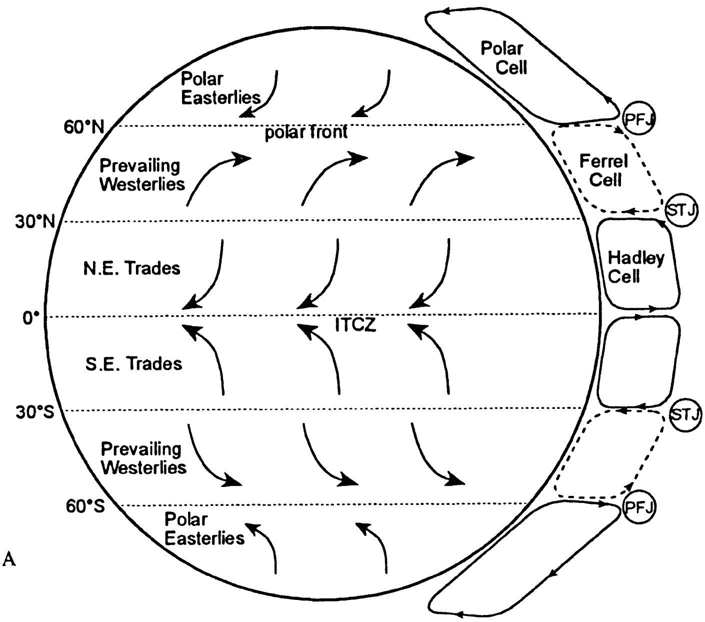
Regarding any cyclical tendency in the increases, the authors suggest that cooler climates recurred at intervals of about 2,600 years. Earlier, a similar oscillation was inferred by Dansgaard *et al.* (1984) using the Camp Century $\partial^{18}O$ Greenland ice core record. Then, as reported by Kerr (1995), Bond found evidence for small increases of ice-rafted debris in the Holocene part of North Atlantic Ocean cores at intervals of 1,000 to 2,400 years, thus possibly coinciding with the quasi-cycle found by O'Brien *et al.* (1995). Bond and Lotti (1995) already found evidence for these quasi-cycles in the Late Pleistocene part (40 to 10 kyr BP) of North Atlantic deep sea cores.

It is interesting to note that, as O'Brien *et al.* (1995) suggested, these phases may correspond to so-called 'triple oscillation' events at which the ^{14}C production rate increased considerably, as reconstructed with the aid of tree rings (Stuiver and Braziunas, 1989; Stuiver *et al.*, 1991). During these 'triple oscillation' events, at least two, but most often three Spörer and Maunder type patterns occurred. The Spörer- and Maunder events were periods (1416–1534 AD and 1645–1715 AD, respectively) during which a minimum number of sun spots were present, thus coinciding with a reduced solar activity (estimated at a 0.4% reduction) and with increases in atmospheric ^{14}C. Stuiver and Braziunas (1989) argue that such century scale $\Delta^{14}C$ variations during the Holocene are best explained by variations in ^{14}C production rate induced by solar change. This conclusion is partly based on the similarity of the ^{10}Be record (*i.e.* an isotope which, like ^{14}C, is produced by interaction of cosmic ray particles with the atmosphere) and the atmospheric ^{14}C record (Beer *et al.*, 1994). Stuiver and Braziunas (1989) note that there are no indications that changes in the ocean circulation caused the discussed ^{14}C variations (*viz.*, by a reduced CO_2 gas exchange at the air-sea interface or a reduced upwelling of ^{14}C deficient deep ocean water). In summary, it is most likely that the triple oscillations in $\Delta^{14}C$ are caused by a reduced solar activity. The triple oscillations were reconstructed at *ca.* 8,500–7,800 cal yr BP, *ca.* 5,400–4,700 cal yr BP, *ca.* 2,680–2,200 cal yr BP and 1,100–400 cal yr BP. Accordingly, a *ca.* 2,500 year quasi-cycle of solar variability seems present in the ^{14}C record in tree rings. As changes in the magnetic-dipole moment follow a cycle of 8 to 10 thousand years (Harvey, 1980), fluctuations in the geomagnetic field can be excluded as a source for the triple $\Delta^{14}C$ variations. A possible correlation between the $\Delta^{14}C$ triple oscillation and the quasi-cycles of cold periods reported by Harvey (1980), O'Brien *et al.* (1995), Dansgaard *et al.* (1984) and others, may indicate that the forcing mechanism behind the cool events is a variation in solar radiation. Such a relation was, in fact, suggested thirty years ago by Bray (1968).

How could a relatively small reduction in solar activity induce the relatively large change in global climate inferred for 2,650 BP? Answering this question involves a considerable degree of speculation, since the effect of solar variability on the Holocene climate is still a controversy (*e.g.* Wigley, 1981; Roederer, 1995). Anyhow, to provide an answer, it is necessary to look at the effect of solar variations on the atmosphere. An important effect of reduced solar activity is an increase in the cosmic-ray flux (Stuiver and Braziunas, 1989). The latter effect is related to the deflection of the cosmic-ray flux by solar-wind plasma in interplanetary space. An increased flux of cosmic rays may in turn produce more ^{14}C in the stratosphere. Two theories are available that explain how relatively small reduction of solar radiation and an accompanying increase in cosmic-ray flux may affect the lower atmosphere.

The first theory is based on the notion that a reduction of (ultraviolet) radiation may also lead to a decline in ozone production in the lower stratosphere (Harvey, 1980). The latter process could be the trigger mechanism responsible for the inferred climate changes, as may be deduced from recent climate modelling studies by Haigh (1994, 1996). She performed simulations with climate models to study the relation between the 11-year solar activity cycles, ozone production and climate change. First, Haigh (1994) used a chemical model of the atmosphere and found that a one percent increase in UV radiation at the peak of a solar activity cycle generated one to two percent more ozone in the stratosphere. Subsequently, Haigh (1996) used this increase as an input in a January climate model experiment. In the simulation, this produced a warming of the lower stratosphere by the absorption of more sunlight. In addition, the stratospheric winds were strengthened and the tropospheric westerly jet streams displaced poleward. Since the position of these jets determines the latitudinal extent of the Hadley cells, this poleward shift resulted in a similar displacement of the descending parts of the Hadley Cells (see also Sadourny, 1994). This in turn led to a poleward relocation of the mid latitude storm tracks. These simulation results of Haigh (1996) were similar (though smaller in magnitude) to observations made by van Loon and Labitzke (1994).

Although the present study deals with solar variations on a different time scale, an effect opposite to the one simulated by Haigh (1996) may have played a role in the discussed climate change around 2,650 BP (800 cal BC). The observed strong increase of atmospheric ^{14}C during the period around 2,650 BP may have been caused by a reduced solar activity as part of the quasi-cycle of 2,500 to 2,600 calendar years, as explained above. Such a reduction in solar activity could also have resulted in a decrease in the stratospheric ozone content. If one assumes that this decrease in stratospheric ozone content leads to an opposite effect to the one simulated by Haigh (1996), a constriction of the latitudinal extent of the Hadley Cell circulation follows. Furthermore, an expansion of the Polar Cells and a realignment of the main depression tracks at mid-latitudes towards the equator may be inferred. This reorganization is illustrated by Figures 2.2A and 2.2B, which show a simplified representation of the tropospheric circulation before (*i.e.* similar to present situation) and after the adjustment (grey arrows denote changes). The mid-



2A

2B

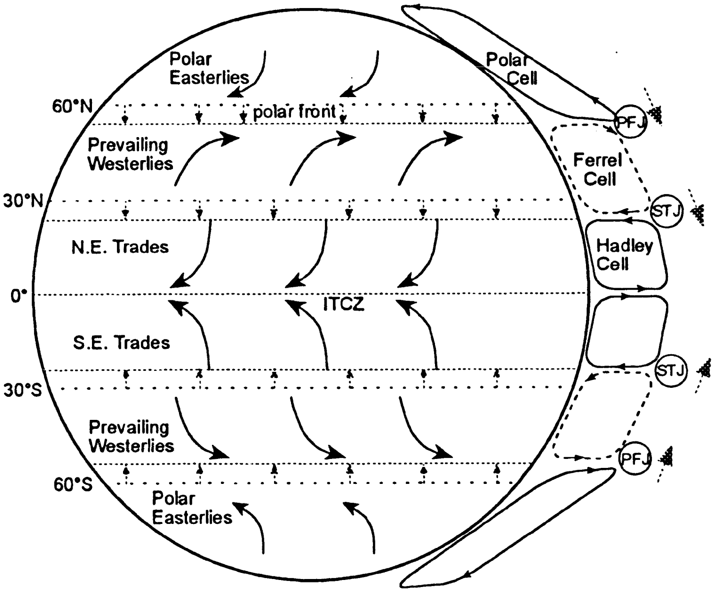


Figure 2.2A. Simplified model of the tropospheric circulation (similar to present situation) before the discussed climate change around 800 cal BC. ITCZ = Intertropical Convergence Zone; STJ = Subtropical Jet; PFJ = Polar Front Jet.

Figure 2.2B. As in Fig 2.2A, but for the period directly after the climate change around 800 cal BC. Grey arrows denote changes corresponding to climate change.

latitudinal depression tracks are not shown but will follow the polar front. In the Northern Hemisphere these tracks are only clearly defined over the oceans. The two jet streams are present in the upper troposphere: the one at the boundary between the Hadley Cell and the Ferrel Cell (STJ, the Subtropical Jet) and the other at the border between the Ferrel Cell and the Polar Cell (PFJ, the Polar Front Jet).

A contraction of the Hadley Cell circulation and an associated weakening of the monsoons would be consistent with inference of drier conditions in the tropics (*e.g.* the Cameroun and Caribbean regions) around 2,650 BP. Similarly, an expansion of the Polar Cells and a shift of the storm tracks closer to the equator could be compatible with the encroachment of cooler and wetter conditions at middle latitudes in both hemispheres (*e.g.* Europe, N. and S. America, Japan, New Zealand). Based on a recent climatic history, Bryson (1975) discussed the relationship between the cooling of high latitudes and a weakening of the monsoons.

The second theory is based on the idea that an increase of the cosmic ray flux may directly lead to an increase in global cloud cover. Evidence for this phenomenon is provided by Svensmark and Friis-Christensen (1997), who found—for the most recent solar cycle—an excellent correlation between variations in cosmic ray flux and the observed global cloud cover. This relationship may be explained by ionization of the atmosphere by cosmic rays, affecting positively aerosol formation and cloud nucleation (Pudovkin and Raspopov, 1992; Raspopov *et al.*, 1997). Tinsley (1996, 1997) suggests that changes in solar activity affect cloud formation through variations in electrically induced cloud nucleation (clouds forming through so-called electrofreezing). An increase in the global cloud cover is believed to cause a cooling of the Earth, especially when low altitude clouds are involved, because more incoming radiation is reflected (Svensmark and Friis-Christensen, 1997). Earlier, Friis-Christensen and Lassen (1991) analyzed for the period 1861–1989 the similarity between the northern hemisphere temperature record and the length of the solar cycle (as an indicator of solar activity), and found a close match. Moreover, it is expected that the effect would be most marked at high latitudes, since the shielding effect of the geomagnetic field is larger near the equator. Indeed, the correlation between cosmic ray flux and cloud cover does become stronger as one goes from the equator towards the poles (Svensmark and Friis-Christensen, 1997). A direct increase in cloudiness and accompanying cooling would be in agreement with the reconstructed wetter and cooler conditions at middle latitudes around 2,650 BP. The inferred drier conditions in the tropics are less easily explained by this second theory. One may speculate, however, that the proposed changes in cloud cover and temperature may induce changes in the atmospheric circulation, possibly involving an increase in the number of El Niño events and drier conditions at several places in the tropics (Svensmark, pers. comm.).

Quite possibly, too, a decrease in solar activity induced adjustments in the ocean circulation through the above inferred changes in the atmosphere. An increase in precipitation at middle latitudes—by a change in the position of the storm tracks or by a direct increase in cloud cover—could well have disturbed the thermohaline circulation in the North Atlantic. This relationship was tentatively postulated by Stuiver and Braziunas (1993) for the Maunder minimum. Through its association with the Gulf Stream, this thermohaline circulation releases significant amounts of heat to the atmosphere at mid-latitudes, contributing to the relatively mild climate of Europe today.

The key to the thermohaline circulation is the formation of North Atlantic Deep Water (NADW). During this process, surface waters sink in the Greenland-Iceland-Norway seas due to a high salt content (and correspondingly higher density) compared to the surrounding water masses. A cessation or weakening of NADW formation through an extra influx of fresh water (melt water; rain water, snow) is proposed by many authors as a mechanism for the cooling of Europe during the Pleistocene glacial events (*e.g.* Broecker, 1992). Modelling studies have shown that the thermohaline circulation may indeed be very sensitive to changes in the fresh water flux (*e.g.* Rahmstorf, 1994). A weakening of the thermohaline circulation would have three major effects. First, it would directly cause a relatively intense cooling of Europe. Second, the cooling could have caused an increase in the area covered by sea-ice and snow, generating further cooling indirectly through the positive ice-albedo feedback. Third, a decrease in sea surface temperature would lead to thermal contraction of the world oceans and a drop in sea level, which is in agreement with the inferred colonization of salt marshes in the Netherlands. This third effect may be inferred from recent studies in which a future warming of a few degrees Celsius is expected to cause a sea level rise through thermal expansion (Mörner, 1995; Wigley and Raper, 1993). In conclusion, around 2,650 BP a weakening of the thermohaline circulation could have amplified the climate change initially initiated by the reduced solar activity. It should be noted, however, that the thermohaline circulation may also be subject to the internal variability of the atmosphere-ocean system, this working with or against fluctuations in solar activity.

The hypothesis enunciated above—involving a weakening of the thermohaline circulation—agrees with the reconstruction of the surface salinity and density for an ocean core at the Rockall plateau (55°N, 14'W) by Duplessy *et al.* (1992). The reconstructed surface salinity and density of the ocean water show a clear minimum around 2,500–3,000 BP. Furthermore this minimum coincides with a small decrease (1°C) in the sea surface temperature.

In addition, a water vapor positive feedback may also have played a role around 2,650 BP, since the inferred cooling could have caused a decrease in evaporation and thus in the water vapor content of the atmosphere. Such a decrease would lead to further cooling, since water vapor is a powerful greenhouse gas. A further factor that possibly influenced the climate at that time is the variation in atmospheric CO_2 content (Harvey, 1980).

In conclusion:

- A sudden and sharp rise in the atmospheric ^{14}C content around 2,650 BP has been found to be contemporaneous with an abrupt and global climate change. At middle latitudes of the Northern Hemisphere (Europe, N. America, Japan) and Southern Hemisphere (New Zealand, S. America), this is to a cooler and wetter climate; and in the tropics (Africa, Caribbean), it is to a drier climate, as evidenced by both archeological and palaeoecological data.
- The variations in atmospheric ^{14}C content and in climate may be tentatively explained by a reduction in solar activity. Two possible mechanisms, which may even have functioned simultaneously, are given here. The first one is based on the idea that a reduced solar input could have reduced the stratospheric ozone content. The latter process may have been the trigger mechanism responsible for a decreased latitudinal extent of the Hadley Cells, an expansion of the Polar Cells and an equator-ward displacement of the mid-latitude storm tracks. The second theory is based on the notion that an increase in cosmic ray flux, accompanying the reduction in solar activity, could have directly caused an increase in global cloud cover through the formation of condensation nuclei. An increase in cloud cover would probably have led to more precipitation and cooler conditions in middle latitudes. Both theories are consistent with the reconstructed climatic changes based on archeological and palaeoecological data.
- A consequent weakening of the thermohaline circulation, as a result of the displacement of the mid-latitude storm tracks, could have been especially significant in the North Atlantic. For such a weakening could have caused a relatively strong cooling of Europe through the reduced release of heat by the Gulf Stream and through the positive ice-albedo feedback. This helps explain, for instance, the particular sensitivity of communities around the Netherlands coast which is where this enquiry began.

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References

Ballantyne, C.K. (1993) Holocene mass movements on Scottish mountains: dating, distribution and implications for environmental change. *In* B. Frenzel (ed.), *Solifluction and Climatic Variation* *in the Holocene.* Special Issue ESF Project 'European Palaeoclimate and Man' Palaeoclimate Research 11, Fischer Verlag, Stuttgart, pp. 71–85.

Barber, K.E. (1982) Peat-bog stratigraphy as a proxy climate record. *In* A. Harding (ed.), *Climatic Change in Later Prehistory,* Edinburgh University Press, Edinburgh, pp. 103–113.

Barber, K.E., Chambers, F.M., Maddy, D., Stoneman, R. and Brew, J.S. (1994) A sensitive high-resolution record of late Holocene climatic change from a raised bog in northern England. *The Holocene* 4, 198–205.

Becker, B., Billamboz, A., Egger, H., Gassmann, P., Orcel, A. and Ruoff, U. (1985) Dendrochronologie in der Ur- und Frühgeschichte—Die absolute Datierung von Pfalbausiedlungen nördlich der Alpen im Jahrringkalender Mitteleuropas. *Antiqua* 11, 1–68.

Beer, J., Joos, F., Lukasczyk, C., Mende, W., Rodriguez, J., Siegenthaler, U. and Stellmacher, R. (1994) ^{10}Be as an indicator of solar variability and climate. *In* E. Nesme-Ribes (ed.), *The Solar Engine and its Influence on Terrestrial Atmosphere and Climate,* NATO ASI series, I 25. Springer Verlag, Berlin Heidelberg, pp. 221–233.

Behling, H. (1997) Late Quaternary vegetation, climate and fire history of the *Araucaria* forest and campos region from Serra Campos Gerais, Paraná State (South Brazil). *Review of Palaeobotany and Palynology* 97, 109–121.

Berglund, B.E. (1991) The Late Bronze Age landscape. *In* B.E. Berglund (ed.), *The Cultural Landscape during 6,000 Years in Southern Sweden,*. Ecological Bulletins 41, 73–77.

Bhiry, N and Filion, L. (1996) Holocene plant succession in a dune-swale environment of southern Québec: a macrofossil analysis. *Ecoscience* (in press)

Blikra, L.H. and Nemec, W. (1993) Postglacial avalanche activity in western Norway: depositional facies sequences, chronotratigraphy and palaeoclimatic implications. *In* B. Frenzel (ed.), *Solifluction and Climatic Variation in the Holocene.* Special Issue ESF Project 'European Palaeoclimate and Man' Palaeoclimate Research 11, Fischer Verlag, Stuttgart, pp. 143–162.

Boersma, J.W. (1983) De opgraving Middelstum-Boerdamsterweg in een notedop. *In* P.B. Kooi, *Leven langs de Fivel: van Helwerd tot Zwart Lap.* In Middelstum-Kantens. Bijdragen tot de plattelandsgeschiedenis met een beschrijving van de boerderijen en hun bewoners. Kantens, 31–35.

Bond, G.C. and Lotti, R. (1995) Iceberg discharges into the North Atlantic on millennial time scales during the last glaciation. *Science* 267, 1005–1010.

Bray, J.R. (1968) Glaciation and solar activity since the fifth century BC and the solar cycle. *Nature* 220, 672–674.

Bridge, M.C., Haggart, B.A. and Lowe, J.J. (1990) The history and palaeoclimatic significance of subfossil remains of Pinus sylvestris in blanket peats from Scotland. *Journal of Ecology* 78, 77–99.

Broecker, W.S. (1992) The strength of the nordic heat pump. *In* E. Bard and W.S. Broecker (eds.), *The Last Deglaciation: Absolute and Radiocarbon Chronologies,* NATO ASI Series, Vol. 12, Springer, Berlin, pp. 173–180.

Bryson, R.A. (1975) The lessons of climatic history. *Environmental Conservation* 2, 163–170.

Burga, C.A. (1993) Swiss alpine palaeoclimate during the Holocene: Pollen analytical evidence and general features. *In* B. Frenzel (ed.), *Solifluction and Climatic Variation in the Holocene.* Special Issue ESF Project 'European Palaeoclimate and Man' Palaeoclimate Research 11, Fischer Verlag, Stuttgart, pp. 11–21.

Charman, D.J. (1990) Origins and Development of the Flow Country Blanket Mire, Northern Scotland, with Particular Reference to Patterned Fens. Ph.D. thesis, University of Southampton.

——. (1995) Patterned fen development in northern Scotland: hypothesis testing and comparison with ombrotrophic blanket peats. *Journal of Quaternary Science* 10, 327–342.

Curtis, J.H. and Hodell, D.A. (1993) An isotopic and trace element study of ostracods from Lake Miragoane, Haiti: A 10,500 year record of paleosalinity and paleotemperature changes in the Caribbean. *In* P.K. Swart, K.C. Lohmann, J. McKenzie, and S. Savin (eds.), *Climate Change in Continental Isotopic Records,* Geophysical Monograph 78, 135–152.

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N., and Hammer, C.U. (1984) North Atlantic climatic oscillations revealed by deep Greenland ice cores. *In* J.E. Hansen and T. Takahashi (eds.), *Climate Processes and Climate Sensitivity*, Geophysical Monograph 29, 288–298.

Davis, O.K., Jirikowic, J. and Kalin, R.M. (1992) Radiocarbon record of solar variability and holocene climatic change in coastal southern California. *In* K.T. Redmond (ed.), *Proceedings of the Eighth Annual Pacific Climate (PACLIM) Workshop, March 10–13, 1991*, California Department of Water Resources, Interagency Ecological Studies Program Technical Report 31, 19–33.

Duplessy, J.C., Labeyrie, L., Arnold, M. Paterne, M., Duprat, J. and Van Weering, T.C.E. (1992) Changes in surface salinity of the North Atlantic Ocean during the last deglaciation. *Nature* **358**, 485–488.

Elenga, H., Schwartz, D. and Vincens, A. (1994) Pollen evidence of late Quaternary vegetation and inferred climate changes in Congo. *Palaeogeography, Palaeoclimatology, Palaeoecology* **109**, 345–356.

Eronen, M. and Huttunen, P. (1993) Pine megafossils as indicators of Holocene climatic changes in Fennoscandia. *In* B. Frenzel (ed.), *Oscillations of the Alpine and Polar Tree Limits in the Holocene*, Special Issue ESF Project 'European Palaeoclimate and Man' Palaeoclimate Research 9, Fischer Verlag, Stuttgart, pp. 29–40.

Filion, L., Saint-Laurent, D., Desponts, M. and Payette, S. (1991) The late Holocene record of aeolian and fire activity in northern Québec, Canada. *The Holocene* **1**, 201–208.

Fokkens, H. (1991) Verdrinkend Landschap, archeologisch onderzoek van het westelijk Fries-Drents Plateau 4400 BC to 500 AD. Ph.D. thesis, University of Groningen.

Friis-Christensen, E. and Lassen, K. (1991) Length of the solar cycle: An indicator of solar activety closely associated with climate. *Science* **254**, 698–700.

Gaillard, M.-J. and Berglund, B.E. (1988) Land-use history during the last 2700 years in the area of Bjäresjö, Southern Sweden. *In* H.H. Birks, H.J.B. Birks, P.E. Kaland, and D. Moe (eds.), *The Cultural Landscape: Past, Present and Future*, Cambridge University Press, Cambridge, pp. 409–428.

Gajewski, K., Payette, S. and Ritchie, J.C. (1993) Holocene vegetation history at the boreal-forest—Shrub-tundra tarnsition in north-western Québec, *Journal of Ecology* **81**, 433–443.

Gamper, M. (1993) Holocene solifluction in the Swiss Alps: Dating and climatic implications. *In* B. Frenzel (ed.), *Solifluction and Climatic Variation in the Holocene*. Special Issue ESF Project 'European Palaeoclimate and Man' Palaeoclimate Research 11, Fischer Verlag, Stuttgart, pp. 1–9.

Giresse, P., Maley, J. and Brenac, P. (1994) Late Quaternary palaeoenvironments in the Lake Barombi Mbo (West Cameroon) deduced from pollen and carbon isotopes of organic matter. *Palaeogeography, Palaeoclimatology, Palaeoecology* **107**, 65–78.

Godwin, H. (1975) *History of the British Flora*. Cambridge University Press.

Haigh, J.D. (1994) The role of stratospheric ozone in modulating the solar radiative forcing of climate. *Nature* **370**, 544–546.

——. (1996) The impact of solar variability on climate. *Science* **272**, 981–984.

Harvey, L.D.D. (1980) Solar variability as a contributing factor to Holocene climatic change. *Progress in Physical Geography* **4**, 487–530.

Heusser, C.J. (1990) Ice age vegetation and climate of subtropical Chile. *Palaeogeography, Palaeoclimatology, Palaeoecology* **80**, 107–127.

——. (1995) Palaeoecology of a Donatia-Astelia cushion bog, Magellanic Moorland-Subarctic Evergreen Forest, southern Tierra del Fuego, Argentina. *Review of Palaeobotany and Palynology* **89**, 429–440.

Jirikowic, J.L., Kalin, R.M. and Davis, O.K. (1993) Tree-ring ^{14}C as a possible indicator of climate change. *In Climate Change in Continental Isotopic Records*. Geophysical Monograph **78**, 353–366.

Karlén, W. (1993) Glaciological, sedimentological and palaeobotanical data indicating Holocene climatic change in Northern Fennoscandia. *In* B. Frenzel (ed.), *Oscillations of the Alpine and Polar Tree Limits in the Holocene*, Special Issue ESF Project 'European Palaeoclimate and Man' Palaeoclimate Research 9, Fischer Verlag, Stuttgart, pp. 69–83.

Kerr, R. (1995) A fickle sun could be altering earth's climate after all. *Science* **269**, 633.

Kilian, M.R., van der Plicht, J. and van Geel, B. (1995) Dating raised bogs: New aspects of AMS ^{14}C wiggle matching, a reservoir effect and climatic change. *Quaternary Science Reviews* 14, 959–966.

Klaver, E.R. (1981) Een Holocene vegetatie successie in het Fochtelooerveen. Internal Report Hugo de Vries-Laboratorium, no. 101, University of Amsterdam.

Kuhry, P. (1988) Palaeobotanical-palaeoecological Studies of Tropical High Andean Peatbog Sections (Cordillera Oriental, Colombia). *Dissertationes Botanicae* 116.

Leuschner, H.H. (1992) Subfossil trees. *In* T.S. Bartholin, B.E. Berglund, D. Eckstein, and F.H. Schweingruber (eds.), *Tree Rings and Environment, Proceedings of the International Dendrochronological Symposium,* Ystad, South Sweden, 3–9 September 1990, *Lundqua Report* 34, 193–197.

Lumley, S.H. (1993) Late Quaternary Vegetational and Environmental History of the Taitao Peninsula, Chile. Ph.D. thesis, Cambridge University.

Magny, M. (1993a) Solar influences on Holocene climatic changes illustrated by correlatons between past lake-level fluctuations and the atmospheric ^{14}C record. *Quaternary Research* **40**, 1–9.

——. (1993b) Un cadre climatique pour les habitats lacustres préhistoriques? *Compte Rendu de l'Académie des Sciences Paris* 316, Série II, 1619–1625.

Matthews, J.A., Ballantyne, C.K., Harris, C. and McCarroll, D. (1993) Solifluction and climatic variation in the Holocene: discussion and synthesis. *In* B. Frenzel (ed.), *Solifluction and climatic variation in the Holocene*, Special Issue ESF Project 'European Palaeoclimate and Man' Palaeoclimate Research 11, Fischer Verlag, Stuttgart, pp. 339–361.

McGlone, M.S. and Moar, N.T. (1977) The *Ascarina* decline and post-glacial climatic change in New Zealand. *New Zealand Journal of Botany* **15**, 485–489.

Melief, A.B.M. (1985) Late Quaternary Paleoecology of the Parque Nacional Natural los Nevados (Cordillera Central), and Sumapaz (Cordillera Oriental) areas, Colombia. Ph.D. thesis, Universiteit van Amsterdam.

Mercer, R.J. (1970) The excavation of a Bronze Age hut-circle settlement, Stannon Down, St. Breward, Cornwall, 1968. *Cornish Archaeology* 9, 17–46.

Meyer, G.A., Wells, S.G., Balling, R.C. and Jull, A.J.T. (1992) Response of alluvial systems to fire and climate change in Yellowstone National Park. *Nature* 357, 147–150.

Moe, D., Indrelid, S. and Fasteland, A. (1988) The Halne area, Hardangervidda. Use of a high mountain area during 5000 years—An interdisciplinary case study. *In* H.H. Birks, H.J.B. Birks, P.E. Kaland, and D. Moe (eds.), *The Cultural Landscape—Past, Present and Future*, Cambridge University Press, Cambridge, pp. 431–444.

Morin, H. and Payette, S. (1988) Holocene gelifluction in a snow-patch environment at the forest-tundra transition along the eastern Hudson Bay coast, Canada. *Boreas* 17, 79–88.

Mörner, N.A. (1995) Recorded sea level variability in the Holocene and expected future changes. *In* D. Eisma (ed.), *Climate Change. Impact on Coastal Habitation*, Lewis Publishers, London, pp. 17–28.

Nesje, A. (1993) Neoglacial gelifluction in the Jostedalsbre region, western Norway: evidence from dated buried soils. *In* B. Frenzel (ed.), *Solifluction and Climatic Variation in the Holocene*. Special Issue ESF Project 'European Palaeoclimate and Man' Palaeoclimate Research 11, Fischer Verlag, Stuttgart, pp. 37–47.

Niewiarowski, W., Noryskiewicz, B., Piotrowski, W. and Zajaczkowski, W. (1992) Biskupin fortified settlement and its environment in the light of new environmental and archaeological studies. *In* B. Coles (ed.), *The Wetland Revolution in Prehistory*. WARP Occasional Paper **6**, 81–92.

O'Brien, S.R., Mayewski, P.A., Meeker, L.D., Meese, D.A., Twickler, M.S., and Whitlow, S.I. (1995) Complexity of Holocene climate as reconstructed from a Greenland ice core. *Science* **270**, 1962–1964.

Orombelli, G. and Pelfini, M. (1985) Una fasa di avanzata glaciale nell'olocene superiore, precedente alla piccola glaziazione, nelle Alpi Centrali. *Rendiconti della Società Geologica Italiana* **8**, 17–20.

Payette, S. and Gagnon, R. (1985) Late Holocene deforestation and tree regeneration in the forest-tundra of Québec. *Nature* **313**, 570–572.

Piotrowski, W. (1995) Biskupin—The fortified settlement from the first millennium BC. *Quaternary Studies in Poland* **13**, 89–99.

Pudovkin, M.I. and Raspopov, O.M. (1992) The mechanism of action of solar activity on the state of the lower atmosphere and meteorological parameters (a review). *Geomagnetism and Aeronomy* **32**, 593–608.

Rahmstorf, S. (1994) Rapid climate transitions in a coupled ocean-atmosphere model. *Nature* **372**, 82–85.

Raspopov, O.M., Shumilov, O.I., Kasatkina, E.A., Dergachev, V.A., and Creer, K.M. (1997) Impact of cosmic ray flux variations caused by changes in geomagnetic dipole moment on climate variability. Russian Academy of Sciences IOFFE Physical-Technical Institute, Preprint 1693, 1–41.

Reynaud-Farrera, I., Maley, J. and Wirrmann, D. (1996) Végétation et climat dans les forêts du Sud-Ouest Cameroun depuis 4770 ans BP: Analyse pollinique des sédiments du Lac Ossa. *Comptes Rendus de l'Académie des Sciences Paris* **322**, série IIa: 749–755.

Ritchie, J.C. (1995) Tansley Review No. 83. Current trends in studies of long-term plant community dynamics. *New Phytologist* **130**, 469–494.

Roederer, J.G. (1995) Solar variability effects on climate. *In* B. Frenzel, T. Nanni, M. Galli, and B. Gläser (eds.), *Solar Output and Climate during the Holocene*, Paläoklimaforschung—Palaeoclimate Research **16**, 1–13.

Sadourny, R. (1994) Sensitivity of climate to long-term variations of solar output . *In* E. Nesme-Ribes (ed.), *The Solar Engine and Its Influence on Terrestrial Atmosphere and Climate*. NATO ASI series I **25**. Springer Verlag Berlin Heidelberg, pp. 479–491.

Salomons, J.B. (1986) Paleoecology of volcanic soils in the Colombian Central Cordillera. *Dissertationes Botanicae* **95**, 1–212.

Smith, D.J. (1993) Solifluction and climate in the Holocene: A North American perspective. *In* B. Frenzel (ed.), *Solifluction and Climatic Variation in the Holocene*, Special Issue ESF Project 'European Palaeoclimate and Man', Palaeoclimate Research 11, Fischer Verlag, Stuttgart, pp. 123–141.

Stuiver, M., Braziunas, T.F., Becker, B. and Kromer, B. (1991) Climatic, solar, oceanic, and geomagnetic influences on Late-Glacial and Holocene atmospheric $^{14}C/^{12}C$ change. *Quaternary Research* **35**, 1–24.

Stuiver, M., and Braziunas, T.F. (1989) Atmospheric ^{14}C and century-scale solar oscillations. *Nature* **338**, 405–408.

Stuiver, M., and Braziunas, T.F. (1993) Sun, ocean, climate and atmospheric $^{14}CO_2$: An evaluation of causal and spectral relationships. *The Holocene* **3**, 289–305.

Stuiver, M., Long, A., and Kra, R.S. (1993) Calibration 1993. *Radiocarbon* **35**, 215–230.

Svensmark, H. and Friis-Christensen, E. (1997) Variation of cosmic ray flux and global cloud coverage—A missing link in solar-climate relationships. *Journal of Atmospheric and Solar-Terrestrial Physics* **59**, 1225–1232.

Tinsley, B.A. (1996) Correlations of atmospheric dynamics with solar wind-induced changes of air-earth current density into cloud tops. *Journal of Geophysical Research* **101**, 29701–29714.

Tinsley, B.A. (1997) Do effects of global atmospheric electricity on clouds cause climate changes? *EOS* **78**, 341.

Tsukada, M. (1967) Pollen succession, absolute pollen frequency, and recurrence surfaces in Central Japan. *American Journal of Botany* **54**, 821–831.

van Geel, B. (1978) A palaeoecological study of Holocene peat bog sections in Germany and the Netherlands. *Review of Palaeobotany and Palynology* **25**, 1–120.

van Geel, B. and Mook, W.G. (1989) High-resolution ^{14}C dating of organic deposits using natural atmospheric ^{14}C variations. *Radiocarbon* **31**, 151–156.

van Geel, B., Buurman, J. and Waterbolk, H.T. (1996) Archaeological and palaeoecological indications of an abrupt climate change in The Netherlands, and evidence for climatological teleconnections around 2650 BP. *Journal of Quaternary Science* **11**, 451–460.

van Geel, B., van der Plicht, J., Kilian, M.R., Klaver, E.R., Kouwenberg, J.H.M., Renssen, H., Reynaud-Farrera, I., and Waterbolk, H.T. The sharp rise of $\Delta^{14}C$ around 800 cal BC: Possible causes, related climatic teleconnections and the impact on human environments, *Radiocarbon* (submitted).

van Loon, H. and Labitzke, K. (1994) The 10–12 year atmospheric oscillation. *Meteorologisches Zeitschrift* **3**, 259–266.

Veit, H. (1993) Holocene solifluction in the Austrian and southern Tyrolean Alps: dating and climatic implications. *In* B. Frenzel (ed.), *Solifluction and Climatic Variation in the Holocene.* Special Issue ESF Project 'European Palaeoclimate and Man', Palaeoclimate Research 11, Fischer Verlag, Stuttgart, pp. 23–32.

Waterbolk, H.T. (1959) Nieuwe gegevens over de herkomst van de oudste bewoners der kleistreken. *Koninklijke Nederlandse Akademie van Wetenschappen, Akademiedagen* **11**, 16–37.

Waterbolk, H.T. (1966) The occupation of Friesland in the prehistoric period. *Berichten van de Rijksdienst voor het Oudheidkundig Bodemonderzoek* **15/16**, 13–35.

Waterbolk, H.T. (1995a) The Bronze Age settlement of Zwolle-Ittersumerbroek: Some critical comments. *Palaeohistoria* **35/36**, 73–87.

Waterbolk, H.T. (1995b) De prehistorische nederzetting van Zwolle-Ittersumerbroek. *Archeologie en Bouwhistorie in Zwolle* **3**, 123–173.

Wazny, T. (1994) Dendrochronology of Biskupin—Absolute dating of the Early Iron Age settlement. *Bulletin of the Polish Academy of Sciences, Biological Sciences* **42**, 283–289.

Wigley, T.M.L. (1981) Climate and paleoclimate: What can we learn about solar luminosity variations? *Solar Physics* **74**, 435–471.

Wigley, T.M.L. and Kelly, P.M. (1990) Holocene climatic change, ^{14}C wiggles and variations in solar irradiance. *Philosphical Transactions of the Royal Society,* London A **330**, 547–560.

Wigley, T.M.L. and Raper, S.C.B. (1993) Future changes in global mean temperature and sea level. *In* R.A. Warrick, E.M. Barrow, and T.M.L. Wigley (eds.), *Climate and Sea Level Change,* Cambridge University Press, Cambridge, pp. 111–133.

Chapter 3

Medieval Climatic Anomaly in the Americas

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3.1. Introduction

Discussions of future climate, whether in forums of science or policy, have come to be dominated by the debate over human-induced global warming. Typically, these discussions focus on the temperature departures that might result from continued anthropogenic emission of greenhouse gasses; on the response of the earth's physical and biotic systems to those departures; and on the socio-economic impacts of those responses. Strong differences of opinion fuel these discussions. Many scientists, citing computerized simulations of an atmosphere with increased heat-trapping capacity, conclude that significant global warming is likely in the future, and perhaps already underway. Others, stressing the uncertainties of the computer models and questioning the assumptions on which the simulations are based, proclaim that predictions of human-induced global warming are premature, unscientific, and alarmist. The socio-economics of the debate are no less polarized. One side warns that failure to curtail fossil-fuel emissions could threaten the food and water supply of the Earth's burgeoning human population; the other raises the specter that restricting fossil fuel consumption might lead to global economic collapse.

Preoccupied with the greenhouse factor and reduced to a polemic, these discussions of future climate too often overlook a critical inevitability: *with or without anthropogenic warming, the Earth's climate will change appreciably in the years ahead, just as it has on many occasions during the late Holocene.* A large and growing body of evidence documents the decades- to centuries-scale climatic variations that have characterized the latter portion of the Holocene epoch, and indicates that not a few of these were severe and widespread enough to have caused significant environ-

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mental modification and social strain. If such changes in climate were to occur today, the environmental and socio-economic impact could be profound. Yet we too often ignore the possible recurrence of these natural climatic precedents, hypothesizing instead about an unprecedented greenhouse threat. Surely, prudence dictates that societies study, plan for, and as far as possible prevent, anthropogenic global warming. But that same prudence should require that we consider the consequences of, and plan for, significant changes in temperature and precipitation such as those that have actually and naturally occurred in the recent past.

Such an assessment must begin by identifying from the recent geological record the secular- to century-scale intervals when climate differed substantially from that of the modern (twentieth-century) period, and by characterizing these climatically anomalous intervals in terms of duration, areal extent, and associated circulation patterns. This paper addresses one such episode, the so-called 'Medieval Climatic Anomaly' (Stine, 1994), and its expression in seven regions of the Americas: California and the northwestern Great Basin; the northern Rocky Mountains and the adjacent Great Plains; the Upper Midwest and eastern sub-arctic Canada; southern Alaska and western Canada; southern Patagonia; the Yucatan Peninsula; and the south-tropical Andes. In five of these regions the late Medieval period brought prolonged intervals of dryness; the Upper Midwest and Alaska/British Columbia, in contrast, experienced a much wetter regime.

This review places particular emphasis on intermontane North America, where a relatively dense network of study sites has become available during the past decade. It does not cover the voluminous literature concerning Medieval climate in the southern Rocky Mountains and on the Colorado Plateau. This mass of literature derives from an interest in the role that climate may have played in the rise and demise of the Anasazi people during this period. Detailed proxy records of climate are available (see, for example, Euler *et al.*, 1979), and the apparent link between meteorological/ climatological events and cultural changes is to many workers compelling. But the extent to which these climatic conditions were peculiar to the Medieval period is questionable (Dean, 1994). It is therefore assumed, for the purpose of this paper, that the southern Rocky Mountains and the Colorado Plateau did not experience so singular a climatic anomaly in medieval times.

3.2. Background

3.2.1. CONTEXT, PREMISES, SCOPE, AND APPROACH

In a compilation of evidence published more than three decades ago, H.H. Lamb (1963), pursuing a theme raised by Huntington (1924) and Brooks (1926), presented a compelling case for a widespread climatic departure during the Middle Ages. In this and subsequent works (Lamb, 1965, 1977) he demonstrated that in

Greenland, Iceland, and portions of northern Europe temperatures began to rise around AD 800, and that within two centuries they exceeded average twentieth-century values by from one to several degrees centigrade. Excepting a possible cool interlude sometime between AD 1050 and 1150, this 'Medieval Warm Epoch' persisted until around AD 1300.

Lamb's catalogue of evidence from Medieval times extended beyond the North Atlantic. He concluded, for example, that by twentieth-century standards south-western Australia was dry; the Sahara was moist; and eastern China and Japan were cool. Building on Bryson's (1963) evidence of a 'Little Climatic Optimum' in North America, Lamb reported that the latter half of Medieval time brought warm, moist conditions to the Upper Midwest, and warmth to the arctic regions of Alaska (Fritts, 1963) and Canada (Bryson, 1963, Bryson *et al.,* 1965; Nichols, 1967, 1970). Detailed records of Medieval climatic conditions from elsewhere in the Americas were rare, though Southern Patagonia was believed to have been dry (Auer, 1958, 1960), and the Yucatan Peninsula was understood to have experienced great, culturally decisive swings in moisture availability (Brooks, 1926). The pollen diagrams of Heusser (1966) suggested to Lamb that any Medieval warmings in British Columbia and southern Chile, at latitudes comparable to northern Europe, were unremarkable 'if they were felt at all'.

The concept of a 'Medieval Warm Epoch' has stimulated a vibrant, interdisciplinary interest in, and a widespread search for, proxy records of climate for the ninth through fourteenth centuries. Especially within the last decade, this search has turned up evidence of anomalous climatic conditions in many regions of the Americas, though most of this newfound information points to altered precipitation patterns, rather than temperature variations *per se.* With this in mind, the present paper concerns itself not with a 'Medieval Warm Epoch', but with a 'Medieval climatic anomaly'.

To understand the nature and consequences of this anomaly it is necessary to know not only what the climate changed to, but what it changed from. It appears that, in many regions of the Americas at least, the climate that forms the backdrop for the Medieval anomaly became established sometime between AD 0 and 300. This pre-anomaly condition varied in character from region to region. In discussing the individual regions an attempt has been made to place the climatic anomaly of Medieval time into this longer context.

3.2.2. DATING THE CLIMATIC ANOMALY: SOME PROBLEMS OF RECORD-TO-RECORD AND REGION-TO-REGION COMPARISON

Several different techniques, and combinations of techniques, have been used in the studies cited herein to date the climate changes of Medieval time. Because these techniques differ in precision, accuracy, and resolution, site-to-site and study-to-study comparisons of the commencement times, termination times, and durations of the climatically anomalous intervals are problematical. The tree-ring series and ice cores available from the Americas, while extremely accurate chronologically, only rarely extend back through the Medieval period. By comparison, radiocarbon dating, which forms the basis for most of the American chronologies, is far less accurate, even when the individual dates have been calibrated. Indeed, because substantial portions of Medieval time were characterized by high solar activity and, consequently, a low rate of atmospheric radiocarbon production, the radiocarbon ages from the period are often triplicated, and in some cases quadruplicated, on the calibration curve. The result is that even calibrated radiocarbon dates from Medieval time seldom yield ages with a resolution higher than plus/minus a half-century.

Regional comparisons involving radiocarbon-based chronologies are further complicated by site-to-site differences in the field settings. Some sites yield datable features that can be directly and causally linked to the onset (or the cessation) of a climatic anomaly. In such cases, to date the feature is to date the commencement (or termination) of the anomaly. Seldom is it possible to date directly both the commencement and the termination of the anomaly, and thus accurately define its duration. Meanwhile, in other studies (most often those involving sediment cores) it is possible to date directly neither the commencement nor the termination of an anomaly. Instead, they must be interpolated from overlying and underlying radiocarbon dates.

Because of these limitations, and to avoid undue interpretation of the dates, all radiocarbon ages cited herein are given as single, calibrated numbers on the BC/AD scale, without stating the 1-sigma error. It should be understood that these dates represent only broad approximations of actual ages. Given these dating constraints, it is not yet possible to say with confidence whether a climate anomaly inferred from one region was strictly synchronous with one inferred from some other region. It is clear, however, that the anomalous climatic conditions that characterized much of the Medieval period were, region to region, at least broadly overlapping in time, if not strictly synchronous.

3.3. California and the Northwestern Great Basin

3.3.1 RELICT STUMPS AND RELATED EVIDENCE IN SIERRAN DRAINAGES

Several different types of proxy paleoclimatic indicators demonstrate that during two periods of Medieval time—for more than two centuries prior to AD ~1110, and for over a century prior to AD ~1350—California and the northwestern Great Basin experienced conditions substantially drier than those of the modern period. The most conspicuous evidence of dryness consists of the relict stumps of trees and shrubs that stand rooted on the beds of present-day lakes, rivers, and marshes. These plants grew at times of reduced water levels, and were drowned when climate returned to a wetter condition. Seven of the stump sites—Mono Lake, Walker Lake, Owens Lake, the West Walker River, Tenaya Lake, Independence Lake, and Osgood Swamp—are described below.

Mono, Owens, and Walker lakes (Fig. 3.1) are terminal sinks of the western Great Basin. The great bulk of their inflow consists of snowmelt from the east side of California's Sierra Nevada. Like all hydrographically closed water bodies, these lakes fluctuate naturally in response to changes in climate, expanding when inflow exceeds evaporative loss, and contracting when loss exceeds inflow. Because of decades-long diversion of their feeder streams for irrigation and/or municipal supply, all three of the lakes presently lie well below their calculated 'natural levels' (the levels they would occupy but for diversions). Presently Mono stands approximately 16 m below its natural level; Walker lies nearly 40 m below it; and Owens, which under natural conditions would contain a lake approximately 15 m deep, has been reduced to a playa.

Relict stumps of trees *(Pinus jeffreyi, Populous fremontii, Shepherdia argentea)* and shrubs *(Salix sp., Artemisia tridentata, Chrysothamnus nauseosus)* stand rooted in their growth positions at numerous locations on the artificially exposed shorelands of Mono Lake (Stine, 1990, 1994). They are most abundant (approximately one hundred individuals) on the deltas of the main feeder streams and near the spring-fed marshes and meadows on the northern, western, and southern shores. Dozens of the tree stumps, with between 40 and 60 annual growth rings, have basal elevations of approximately 1,951 m (~8 m below today's natural level); shrub stumps with up to 20 rings are common at even lower elevations, and have been found as low as 1,941 m (~17 m below today's natural level).

Radiocarbon dates on outermost wood (thus, 'death dates') from 17 of the Mono stumps distinguish two generations of trees and shrubs, and thus two Medieval lowstands of the lake. The Generation-1 stumps ('G-1 stumps') were killed around AD 1110 when the lake recovered from the first of the lowstands; the G-2 stumps were killed approximately AD 1350 during the second recovery (each of these two generalized dates represents the midpoint of the interval of error-bar overlap—see Stine, 1994). The deltaic sedimentary record exposed in the channel walls of Mono Lake's recently incised feeder streams confirms that the lake was low for many decades prior to AD ~1110; that it rose 19 m beginning around that date; that it returned to low levels during the ensuing decades; and that it recovered from that second lowstand around AD ~1350 (Stine, 1990). Note that the 19-m rise that separates the two drought-induced lowstands represents a very wet, if short-lived, interval. The level attained by the lake during this brief transgression—1960.5 m—is higher than any stand in modern (post-1850) times, and the fourth-highest peak of the past 4,000 years. It thus appears that the Medieval period was not only marked by extreme drought, but by extreme sub-century- to century-scale swings in moisture availability.

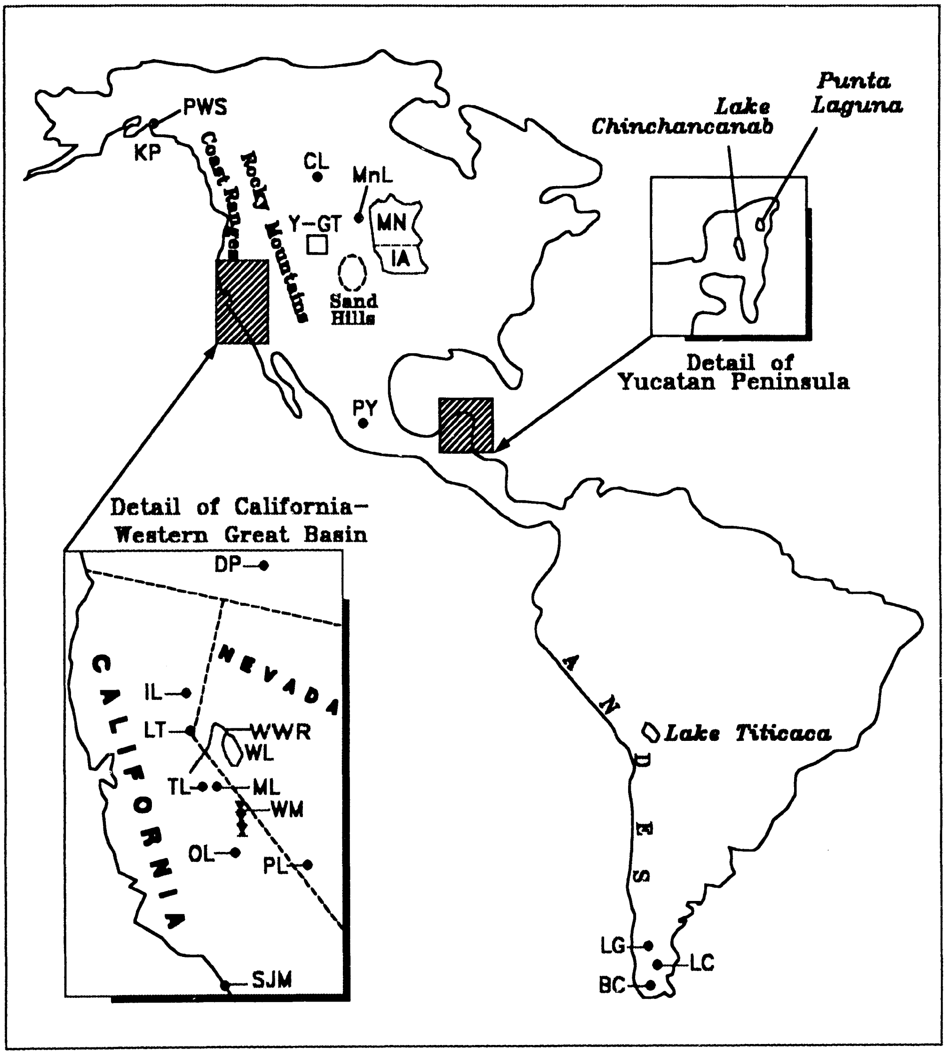

Figure 3.1. Index map showing localities of sites referred to in the text. Abbreviations, California and Western Great Basin: DP=Diamond Pond; IL= Independence Lake; LT=Lake Tahoe; WWR=West Walker River; WL=Walker Lake; TL=Tenaya Lake; ML=Mono Lake; WM=White Mountains; OL=Owens Lake; PL=Pahranagat Lake; SJM=San Joaquin Marsh. Elsewhere in North America: PWS=Prince William Sound; KP=Kenai Peninsula; CL=Chappice Lake; MnL=Moon Lake; Y-GT=Yellowstone-Grand Tetons National Parks; MN=Minnesota; IA=Iowa; PY=Piscina de Yuriria. In Patagonia: LG=Lago Ghio; LC=Lago Cardiel; CM=Catalon Marsh.

At Owens Lake, desiccation due to agricultural and municipal diversions has exposed shrub stumps rooted at elevations as low as 1,084 m—within three vertical meters of the lowest point on the lake floor. Outermost wood from the one specimen thus far sampled provides a calibrated ^{14}C date of AD ~1020. (Note that the death of this stump, which occupies a very low elevation on the Owens Lake floor, does not signal the termination of a drought; rather, it indicates only that Owens Lake was very low in AD 1020, and that it rose slightly at that time.) Also found at low elevations on the artificially exposed Owens playa are assemblages of stone tools. Dr. Robert Bettinger (pers. comm., 1994, 1997) has identified these artifacts from Stine photographs as 'Rose Spring Type', characteristic of the period between AD ~600 and ~1350. The presence of these artifacts, and of the rooted shrub stumps, argues compellingly for the near-desiccation of Owens Lake during at least the first of the Medieval recessions of Mono Lake.

Artificially induced recession has likewise exposed the rooted remains of trees and shrubs at Walker Lake. Wood from one of the trees, collected from the southeastern shorelands in the late 1960s by Dr. Martin Mifflin (pers. comm., 1993), yields a calibrated ^{14}C age of AD ~1025—similar to the death date on the shrub stump from Owens Lake. (The death of this Walker Lake tree, which is found at an elevation more than 40 m below today's calculated natural level, does not represent a drought termination; rather, it represents a slight rise from a very low level around AD 1025.) The age and location of this specimen indicates that, during at least the first of Mono Lake's Medieval lowstands, Walker Lake stood more than 40 m below today's natural level. Intriguingly, a 'drowned forest', presently under approximately 20 m of water, has been discovered by divers near the northern end of Walker Lake (Mr. Glen Bunch, pers. comm., 1995). This forest grew at a time when the lake surface stood at least 60 m below today's natural level. Wood from these trees will be collected during the autumn of 1997. Radiocarbon assays will determine whether they date from the Medieval period, or from a different lowstand of Walker Lake.

Under natural conditions, Walker Lake receives over 90% of its inflow from the combined discharge of two Sierran streams—the West Walker River and the somewhat smaller East Walker River (Fig. 3.1). With evidence that Walker Lake fell to low levels during at least one interval of the Medieval period, one could reasonably hypothesize that the discharge of the West Walker River during that interval must have been substantially lower than it is today. Stumps rooted in the West Walker River Canyon prove this to be the case. At numerous places, the canyon floor is studded, wall-to-wall, with large (to 80 cm diameter), *in situ* stumps of *Pinus jeffreyi*. The bases of these stumps are today fully inundated by the river, even during times of seasonally low flow. Because the canyon bottom is too narrow to permit substantial lateral movement by the stream, and because *P. jeffreyi* cannot tolerate inundation of the entire root system during even brief (two-week) periods of the growing season (Dr. Andrew Leiser, pers. comm., 1994), the presence of the stumps is evidence of greatly reduced flow through the canyon, and thus of greatly reduced inflow to Walker Lake, at some time(s) in the past.

Radiocarbon dates on outermost wood from five of the West Walker River stumps reveal the same two generations of relict vegetation present at Mono—the first (with up to 200+ annual rings) killed around AD 1130, and the second (with as many as 140 annual rings) killed at approximately AD 1340 (Stine, 1994). The stumps of the West Walker Canyon would thus appear to corroborate the conclusions drawn from Walker, Mono, and Owens lakes—that runoff from the Sierra was greatly reduced during substantial intervals of Medieval time.

Unlike Walker, Mono and Owens lakes, Tenaya and Independence lakes are hydrographically open. Lying immediately west and east, respectively, of the Sierran crest, these two water bodies receive runoff from their high-elevation drainages during months of snowmelt, and spill throughout the spring and summer. By early autumn they have typically fallen a short distance below their spillways.

Protruding from the surface of Tenaya Lake are the upright trunks of large trees. Soundings indicate that the bases of these individuals stand in as much as 20 m of water. According to divers who have investigated and photographed the trunks, they appear to be rooted in growth position (Mr. Phil Catarino, pers. comm., 1996). Radiocarbon assays on outermost wood from two of the Tenaya specimens yield ages coeval with the G-1 and G-2 stumps at the previously noted sites (Stine, 1994). At Independence Lake, tree stumps rooted 'well below' lake level date to approximately 1300 AD, near the end of the second of the two Medieval droughts (Lindstrom, 1990). These stumps of the high Sierran lakes would thus seem to point to lowered surface levels, throughout two intervals of Medieval time in the case of Tenaya Lake, and during at least the second of the two intervals in the case of Independence Lake.

In addition to existing in lakes and streams of Sierran drainage, rooted stumps of Medieval age are present in at least one Sierran marshland—Osgood Swamp near Lake Tahoe (Fig. 3.1). Even during multi-year droughts in the modern period (*e.g.* 1928–34; 1987–92), Osgood Swamp has remained perennially sodden, precluding invasion by pines. Dr. David Adam (pers. comm., 1994), who discovered the drowned lodgepole pine *(P. contorta)* stumps and dated outermost wood at around AD 1120 (close to the death dates on G-1 stumps from other localities), interprets their presence to indicate that the marsh was desiccated for a substantial interval of Medieval time.

3.3.2. CAUSE OF FLUCTUATING WATER LEVELS AT STUMP SITES

In attempting to explain the drop in water levels at the stump sites described above it is important to consider all potential causes, geomorphic as well as climatic. For example, episodes of stream piracy, in which a change in a hydrographic divide deflects runoff into a different drainage, might, in principle, account for low flows in the West Walker River, and thus for a lowered Walker Lake. But no evidence of stream piracy can be found in the West Walker River drainage, nor for that matter in any of the main drainages of the Mono, Owens, Tenaya, Independence, or Osgood systems. Alternatively, a change in the spillway elevation might, *a priori,* explain the fluctuations of mountain lakes like Tenaya and Independence. Careful examination, however, reveals no evidence of rises in spillway elevation (due, for instance, to damming by landslide) nor for spillway lowering (due to incision by the outlet streams) at these two water bodies. Tectonically-induced tilting might conceivably explain the drowning of stumps at hydrographically closed lakes. But at Mono the demonstrable lack of substantial tilting on a 3,800-year-old shoreline (Stine, 1987) and the presence of stumps along three quadrants of the lake margin, argue against the tilt hypothesis. At Owens Lake the presence of rooted stumps not just on the playa edge but near its center, likewise argues against tilt as an explanation for the growth and death of the Medieval-age vegetation. Besides, at both Mono and Owens, as well as at Walker Lake, an implausibly large amount of tilting since Medieval times would be required to account for the elevational differences between the drowned stumps and today's natural lake levels.

In the case of Tenaya and Independence lakes it is conceivable that avalanches or other mass-wasting events plucked the trees from the surrounding slopes and carried them to the water. There, one could surmise, large rocks entwined in the root systems forced the trees to settle, roots down, on the lake bottom. While such an explanation cannot be disproven, it seems unlikely, since inspection shows the dated stumps to have a thoroughly normal posture. Moreover, the ages of the stumps at these two sites (G-1 and G-2 at Tenaya; G-2 at Independence) correspond very closely with those at the other localities. Indeed, the apparent site-to-site contemporaneity of drowned vegetation at seven different stump localities spanning a 400 km-long swath of eastern California and the western Great Basin would seem to require a single regional explanation, rather than a number of different locality-specific explanations. Effective drought (due to reduced precipitation and/or to increased evapotranspiration) is not only the most plausible explanation for lowered water levels at each and every site, but it is the only explanation that accounts for this site-to-site correspondence (Stine, 1994).

3.3.3. DURATION OF DROUGHTS INFERRED FROM THE STUMP SITES

Minimum durations for the Medieval droughts can be determined by counting the number of annual rings present in the longest-lived of the drowned stumps. At the West Walker River, stumps of Generation 1 exhibit over 200 rings; those of Generation 2 contain up to 140 rings. (The life-spans of the submerged Tenaya Lake trunks were only slightly shorter than these figures.) Thus, based on the stump evidence, a broad, radiocarbon-dated chronology for the Medieval droughts is as follows. The first drought commenced prior to AD 910 and terminated approximately AD 1110; the second commenced prior to AD 1210 and terminated approximately AD 1350. Note that the death dates on the low-elevation stumps at Owens Lake (AD ~1020) and Walker Lake (AD ~1025), which signify drought-induced lowstands but not drought terminations, comport with this chronology.

3.3.4. SEVERITY OF DROUGHTS INFERRED FROM STUMPS

Beginning in 1987, California experienced six successive dry years. While the severity of this drought varied from place to place and from year to year, abnormally low precipitation prevailed over the entire Sierra Nevada, with runoff averaging some 65% of the modern normal. This recent dry spell can serve as a convenient datum for judging the severity of the effective droughts of Medieval time. Throughout these six years Osgood Swamp remained a morass, precluding invasion by trees. In each of these years the West Walker River experienced springtime and summertime flows too high, and too prolonged, to permit the establishment or growth of Jeffrey pines on its bed. Both Tenaya and Independence lakes overflowed for at least several weeks during the peak snowmelt period in each of the six drought years, and at no time did either lake fall more than 1 m below its sill.

These observations therefore indicate that the modern dry spell was not severe enough to draw Sierran lakes, rivers, and marshes to levels as low as those of Medieval time. But was its moderate nature simply a function of its short duration? Had the modern dry spell continued at 65% of normal runoff for decades, would water levels at the stump sites eventually have fallen as low as those of Medieval time? The evidence suggests not. Because the drainage basins of these four sites are small, and are dominated by bedrock, they store little groundwater. Runoff following a low-precipitation winter is therefore not augmented appreciably by the discharge of groundwater stored during previous years. Simply put, a longer drought, in and of itself, would not have further lowered runoff. It is low severity, rather than short duration, that makes the modern dry spell an inadequate analog for Medieval climate in the Sierra. Moreover, the high-resolution water-balance model for the Mono Basin developed by Vorster (1985) demonstrates that the 'effective inflow' required to account for the Medieval regressions and lowstands of Mono Lake must have been less than 60% of the modern average value.[1] Likewise, calculations suggest that, to account for the desiccation of Owens Lake during Medieval time, effective inflow to the lake must have dropped to between 45% and 50% of the modern value (Stine, unpub. data, 1995).

The Medieval lowstands of Mono, Owens, and Walker lakes contrast sharply with the levels that would exist today under natural conditions. But this is not to say that the lakes fell to their Medieval lowstands from such high levels. At Mono, at least, the lake surface had occupied moderately low levels (by modern standards) for over 600 years before falling to the first of the extreme lowstands. Numerous other proxy records from California and the Great Basin point to moderately dry conditions during the centuries prior to the profound droughts of Medieval time. These other records further confirm the severity and the persistent nature of the Medieval droughts, and can be used to extend the geographical limits of the droughts northward and southward of the known stump sites.

3.3.5. TREE-RING SERIES FROM THE SIERRA AND WHITE MTS.

Analyzing the rings of long-lived bristlecone pines *(Pinus longaeva)* from the White Mountains of eastern California (Fig. 3.1), LaMarche (1974, 1978) demonstrated that trees growing near the upper timberline are sensitive to temperature (and so put on narrow rings in relatively cool years), while those growing near the lower forest border are more sensitive to moisture availability (and so put on narrow rings in relatively dry years). By sampling trees from both elevational extremes, he was thus able to determine whether a given series of years in the past was warm and wet, warm and dry, cool and wet, or cool and dry. LaMarche's (1974) record from the (moisture-sensitive) lower forest boundary shows that two intervals of Medieval time—from prior to AD 900 until 1100, and from AD 1150 until 1350—were characterized by particularly narrow rings, thus corroborating remarkably well the stump-derived evidence of Medieval droughts.[2] It appears from LaMarche's record that the first of the droughts was accompanied by cool conditions (relative to modern time), while the second, even drier, was characterized by abnormal warmth.

The exceptional longevity (>4,000 years) of the bristlecone pine, coupled with the ability of its wood to resist deterioration in the high, dry, White Mountains, enabled Graybill (pers. comm., 1992) to construct a continuous, 9,000-year ring series from living and dead trees at the lower forest border. Not surprisingly, his record shows the same Medieval droughts as were documented by LaMarche. It also reveals that, during the second of the two, the low-border bristlecones produced some of the narrowest rings of the past nine millennia.

Stable-isotope analyses of bristlecone rings by Leavitt (1994) reflect the dry periods discussed above and, even more strikingly, show the interval AD 1080–1129 to have been very wet, as indicated by a striking depression in $\partial^{13}C$ content. The depressed values suggest to Leavitt that AD 1080–1129 was the wettest multi-decade-scale interval of the past millennium. It seems likely that this period (precisely dated from the dendro-record) corresponds with the dramatic (^{14}C-dated) transgression that separates the two Medieval lowstands of Mono Lake.

Dendroclimatological records produced by Graumlich (1993) from foxtail pines *(P. balfouriana)* and western juniper *(Juniperus occidentalis)* in the southern Sierra Nevada (Fig. 3.1) provide further confirmation of drought during Medieval time. Employing response-surface analysis that relates ring width to historical summer temperature and winter precipitation, she demonstrates that the two driest 50-year intervals in the past 1,000 years (from AD 1250 to 1299, and from AD 1315 to 1364) occurred during the second of the two Medieval lowstands of Mono Lake; the third-driest 50-year interval of the record (AD 1021–1070) occurred during the first of the two Mono lowstands. Her reconstructions show the Medieval period to have been characterized by anomalously high temperatures, with the four warmest 50-year periods of the past millennium occurring between the tenth and fourteenth centuries (in order of decreasing warmth, AD 1118–1167, AD 1245–1294, AD 911–960, and AD 1343–1392).

3.3.6. EVIDENCE FROM MICROFOSSILS AND PACK-RAT MIDDENS

The San Joaquin Marsh is a small wetland that lies adjacent to Newport Bay on California's southern coast (Fig. 3.1). Prior to modification by the Europeans, the marsh was a freshwater environment that received flow from the ephemeral streams that drain the adjoining hills. A core spanning the past ~7,000 years, extracted from the marsh by Davis and his co-workers (Davis *et al.,* 1991), contains pollen from both freshwater and saltwater marsh species, as well as fossil remains of marine foraminifera and dinoflagellates. By Davis' interpretation (Davis, pers. comm., 1997), levels of the core containing relatively large proportions of freshwater pollen represent periods in which climate was wet enough to instigate stream flows to the marsh. In contrast, levels in which marine and saltwater-marsh taxa predominate represent dry periods with little or no stream-water input.

The segment of the core extending from roughly AD 0 to ~1400 is remarkable for its near-lack of freshwater taxa, indicating that climate during this period was too dry to produce substantial runoff from the coastal hills. From this evidence Davis tentatively concludes that Medieval drought in western North America extended at least as far south as Newport Bay. Cores extracted from two southern California coastal wetlands similar to San Joaquin Marsh support this conclusion (Davis, pers. comm., 1997).

Using pollen and packrat middens, the geographical limits of Medieval drought in the Great Basin can be extended at least as far northward as Diamond Pond in east-central Oregon, and at least as far southward as Lower Pahranagat Lake in southern Nevada (Fig. 3.1). Pollen cores extracted from Diamond Pond, and packrat middens found in the area immediately surrounding it, yield a paleo-vegetation record for the past 5,000+ years (Wigand, 1997; Mehringer and Wigand, 1990; Wigand, 1987). The period from AD 0 to 1400, and particularly from AD 850 to 1400, is characterized by a low incidence of juniper *(Juniperus occidentalis),* indicating relatively dry conditions. A high degree of xericity between AD ~850 and ~1400 is also evidenced by an unmatched abundance of *Tetraedon sp.* and other algae species that thrive in stagnant, low-water conditions. Wigand (pers. comm., 1997) concludes that, at least in the vicinity of Diamond Pond, the interval ending around AD 1350–1400 represents the driest centuries-scale period of the past five millennia.

The 2,000-year-long record from Lower Pahranagat Lake likewise shows the Medieval period to have been climatically aberrant (Wigand, 1996). High concentrations of Chenopodiineae and Cyperaceae pollen indicate that shallow water, and large areas of exposed pond-bottom, dominated the interval from AD 850 to ~1400. A deeper lake generally prevailed during the earlier and later parts of the record.

The sites discussed above provide a clear and consistent record of abnormal and persistent dryness during the Medieval period over the northern, western, and southern portions of the Great Basin. No Medieval climatic anomaly has yet been identified in the central or eastern portions of that region, though this merely reflects a lack of evidence either way. But studies eastward, in the Rocky Mountains and the adjacent Great Plains (see below), document persistent drought during Medieval time, suggesting that abnormal dryness may have prevailed over the entire northern Great Basin.

3.3.7. THE ARCHAEOLOGICAL / ANTHROPOLOGICAL RECORD FROM CALIFORNIA AND THE GREAT BASIN

A detailed discussion of the human response to the Medieval climatic anomaly is beyond the scope of this paper. It is noteworthy, however, that archeologists and anthropologists in California and the Great Basin, as well as in other regions, recognize the Medieval period to have been a time of social disruption. Evidence from burial sites in south-coastal California indicates to Raab (1994; see also Arnold, 1992, and Lambert, 1993) that between AD ~1100 and ~1300 war- and disease-induced mortality reached their highest levels in the past 6,000 years. Along the central coast, as well as in California's Central Valley, the second of the two droughts was concurrent with widespread abandonment of long-occupied sites, and with the disintegration of long-established trade networks (Moratto, 1984; Jones et al., 1997). These and other examples of cultural upheaval in western North America are comprehensively reviewed by Jones and others (1997).

3.4. Northern Rocky Mountains and the Adjacent Great Plains

3.4.1. THE ROCK MOUNTAINS OF THE NORTHERN UNITED STATES

At Jenny Lake in Grand Tetons National Park (Fig. 3.1), Dr. David Love and Ms. Betty Strook (Love and Strook, pers. comm., 1996) describe upright trunks of adult trees standing in 24 to 30 meters of water. Divers who have viewed the trees note that at least some of the relicts are characterized by a sparseness of branches on their north-facing sides, much like the trees living on today's shorelands. This, and the presence of a relict raptor nest on the branches of one of the fully submerged individuals, argues against transport from the surrounding slopes (B. Strook, pers. comm., 1996). Outermost wood from one of the specimens has been radiocarbon dated at AD ~1350, virtually coincident with the death of the G-2 stumps in the Sierran drainages. As in those drainages, the presence of stumps in Jenny Lake strongly suggests that water levels were reduced during the decades prior to AD ~1350.

Nearby, in northern Yellowstone National Park (Fig. 3.1), Meyer *et al.* (1995) demonstrate that the interval AD ~650 to ~1250, and in particular the latter 250 years of that period, was characterized by a high incidence of drought-induced sedimentation events triggered by fire. In this same region, Hadley (1996, 1997) provides evidence that the period AD ~750–1250 brought a marked decline in the populations of small-mammal species associated with mesic vegetation. Accompanying this population decline was a significant decrease in the size of the pocket gopher *Thomomys talpoides tenellus,* to its smallest dimensions of the past three millennia. Based on the above-noted population decline, and on a known relationship between xericity and small size in pocket gophers, Hadley concludes that during Medieval time the northern portion of the Rocky Mountains experienced its driest climate of the past 3,200 years.

3.4.2. THE CENTRAL GREAT PLAINS

Recent work on the widespread deposits of aeolian sand to the lee of the Rockies in western Nebraska and eastern Colorado (Fig. 3.1) points to perhaps three major episodes during the past 10,000 years when drought-induced loss of vegetation led to dune destabilization and mobilization. Radiocarbon dating of sand-buried soils, and archeo-stratigraphy, leads Madole (1994) to conclude that the most recent of these major episodes began sometime after AD ~950, but before AD ~1250. Ongoing investigation by Mr. James Swinehart and Dr. David Loope (pers. comm., 1995–97) confirms and clarifies the timing of initial destabilization, and provides evidence that the causal drought may have consisted of two or more datable stages within Medieval time.

3.4.3. THE NORTHERN GREAT PLAINS

Still farther east, on the plains of east-central North Dakota, fossil diatoms in cores extracted from Moon Lake (Fig. 3.1) register dry conditions (relative to the modern period) throughout most of the interval from AD ~0 to ~1250. Changes to particularly salt-tolerant taxa, signaling especially severe drought-induced drawdowns, occurred (based on interpolation) during the intervals AD ~700–750 and AD ~950–1150 (Laird *et al.,* 1996). This is in general accord with the paleobotanical and geochemical investigation by Vance and others (1992) at Chappice Lake on the plains of southeastern Alberta (Fig. 3.1). Analysis of cores spanning the last ~7,300 years shows that during the period between AD ~990 and ~1350 the lake was characterized by inordinately large quantities of seeds and pollen from *Ruppia,* a taxon known for its high salinity tolerance. The sediment deposited in the pond during this interval is also revealing: rather than the bioturbated silty clays that characterize most levels of the core, the Medieval interval displays carbonate-laminated silty clays—an indication that the turbating organisms had been extirpated, or at least rendered rare, by high salinity.

These records of low water levels from the northern Rocky Mountains and the adjacent Great Plains all indicate that substantial segments of Medieval time were characterized by conditions that were abnormally dry, not only by modern standards but by those of the past two millennia. The dating of these droughts, especially at the cored sites that rely on interpolation between radiocarbon dates with large error bars, makes it impossible to correlate precisely the droughts from region to region, or even from site to site. But a large amount of temporal overlap, if not exact concurrence, appears likely.

3.5. The Upper Midwest and Eastern Sub-Arctic Canada

Bryson (1963; Baerreis and Bryson, 1967; Bryson and Murray, 1977) has long maintained that the Upper Midwest was relatively wet for at least several centuries prior to AD 1100, and that drought become frequent and severe only after AD 1200. His evidence consists of palynological and archeological data indicating that, prior to AD ~1100, northwestern Iowa and southeastern Minnesota were characterized by tallgrass prairie and riparian woodland that supported deer. During the ensuing century trees became much less abundant and the woodland-browsing deer gave way to shortgrass-grazing bison.

Bryson recognized early on that the Medieval climate shifts in the Upper Midwest corresponded with shifts in the position of the tundra/treeline border in central and eastern Canada (Bryson *et al.*, 1965). That work, together with refinements by Nichols (1970; 1967) and Sorenson et al. (1972), indicates that the arctic treeline occupied a relatively southerly position, reflecting cold, dry conditions, around BC 100. Shortly thereafter it began to advance northward in response to an increase in summer temperatures. Poleward movement continued until approximately AD 1130, at which time renewed cooling again drove it south. As in the Upper Midwest, the subarctic of central and eastern Canada seems to have been characterized by increased moisture during the centuries prior to AD ~1130 (Nichols, 1970).

The Medieval wettening of the Upper Midwest and eastern sub-arctic Canada contrasts sharply with the contemporaneous drying that beset California, the Great Basin, the northern Rocky Mountains, and the Great Plains. But as is discussed below, this contrast does not constitute a contradiction. On the contrary, the atmospheric circulation that would explain Medieval wetness in the middle and high latitudes of North America's deep interior could also account for contemporaneous dryness farther west. Such a circulation could also explain the increase in glacier-nourishing wetness that appears to have characterized western Canada and southern Alaska at this time.

3.6. Western Canada and Southern Alaska

3.6.1. WESTERN CANADA

Studies of glacio-stratigraphy and glacier-sheared forests at sites in the Coast Range and Rocky Mountains of British Columbia (Fig. 3.1) point to an expansion of glaciers during Medieval time. Luckman's work (1992, 1995) in the Canadian Rockies indicates that an advancing Robson Glacier overrode forest just prior to AD 1150, and sustained its advance until at least AD 1350. The Peyto Glacier was advancing prior to AD 1246, and continued to expand until at least AD 1324. In the Canadian Coast Ranges, Ryder and Thomson (1986) have found that substantial advances of the Klinaklini and Franklin glaciers commenced prior to AD ~1150, and that the Bridge glacier began to extend sometime prior to AD ~1350.

3.6.2. TIDEWATER GLACIERS OF PRINCE WILLIAM SOUND

A Medieval extension of glaciers has likewise been documented in the Prince William Sound region of Alaska's gulf coast (Fig. 3.1). According to Wiles and others (Wiles and Calkin, 1994; Wiles et al., 1995) and to Tuthill and others (1968), a large number of tidewater glaciers (including the Princeton, Sunlight, McCarty, Northwestern, Aialik, Sheridan, and Stellar glaciers, and perhaps the Holgate Glacier) advanced between AD 900 and 1300. As with the Canadian glaciers noted above, the magnitude of this Medieval advance appears to have been exceeded in Holocene times only during the Little Ice Age of recent centuries.

3.7. The Yucatan Peninsula

A topic of speculation for over 70 years (Huntington, 1924), the paleoclimate of the Yucatan Peninsula (Fig. 3.1) has recently been illuminated by Hodel and his coworkers. Analysis of the sediments and oxygen isotopes in cores taken from Lake Chichancanab on the central part of the peninsula (Hodel *et al.*, 1995), and from Lake Punta Laguna slightly farther north (Curtis *et al.*, 1996), reveals a climatically aberrant interval, encompassing Medieval times, in which aridity indicators (gypsum and $\partial^{18}O$ in the Chichancanab record, $\partial^{18}O$ in the Punta Laguna record) attained high values. Together, these records reveal that a dryness unprecedented in the prior thousands of years, and unmatched in the past six centuries, commenced in this portion of the peninsula around AD 300. With the exception of a wet, subcentury interlude centered on approximately AD 1150, this aridity persisted, at Punta Laguna at least, until around AD 1390. Conditions were particularly xeric for multidecade intervals centered on AD 585, 860, 985, 1050, and 1390.

Sites in tropical mainland Mexico, to the west of the Yucatan Peninsula, likewise record drought during major portions of this interval. According to Metcalfe and Hales (1990), cores from Piscina de Yuriria (Fig. 3.1), a maar lake at 20° N Lat., display diatoms and chemical characteristics indicative of high salinity and low water levels in the interval from AD ~500 to ~1110.

Longstanding interest in the Medieval paleoclimate of the Yucatan stems from theories concerning the collapse of the Classical Mayan Civilization around AD 800. The above-cited work by Hodel, Curtis, and their colleagues provides the clearest and best-dated evidence yet that extreme dryness coincided with (and, by implication, may have precipitated) this collapse. A similar climate/culture correspondence suggests to some that the Tiwanaku of the south-tropical Andes may likewise have been a casualty of Medieval drought.

3.8. The South-Tropical Andes

According to Abbott and others (1997), Lake Titicaca (Fig. 3.1) fluctuated widely during Medieval time. Geomorphic and core-stratigraphic evidence points to a low stand that persisted from roughly AD 0 to 300. The lake then rose, and spent the ensuing seven centuries at or slightly below levels approximating those of the modern period. Around AD 1050 drought again drove the lake to low levels (roughly 15 m below the average elevation of the past 75 years), where it remained until approximately AD 1400. This Medieval lake regression and lowstand (and by inference, the causal drought) appears to be one of four such events that have occurred at Lake Titicaca during the past 3,500 years.

High-elevation glaciers from the south-tropical Andes have yielded precisely-dated ice cores that encompass Medieval time. The 1,500-year record from the Quelccaya Ice Cap (Thompson and Mosley-Thompson, 1987; Thompson *et al.*, 1985; Thompson *et al.*, 1986; Ortloff and Kolata, 1993), located near the northern boundary of the Titicaca watershed in southeastern Peru (Fig. 3.1), appears to register the same dryness that instigated Lake Titicaca's Medieval regression. The interval from AD 1040 to 1490 was characterized by low snow-accumulation rates. These were particularly low (and, by inference, the climate particularly dry) during most of the period AD 1040–1325. The peak dryness appears to have occurred between around AD 1245 and 1310.

Binford, Ortloff, and Kolata and their collaborators (Binford *et al.*, 1997; Ortloff and Kolata, 1993) document a temporal correspondence between the florescence of the Tiwanaku civilization, on the south end of Lake Titicaca, and the development of highly productive raised-field agriculture. These events occurred around AD 600, under moderately wet climatic conditions. Over the ensuing ~500 years, as Lake Titicaca remained at relatively high levels, the human population on this portion of the Altiplano grew to more than ten times the modern number. Around AD 1100, as Lake Titicaca and the Quelccaya Ice Cap registered drought, the Tiwanaku culture collapsed. The above-mentioned researchers attribute this fall to a decline in the productivity of the raised fields under the new, drier climate regime. They note that the drought, by decreasing the recharge to both surface-water and ground-water systems, would have diminished the land area available for raised-field cultivation, and that the lowered lake and resultant loss of wetlands would have increased the susceptibility of crops to frost damage in this high-elevation environment.

3.9. Southern Patagonia

Evidence of low lakes and desiccated marshes, much like that described above for California and the western Great Basin, has been found at several sites in southern Patagonia, in the lee of the Andes (Fig. 3.1). It indicates that at latitudes between 47°S and 51°S the climate became abnormally dry for several centuries prior to AD 1130. Water in the 'Catalon marsh' near the western end of Lake Argentino fell to levels that permitted colonization by southern Beech *(Nothofagus sp.)*. Remnant stumps from that colonization, exhibiting up to 100 annual rings, remain rooted in place today (Stine, 1994). The hydrographically closed Lake Cardiel (Stine and Stine, 1990) fell to one of its lowest stands of the Holocene, leaving behind deltaic sedimentary sequences and rooted stumps as evidence of the regression. Similar evidence points to an extreme lowstand of Lake Ghio at this same time (Stine, unpub. data, 1990).

3.10. Summary, Discussion, and Conclusions

3.10.1. SUMMARY

It is clear from over thirty studies, most of them published within the past decade, that Medieval climate over large parts of the Americas was, by modern standards, aberrant. In five regions—California and the northwestern Great Basin, the northern Rocky Mountains/Great Plains, southern Patagonia, the Yucatan Peninsula, and the south-tropical Andes—the Medieval period brought prolonged intervals of drought. In contrast, two North American regions—the Upper Midwest/sub-arctic Canada, and Alaska/British Columbia—experienced marked increases in wetness. In most of these regions the Medieval climatic departure was extreme, not only by modern standards, but by middle- and late-Holocene standards as well.

Because of dating limitations, a precise chronology of Medieval climatic change in the Americas cannot yet be constructed. But the following loosely constrained sequence of events does provisionally emerge:

- *Within two centuries of AD 100* the climate at many sites changed markedly. The new climate persisted for centuries.
- *Within a century of AD ~850* many sites registered a change to the conditions that define the Medieval climatic anomaly. Mono, Owens, Walker, Tenaya, and Lower Pahranagat lakes fell; Diamond Pond, the West Walker River, Osgood Swamp, the ponds of the northern plains, and the Patagonian lakes and wetlands shrank; northern Yellowstone dried; the Sand Hills of the west-central plains lost their vegetation cover and mobilized; and at least some of the Alaskan and Canadian glaciers began to advance.
- *Within 3 decades of AD 1110* many sites experienced an abrupt and marked alteration in climate. The forest/tundra border in central and eastern Canada began a southward migration, indicating cooling; the Upper Midwest began to dry; and Patagonia became moist. California and the western Great Basin shifted abruptly to wetness, and remained wet for ~5 decades (tree-drownings at Mono Lake, West Walker River, Osgood Swamp, and Tenaya Lake, and wetness recorded in the bristlecones) before returning to dryness (regressions at Mono Lake, West Walker River, Independence Lake, and Tenaya Lake; drought again recorded in the bristlecones). A similar dry-wet-dry shift seems to have characterized Lake Punta Laguna on the Yucatan Peninsula.
- *Within 50 years of AD 1350* the Medieval climatic anomaly had ended. Water bodies in California and the western Great Basin recovered from their lowest stands (Mono, Tenaya, Independence, and Lower Pahranagat lakes rose, as did Diamond Pond and the West Walker River; the San Joaquin Marsh freshened). Northern Yellowstone became wet (Jenny Lake rose, fire and mass-wasting decreased, and Yellowstone mammals adapted to wetter conditions). Ponds on the northern plains rose (Moon Lake, Chappice Lake); the Yucatan lakes rose (Chichancanab and Punta Laguna); and drought broke in the south-tropical Andes (Lake Titicaca recovered from its lowstand; snow accumulation rates rose on the Quelccaya Ice Cap).

3.10.2. DISCUSSION: ATMOSPHERIC CIRCULATION

Such widespread and considerably coherent shifts in climate must be attributable to changes in the planetary oceanic/atmospheric circulation system.[3] With regard to the Upper Midwest and the Canadian sub-arctic, a reorganization of the atmospheric circulation was proposed early on by Bryson (Bryson *et al.,* 1965; Baerreis and Bryson, 1967). During the wet centuries prior to AD 1100–1200, he reasoned, the summer polar front must have occupied a position over central Canada. This placed the strong westerly flow of the jet stream well to the north of Iowa. In the absence of this obstructive flow, incursions of warm wet air from the Gulf of Mexico brought abundant precipitation to the Upper Midwest. That the arctic treeline (which approximates the summer position of the polar front—Bryson, 1966) was migrating northward until AD ~1130 (see above) corroborates this view.

A contracted circumpolar vortex prior to AD ~1130, as proposed by Bryson, could also account for the aberrant dryness of California and the western Great Basin, as well as for the wetness of southern Alaska and western Canada. (It would likewise explain the recently reported Medieval wetness in Greenland—Meese *et al.*, 1994.) Such a circulation would have placed the wintertime polar front, which today brings the northeastern Pacific the preponderance of its precipitation, far poleward of its modern mean position, steering cyclonic disturbances that would otherwise reach California and Oregon into British Columbia and southern Alaska. Wiles and Calkin (1994) present a convincing argument that high wintertime precipitation (and by inference, a northwardly-shifted polar front) did indeed drive the Medieval advance of the south-Alaskan glaciers. They point out that land-terminating glaciers on the eastern, relatively maritime slope of the Kenai Peninsula have been observed to expand during modern intervals of high wintertime precipitation. Land-terminating glaciers on the less-maritime western slope, in contrast, have responded positively to cool summer temperatures. During Medieval time it was the precipitation-sensitive glaciers of the east-Kenai slope that expanded. No such expansion occurred on the west slope.

Bryson attributed the change from wet to dry conditions in the Upper Midwest around AD 1100–1200 to a shift in the position of the polar front. Around that time, he hypothesized, the circumpolar vortex expanded. Strong, dry westerlies now flowed across the Upper Midwest in summer, preventing penetration by rain-bearing gulf air. The documented southward shift in arctic treeline beginning around AD 1130 (see above) supports this hypothesis.

An expansion of the circumpolar vortex provides a plausible and convincing explanation for the climate change that commenced in the Upper Midwest and sub-arctic Canada around AD 1130. But it does not, in and of itself, explain the persistence, for a further 150–200 years, of extreme winter drought in portions of the mid-latitude west and of aberrant winter wetness in Alaska/British Columbia. This combination of conditions seems to necessitate not just a vortex expansion, but a concomitant change from zonal to azonal flow, with a strong ridge setting up and persisting over the eastern Pacific and Western Cordillera. Such a circulation would have deflected Pacific storms to the north of California and Oregon, and into Alaska and British Columbia. Intriguingly, AD 1130, when the arctic treeline of the Canadian interior began to lose ground to tundra, corresponds closely to the onset of the wet interval that separates the two California/Great Basin droughts, as well as to the short-lived wet period that punctuates the drought on the Yucatan. It seems possible that these brief wet spells reflect a time of atmospheric reorganization, as the northern circumpolar vortex switched from zonal to azonal flow.

Shifts in the circumpolar circulation during Medieval time were not restricted to the northern hemisphere, as evidenced by the climate changes in Patagonia. But with records from Patagonia presently limited to a latitudinal belt of just 4°, and with little other landmass existing at comparable southern latitudes, reconstruction of the austral circumpolar circulation is necessarily restricted to a narrow longitudinal band. None the less, just as in the northern hemisphere, such evidence as we have points to a contracted vortex throughout the centuries prior to AD 1130. This induced strong westerly flow directly over southern Patagonia, generating a strong Andean rainshadow and preventing incursions of moisture from the east. An increase in water levels of south-Patagonian lakes and marshes around AD 1130 signals a weakening of the rainshadow and an increased incidence of moist, easterly winds. This was due, presumably, to vortex expansion, and a resultant equatorward shift in the jet stream.

A more informed and sophisticated analysis will eventually be required to understand fully the Medieval oceanic/atmospheric circulation system. But the simple circulation reconstructions outlined above can serve as a point of departure for such analysis, and for predicting the nature of the Medieval climatic departures that characterized other regions of the Americas and the world. Such predictions can, in their turn, be field-tested by developing proxy records of climate change for the regions in question.

3.10.3. CONCLUSIONS

Large numbers of the world's scientists, economists, and policy analysts are turning their attention to a pressing question regarding future climate: "What will happen if anthropogenic global warming occurs?" But history and paleoclimatology suggest an equally pressing, though less-often asked question about our climatic future: "What will happen if anthropogenic global warming does not occur?" The historical and paleoclimatological records from the recent past reveal numerous multi-decade to century-scale intervals—the Medieval period included—when climate differed markedly from that of the twentieth century. Each of these past intervals represents a plausible and potential climatic future. What would be the environmental and socio-economic impact if such conditions were to return? It behooves a prudent society to plan for its climatic future by examining its past.

Notes

1. The concept of 'effective inflow' attributes all fluctuations of a lake to changes in inflow; evapotranspiration is treated as a constant, set at the average modern rate. Assessments of effective inflow using the Vorster model (Vorster, 1985) take into account both the magnitude and the duration of the Medieval lake-level fluctuations.

   In order to assess the reasonableness of attributing fluctuations in lake level to changes in inflow rather than evaporation, the Vorster model was employed in sensitivity analyses. For these, inflow was set at the average natural value for the modern period, and treated as a constant. When the evaporation rate was set at the highest annual rate recorded during the past 60 years, and the model was run until the lake equilibrated, the lake surface stabilized at an elevation fully 12 m higher than the Medieval lowstands. In other words, to draw Mono Lake to Medieval levels by changing only the rate of evaporation would require that the mean annual evaporation rate be far higher than the highest single annual rate recorded during the past six decades. In contrast, precipitation has been low enough in numerous individual years of the instrumental record (though during no extended string of years) to draw the lake to the Medieval low stand. While increased evaporation may have played a role in the desiccation of the stump sites, it must have been a secondary one.

2. Dates estimated from graphs illustrating 20-year running means.

3. Relatively little is known of oceanic conditions off the coasts of the Americas during Medieval time. Exceptions derive from the investigations of Pisias (1978), Baumgartner and others (1991), and Keigwin (1996). Pisias reported that ocean temperatures in the Santa Barbara Basin off southern California during the period AD 200–1250 were among the highest of the past 8,000 years. Analysis of charcoal and sardine scales (both indicative of warm conditions) in varved cores from the Santa Barbara Basin led Baumgartner to conclude that the low ocean temperatures of around AD 800 rose to a maximum value by AD 1000, and that by AD ~1400 temperatures were again low.

   Keigwin (1996) found that, around AD ~950, surface temperatures of the northern Sargasso Sea near the Bermuda Rise stood at least 1°C above modern values. The influence of these higher Atlantic temperatures were likely profound, not only in Greenland, Iceland, and Northern Europe (due to increased heat influx), but in eastern North America as well. In this regard, note that Stahle and his co-workers (Stahle and Cleaveland, 1994; Stahle et al., 1988) have produced >1,000-year-long ring series from bald cypress (Taxodium distichum) of the Carolina and Georgia coasts. These series, which track well the springtime rainfall recorded during the past century, provide "weak evidence that dry conditions [along the southeastern seaboard] were more prevalent from [AD] 1020 to 1275..." than during other parts of the record.

References

Abbott, M.B., Binford, M.W., Brenner, M. and Kelts, K.R. (1997) A 3,500 ^{14}C yr. high-resolution record of water-level changes in Lake Titicaca, Bolivia/Peru. *Quat. Res.* 47, 169–180.

Arnold, J.E. (1992) Complex hunter-gatherer-fishers of prehistoric California: Chiefs, specialists, and maritime adaptations of the Channel Islands. *Amer. Antiq.* 57, 60–84.

Auer, V. (1958) The Pleistocene of Fuego-Patagonia. Part II: History of the Flora and Vegetation. Helsinki. *Annales Academiae Scientiarum Fennicae*, Series A, III (Geologica-Geographica).

Auer, V. (1960) The Quaternary history of Fuego-Patagonia. *Proc. of the Roy. Soc.*, B-152, pp. 507–516 and 533–538.

Baerreis, D.A. and Bryson, R.A. (1967) Climate change and the Mill Creek Culture of Iowa. *Archives of Arch.* 29, 1–673.

Baumgartner, T., Souter, A., Vicente, F. and Byrne, R. (1991) Is the Medieval Warm Period recorded in the varved sediments of the Santa Barbara Basin, California? Workshop on the Medieval Warm Period, Tucson, November 5–8, 1991.

Binford, M.W., Kolata, A.L., Brenner, M., Janusek, J.W., Seddon, M.T., Abbott, M. and Curtis, J.H. (1997) Climate variation and the rise and fall of an Andean civilization. *Quat. Res.* 47, 235–248.

Brooks, C.E.P. (1926) *Climate Through the Ages.* Ernest Benn, London.

Bryson, R.A. (1963) *Proceedings of the Conference on Climate of the 11th and 16th Centuries,* Aspen, Colo., June 16–24, 1962. National Center for Atmospheric Research Technical Notes 63–1. R.A. Bryson and P.R. Julian (eds.)

——. (1966) Airmasses, streamlines, and the boreal forest. *Geog. Bull.* **8**, 228–269.

Bryson, R. A., Irving, W.N. and Larson, J.A. (1965) Radiocarbon and soil evidence of former forest in the southern Canadian tundra. *Science* **147**, 46–48.

Bryson, R.A. and Murray, T.J. (1977) *Climates of Hunger: Mankind and the Earth's Changing Weather.* University of Wisconsin Press, Madison.

Curtis, J.H., Hodell, D.A. and Brenner, M. (1996) Climatic variability on the Yucatan Peninsula (Mexico) during the past 3,500 years, and implications for Maya Cultural Evolution. *Quat. Res.* **46**, 37–47.

Davis, O.K., Jirikowic, J. and Kalin, R.M. (1991) The radiocarbon record of solar variability and Holocene climatic change in coastal Southern California. *In* K. Redmond (ed.), *Proceedings of the Eighth Annual Pacific Climate (PACLIM) Workshop,* March, 1991, pp. 19–33. Interagency Ecological Program Technical Report No. 31, California Department of Water Resources.

Dean, J.S. (1994) The Medieval Warm Period on the southern Colorado Plateau. *Climatic Change* **26**, 225–241.

Euler, R.C., Gumerman, G.J., Karlstrom, T.N.V., Dean, J.S., and Hevly, R.H. (1979) The Colorado Plateaus: Cultural dynamics and paleoenvironment. *Science* **205**, 1089–1101.

Fritts, H.C. (1963) *Proceedings of the Conference on Climate of the 11th and 16th Centuries,* Aspen, Colo., June 16–24, 1962. National Center for Atmospheric Research Technical Notes 63–1. R.A. Bryson and P.R. Julian (eds.).

Graumlich, L.J. (1993) A 1,000-year record of temperature and precipitation in the Sierra Nevada. *Quat. Res.* **39**, 249–255.

Hadley, E.A. (1996) Influence of late-Holocene climate on northern Rocky Mountain mammals. *Quat. Res.* **46**, 298–310.

Hadley, E.A. (1997) Evolutionary and ecological response of pocket gophers *(Thomomys talpoides)* to late-Holocene climatic change. *Biological Jour. of the Linnean Soc.* **60**, 277–296.

Heusser, C.J. (1966) Polar hemispheric correlation; Palynological evidence from Chile and the Pacific north-west of America. *In* J.S. Sawyer (ed.), *World Climate, 8000–0 BC, Proceedings, International Conference.* Roy. Met. Soc., London, pp. 124–41.

Hodel, D.A., Curtis, J.H. and Brenner, M. (1995) Possible role of climate in the collapse of Classic Maya civilization. *Nature* **375**, 391–394.

Huntington, E. (1924) *Civilization and Climate* (3rd ed.). Yale University Press, New Haven.

Jones, T.L., Brown, G.M., Raab, M.L., McVickar, J.L., Kennett, D.J., York, A., Spaulding, W.G., and Walker, P.L. (1997) The Medieval Climatic Anomaly in western North American prehistory, unpublished manuscript.

Keigwin, L.D. (1996) The Little Ice Age and Medieval Warm Period in the Sargasso Sea. *Science* **274**, 1504–1508.

Laird, K.R., Fritz, S.C., Maasch, K.A. and Cumming, B.F. (1996) Greater drought intensity and frequency before AD 1200 in the northern Great Plains, U.S.A. *Nature* **384**, 552–554.

LaMarche, V.C. (1974) Paleoclimatic inferences from long tree-ring records. *Science* **183**, 334–338.

——. (1978) Tree-ring evidence of past climatic variability. *Nature* 276, 1043–1048.

Lamb, H.H. (1963) On the nature of certain climatic epochs which differed from the modern (1900–1939) normal. *Proceedings of the WMO-UNESCO Symposium on Climate Changes, Rome, 1961.* UNESCO Arid Zone Res. **20**, 125–150.

——. (1965) The early Medieval Warm Epoch and its sequel. *Palaeogeog., Palaeoclim., Palaeoecol.* **1**, 13–37.

——. (1977) *Climatic History and the Future,* Vol. 2, *Climate: Past, Present, and Future.* Methuen, London.

Lambert, P.M. (1993) Health in prehistoric populations of the Santa Barbara Channel Islands. *Amer. Antiq.* **58**, 509–522.

Leavitt, S.W. (1994) Major wet interval in White Mountains Medieval Warm Period evidenced in $\partial^{13}C$ of bristlecone pine tree rings. *Climatic Change* **26**, 299–307.

Lindstrom, S. (1990) Submerged tree stumps as indicators of mid-Holocene aridity in the Lake Tahoe Basin. *Jour. of California and Great Basin Arch.* **12**, 146–157.

Luckman, B.H. (1995) Calendar-dated, early 'Little Ice Age' glacier advance at Robson Glacier, British Columbia, Canada. *The Holocene* **5**, 149–159.

Luckman, B.H. (1992) Glacier and dendrochronological records for the Little Ice Age in the Canadian Rocky Mountains. *In* T. Mikami (ed.), *Proceedings of the International Conference on the Little Ice Age Climate*, Tokyo Metropolitan University, Tokyo, pp. 75–80.

Madole, R.F. (1994) Stratigraphic evidence of desertification in the west-central Great Plains within the past 1,000 years. *Geology* **22**, 483–486.

Meese, D.A., Gow, A.J., Grootes, P.M., Mayewski, P.A., Ram, M., Stuiver, M., Taylor, K.C., Waddington, E.D., and Zielinski, G.A. (1994) An accumulation record from the GISP2 ice core as an indicator of climate change throughout the Holocene. *Science* **266**, 1680–1682.

Mehringer, P.J. and Wigand, P.E. (1990) Comparison of late Holocene environments from woodrat middens and pollen: Diamond Craters, Oregon. *In* J.L. Bettencourt, T.R. Van Devender and P.S. Martin (eds.), *Fossil Packrat Middens: The Last 40,000 Years of Biotic Change*, University of Arizona Press, Tucson, pp. 294–325.

Metcalfe, S. and Hales, P. (1990) Holocene diatoms from a Mexican crater lake. *In* Kociolek, J.P. (ed.), *Proceedings of the Eleventh Diatom Symposium*, Memoirs of the California Academy of Sciences, San Francisco, pp. 501–514.

Meyer, G.A., Wells, S.G., Jull, A.J.T. (1995) Fire and alluvial chronology in Yellowstone National Park: Climatic and intrinsic controls on Holocene geomorphic processes. *Geol. Soc. Amer. Bull.* **107**, 1211–1230.

Moratto, M.J. (1984) *California Archeology*. Academic Press, Orlando.

Nichols, H. (1967) Central Canadian palynology and its relevance to northwestern Europe in the late Quaernary period. *Rev. of Palaeobot. and Palynol.* **2**, 231–43.

Nichols, H. (1970) Late Quaternary pollen diagrams from the Canadian Arctic barren grounds at Pelly Lake, Keewatin, N.W.T. *Arc. and Alp. Res.* **2**, 43–61.

Ortloff, C.R. and Kolata, A.L. (1993) Climate and Collapse: Agro-ecological perspectives on the decline of the Tiwanaku State. *J. of Archeo. Sci.* **20**, 195–221.

Pisias, N.G. (1978) Paleoceanography of the Santa Barbara Basin during the last 8,000 years. *Quat. Res.* **10**, 366–384.

Raab, M. (1994) The Dead at Calleguas Creek: A Study of Punctuated Cultural Evolution during the Middle-Late Period Transition in Southern California. Unpublished report prepared for the Environmental Division, Naval Air Weapons Station, Point Mugu, California.

Ryder, J.M. and Thomson, B. (1986) Neoglaciation in the southern Coast Mountains of British Columbia: Chronology prior to the late Neoglacial maximum. *Can. Jour. of Earth Sci.* **23**, 273–287.

Sorenson, C.J., Knox, J.C., Larson, J.A., and Bryson, R.A. (1972) Paleosols and the forest border in Keewatin, N.W.T. *Quat. Res.* **1**, 468–473.

Stahle, D.W., Cleaveland, M.K. and Hehr, J.G. (1988) Carolina climate changes reconstructed from tree rings: AD 372 to 1985. *Science* **240**, 1517–1519.

Stahle, D.W. and Cleaveland, M.K. (1994) Tree-ring reconstructed rainfall over the southeastern U.S.A. during the Medieval Warm Period and Little Ice Age. *Climatic Change* **26**, 199–212.

Stine, S. (1987) Mono Lake: The Past 4,000 Years. Unpublished Ph.D. dissertation, University of California, Berkeley, 615 pp.

Stine, S. (1990) Late Holocene Fluctuations of Mono Lake, eastern California. *Palaeogeog., Palaeoclim., Palaeoecol.* **78**, pp. 333–381.

Stine, S. and Stine, M. (1990) A record from Lake Cardiel of climate change in southern South America. *Nature* **345**, 705–708.

Stine, S. (1994) Extreme and persistent drought in California and Patagonia during Mediaeval time. *Nature* **369**, 546–549.

Thompson, L.G. and Mosley-Thompson, E. (1987) Evidence of abrupt climatic change during the last 1,500 years recorded in ice cores from the tropical Quelccaya Ice Cap, Peru. *In* W.H. Berger and L.D. Labeyrie (eds.), *Abrupt Climate Change: Evidence and Implications,* Reidel, Dordrecht, pp. 99–110.

Thompson, L.G., Mosley-Thompson, E., Bolzan, J.F., and Koci, B.R. (1985) A 1,500-year record of tropical precipitation in ice cores from the Quelccaya Ice Cap, Peru. *Science* **229**, 971–973.

Thompson, L.G., Mosley-Thompson, E., Dansgaard, W., and Grootes, P.M. (1986) The Little Ice Age as recorded in the stratigraphy of the tropical Quelccaya Ice Cap. *Science* **234**, 361–364.

Tuthill, S.J., Field, W.O. and Clayton, L. (1968) Post-earthquake studies at Sherman and Sheridan Glaciers. *In The Great Alaska Earthquake of 1964* (Hydrology Volume). NAS Publication 1603. Washington: National Academy of Sciences, pp. 318–328.

Vance, R.E., Mathewes, R.W. and Clague, J.J. (1992) A 7,000-year record of lake-level change on the northern Great Plains: A high-resolution proxy of past climate. *Geology* **20**, 879–882.

Vorster, P.T. (1985) A water-balance forecast model for Mono Lake, California. Earth Resources Monograph No. 10, USDA Forest Service Region 5, 350 pp.

Wigand, P.E. (1987) Diamond Pond, Harney County, Oregon: Vegetation history and water table in the eastern Oregon desert. *Great Basin Nat.* **47**, 427–458.

Wigand, P.E. (1997) A late-Holocene pollen record from Lower Pahranagat Lake, southern Nevada, U.S.A.: high-resolution paleoclimatic records and analysis of environmental responses to climate change. *In* C.M. Isaacs and W.L. Thorpe (eds.), *Proceedings of the Thirteenth Annual Pacific Climate (PACLIM) Workshop, April 15–18, 1996,* Interagency Ecological Program, Technical Report 53, California Department of Water Resources, pp. 63–77.

Wiles, G.C. and Calkin, P.E. (1994) Late Holocene, high-resolution glacial chronologies and climate, Kenai Mountains, Alaska. *Geol. Soc. Amer. Bull.* **106**, 281–303.

Wiles, G.C., Calkin, P.E. and Post, A. (1995) Glacier Fluctuations in the Kenai Fjords, Alaska, U.S.A: An evaluation of controls on iceberg-calving glaciers. *Arc. and Alp. Res.* **27**, 234–245.

Chapter 4

Approaching the Medieval Optimum, 212 to 1000 AD

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Most of those who have written sometime this century about the historical impact of climate change have been geographers or climatologists: C.E.P. Brooks, Ellsworth Huntington, Hubert Lamb.... Historians have tended thus far to be either blankly indifferent or somewhat scornful. Some notable exceptions are to be found among the community of French historians associated with *Annales*, a journal founded in 1929 and committed to forging links with other subjects though especially geography. Fernand Braudel was a doyen in this respect. Le Roy Ladurie, 'by common consent the most brilliant of Braudel's pupils' (Burke, 1990:61) produced what may still be the most widely cited of all the climate-and-history studies. In *Times of Feast, Times of Famine*, he stressed how complex is the challenge of assessing the effect on crop yield of secular changes of mean air temperature that may well not exceed a degree Celsius. Also he discerned a disposition to want things both ways on the question of folk migrations, 'The Teutons of the first millennium before Christ are supposed to have left their countries of origin because of the cold. The Scandinavians of the period before AD 1000 are supposed to have done the same thing for exactly the opposite reason—the mildness of the climate, stimulating agriculture and thus also population growth, is said to have led to the departure of surplus male warriors' (Le Roy Ladurie, 1972).

It may, in principle, be possible to square this circle by distinguishing between those migrations, born of economic stress, that have taken the form of the wholesale displacement of the people in question; and those, rooted in enrichment, that have involved a people's expanding out of what still remains their homeland area. In practice, the distinction is not always easy to draw, the Hun migrations and the Islamic expansion post-632 being cases in point. At all events, one can take the

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more fundamental point that Professor Le Roy Ladurie went on to make. This was that judgement on many correlations between climate change and the human historical process ought to be suspended until more thorough climate research has been done. Fortunately, the continual development of techniques for the identification and dating of objective indicators may soon generate time series sufficiently definitive and reliable to compel the interest of historians at large.

Indeed, the provision of such information will confront historians of every hue with stark choices as to how far, or under what circumstances, climate shifts may be causative factors in societal and political change. In the meantime, it may be useful to review existing knowledge in order to identify correlations that will merit more stringent examination when the time is ripe. One general relationship that strongly invites our attention is that between the civilization of Europe and its climate from the peaking out of the Roman Empire to the high Middle Ages. As civilization seems to wane and wax and then eventually to wane again so, too, does climate seem to deteriorate, improve and then peak out. In both spheres, the thirteenth century stands out as an optimum (Brown, N.G., 1994). This particular narrative will effectively conclude with the tenth.

In middle latitudes in that day and age, deterioration was very generally understood to mean cooling rather than warming. This would be true from year to year or, as and when change on this timescale was perceived at all, from century to century. Greater warmth extended growing seasons and allowed of more intensive crop growth. It reduced the spread of ice and of permafrost subsoil. Most likely, any rises in sea level thus induced through ice melt would not be too abrupt. Nor is it probable, in any case, that contemporaries would discern gradual rises in mean sea level or, if they did, see them as due to global warming. There is, of course, a tradition all round the world (from the Congo to Oceania, the Americas, Mesopotamia and the Judaeo-Christian realm) of a sudden and all-but-apocalyptical rise in sea level close to the beginning of time, a rise associated with tremendously heavy rain or, so the Araucanians in Chile believed (Graves, 1968), a huge earthquake. Often part of the myth is that such a deluge will recur at the end of time. Nowhere, however, does there seem to have been much sense of more subtle variations in the height of the sea except, of course, in regard to harbor silting.

Manifestly, the welcome given in the past to extra warmth contrasts with our apprehension of it. However, this contradiction is not as acute as it may appear. The prime concern of analysts today is less with the direction of temperature change than with what its pace may be, particularly in regard to the effect on rainfall patterns. Besides which, there is a counter-current of scientific opinion, especially associated with the State Hydrological Institute at St. Petersburg (Budyko and Izrael, 1991), that has argued that even the present prospects for global warming hold out hope of real advantage for certain regions, either through ice melt or else a shrinking of the bounds of precipitation deficit. The lands that have most regularly been cited in this connection are parts of the Arctic littoral, the Canadian Prairies and much of the Russian Federation. Against that, it can be said that nowhere can thrive nowadays in the absence of a strategy for global management; and also that differential impact between regions could actually make it a lot more difficult to agree on such a strategy. To which one might add that many of us remain persuaded that, when temperature rise is fed into other ecological trends, the outlook for the Arctic environment as a whole looks decidedly adverse.

4.1. Empires and Nomads

In campaigns waged between 106 and 124 AD, the Emperor Trajan captured for Rome the territory that then became Dacia along with Armenia; Assyria and Mesopotamia; and Arabia Petraea. He thus staked out what effectively were the outer limits of Roman rule. The endeavors made under Severus from 208 to conquer Scotland were checked by his death in York in 211. The following year his successor, Emperor Caracalla extended Roman citizenship to all the freemen of the existing Empire. That marked, rather querulously, the peaking of Rome as a political experiment.

Before considering the influence of climate change on European history over the rest of that millennium, it may be appropriate to ask how far such change may itself have been driven by anthropogenic (*i.e.* man-made) factors. Deforestation will have been the paramount one in that era. Its immediate effect will have been to release a lot of carbon dioxide into the atmosphere. But a century or two after the felling has ceased, the more stable result may be a decrease of atmospheric CO_2 because this reduction of biomass will have slowed down the carbon cycle globally. What is more, the albedo of the ground that has been cleared will characteristically be 1.5 to 2.0 times as great as was that of the forest canopy. Therefore, deforestation may ultimately make things cooler.

In Europe, deforestation went ahead quite steadily within the bounds of the Roman Empire. Then it slowed down abruptly during the fifth and sixth centuries: the depths of what have customarily been known as the 'Dark Ages'. In the seventh, it resumed, as witness the extensive clearance of the Ardennes (Latouche, 1981). However, the great phase of forest clearance in Europe was between 1050 and 1250, this against the background of steep population growth. Italy may have acted as something of a pioneer in this clearance (Fichtenau, 1991). German colonization eastwards certainly played a major part (Goudie, 1990).

Another area of great deforestation was China. Estimates of the relevant demography vary. But according to Albert Kolb her population fluctuated between 50 and 60 million between the Han dynasty (206 BC to 220 AD) and the year 1600. However, the percentage living in South China (the Yangtse valley and southwards) rose from eight in the Middle Han to an estimated 49 around the year 1450. This shift betokened a very vigorous and sustained colonization of the South with the peoples indigenous to its fertile valleys being partially assimilated but widely displaced (Kolb, 1971). Associated with this aggrandizement was extensive deforestation. All in all, it will be surprising if more detailed climatic profiles do not confirm significant anthropogenic influence during these centuries.

4.2. The Pulse of Asia?

Turning to the impact on history, the question most central thematically, as well as in a more literal sense, may still be the relevance or otherwise of Huntington's 'Pulse of Asia' thesis. His inference early this century was that the 'relapse of Europe into the Dark Ages... was due apparently to a rapid change of climate in Asia and probably all over the world—a change which caused vast areas which were habitable at the time of Christ to become uninhabitable a few centuries later. The barbarian inhabitants were obliged to migrate, and their migrations were the dominant fact in the history of the known world for centuries' (Huntington, 1907). More specifically, the impetus was 'rapidly decreasing rainfall and rising temperatures during the early centuries of the Christian era' though there was 'evidence of a slight reversal, and of a tendency toward more abundant rainfall and lower temperature during the Middle Ages'. But he also saw strong reason to believe that, underlying this oscillation, was 'a widespread and pronounced tendency', right across the last 2,000 years, towards greater aridity (*ibid.*, 13,14).

This supposedly millennial tendency towards dryness was a part of the then conventional wisdom though it does not readily relate to the notion of oscillation. Nor was the latter, even as Huntington described it, rhythmic enough to count as true pulsation. Nor could one expect it to be if a prime cause was forest removal by humankind. Nor is it always easy to check past claims out in the light of current knowledge. Take, for example, any changes registered in the level of the Caspian Sea. One can assume that these would be relatively easy to discern and measure because of the way a flattish topography around the north Caspian allows any fluctuation in level to be expressed in pronounced shifts in the shoreline. It also seems, at first sight, self-evident that any rise, say, betokens moister conditions in that immediate area. Yet in the 1960s, L.N. Gumilyev, a Soviet historical geographer, took as his point of departure the simple truth that by far the biggest source of Caspian water is the River Volga which is usually deemed to originate in a stream flowing out from the Valday Hills, half way between Moscow and St. Petersburg. He therefore argued that the Caspian was higher whenever rainfall run-off was greater in the upper Volga basin because of the storm tracks being displaced to those latitudes. Never mind that this displacement would have left the northern littoral of the Caspian drier.

Looking at evidence about tribal migration as well as at climate indicators and extending his purview as far east as Lake Aral and the Tarim Basin, Gumilyev concluded that, across the period 200 BC to 400 AD, storms did usually track well north which caused the Caspian to rise even as its own littoral became more droughty. He also said that, around the time of Christ, a warm phase peaked in the Black Sea region with mean temperatures perhaps half a degree above today's (Chappell, 1970), a turning point that is, in fact, matched well enough by evidence from other temperate regions in the Northern Hemisphere (Neumann, 1991). He further deduced that a northward shift of the storm tracks in the tenth century materially contributed to the decline of Khazaria: a fruit-growing, hunting and fishing polity that had dominated the Lower Volga-cum-Don region across the span, 600 to 900 AD. A third viewpoint can be that the storms were diverted more to southerly tracks, at least towards the end of the first era. It gains some corroboration from a peat-bog analysis from the Soviet Academy of Sciences. This found that, in a broad zone from Sweden through the Moscow region, levels of annual precipitation in the fifth century tended to be 50 to 100 mm below the longer term norms (Chernavskaya, 1990).

What we therefore move towards is an interpretation that, in contrast to Huntington's, does not talk simply of a diminution of rainfall right across a continental span. Instead, one needs be much concerned with latitudinal and, indeed, longitudinal shifts of climate patterns within that compass. To get some idea of how these shifts could have influenced tribal migrations, it may be best to review what we know of the Huns, the warrior nomads from East Asia who became prime movers in the *völkwanderung* that interacted so conspicuously with the collapse of the Roman Empire.

The Huns first appear on the stage of recorded history in the third century BC. They do so as the Hsiung-nu, a people prominent in the interplay between China and the barbarians to her North. One kinship group that thus emerged was the Qin. They established a dynasty (246 to 207 BC) that by the year 221 BC had extended bureaucratic imperial government across much of what we today understand by 'China'. They began the earthen ramparts that were successively evolved and extended to become, by 1573, the Great Wall.

At least since Edward Gibbon published his *Decline and Fall of the Roman Empire* (1776–88), the received wisdom has been that Hsiung-nu nomads thrust towards the West because their expansion southwards had been thwarted by China's national solidarity. Various authors have specifically cited the embryonic Wall as a factor in this (Waldron, 1983). But just what one should be talking about in terms of the strength and timing of the tribal pressures remains problematic. Gibbon's belief was that half of a total Hsiung-nu population of 600,000 turned westwards (Gibbon, 1776). Two recent Japanese estimates of the original Hsiung-nu warrior horde have been 60,000 and 300,000 respectively (Chin-Fu Hung, 1981).

At all events, the actual migration track would have been north and then west out of the mixed deciduous forest and grassland of the north-east Chinese plain into the steppes that run continually for three thousand miles near the 50° parallel of latitude, acting thus as a margin of separation between the taiga forests and the deserts to the south. As delineated by present-day geographers, that margin is typically 100 or 200 miles wide east of the Tien Shan mountain chain, *ca.* 85°E. To its west, the steppes broaden out to 400 or 500 miles, merging into the Black Earth region of the Ukraine. Throughout the centuries of which we are talking, their configuration would probably have been similar. Nor may their axial latitude have been much different. Today this axis does not coincide at all closely with the prevailing storm tracks. Nor do those tracks tend to be pronounced in those longitudes (Bartholomew, 1967).

The progress westwards of the Huns was halting to say the least. Only late in the fourth century of the Christian era, did they proceed through the Caspian and Sea of Azov region. As Gibbon observed, 'It is impossible to fill the dark interval of time which elapsed after the Huns of the Volga were lost to the eyes of the Chinese and before they showed themselves to those of the Romans' (Gibbon, *op. cit.*). But from then on the pace of events much accelerated. Having crossed the lower Volga in 372, they impelled German tribes into a major surge into Gaul across a frozen Rhine. Then four years later, most of the Visigoth people, also out to avoid them, fled into the Roman Empire.

Even a few years' climatic adversity on the warrior steppe lands might have been enough to trigger off the Huns push across the Volga. On the other hand, their advance could have been induced simply by a burgeoning awareness that (a) the climate to westward was always more agreeable, warmer in winter and moister in all seasons, and (b) her internal weaknesses were rendering Rome a much more tempting prey than the Chinese or, indeed, the Persian Empire. Moreover, if climate fluctuation was playing a part, it may have been as a general trend across the steppes towards low temperatures, these usually being associated, in winter at least, with greater aridity. It may be relevant to note that two Roman writers prominent in the fourth century were convinced the Huns had chilly origins. Even the swamps 'near the Polar Sea' were cited (Diesner, 1978).

4.3. The Decline and Fall

To pursue this argument further, it is needful to look more closely at Europe Proper plus the Mediterranean: meaning, in particular, at the Roman Empire at its zenith. Perusing evidence from writers such as Pliny the Elder and Strabo about such indicators as the distribution of vineyards, the late Hubert Lamb concluded that the regional climate started to warm sometime before 100 BC. He further inferred more doubtfully that, subject to some cold intermissions, this trend was sustained until 350 AD or thereabouts. It is a perspective that stands rather diametrically in opposition to that adopted by the great classical scholars, French and English, of the eighteenth century. Their consensus was that, throughout the era of Rome on the wane, central Europe was appreciably colder than even in their own chilly times.

Their doyen, Edward Gibbon, saw multiple reasons for Rome's fall. Thus he laid very strong stress on an accelerating decline of civic spirit, a decline he felt had its roots in the metamorphosis from Republic to Empire with the effective accession of Augustus, the first emperor, in 29 BC but which became ever more evident after the Age of the Antonines (138–192 AD). Territorial expansion beyond the limits of proper manageability was held to be a big factor in this decay, not least expansion towards a corruptive East. One 'oriental' influence that especially aroused Gibbon's ire was Christianity: to him a religion for 'enthusiasts', that most favored term of abuse among the eighteenth-century literati (Burrow, 1985).

Nevertheless, he did remark as well on how frequent and thick winter freeze-ups of the Rhine and of the Danube had often afforded the barbarians 'a vast and solid bridge of ice' across which to deploy large armies. He also noted that reindeer flourished in the forests of Germany and Poland. He saw Europe then as akin to the Canada of his own day, putatively that much colder, presumptively because of more forest cover. Yet he did allow that, ice bridges apart, it 'is difficult to ascertain and easy to exaggerate, the influence of the climate of ancient Germany over the minds and bodies of the natives. Many writers have supposed and most have allowed though as it should seem, without any adequate proof, that the rigorous cold of the North was favorable to long life and generative vigor.... The severity of a winter campaign, that chilled the courage of the Roman troops, was scarcely felt by these hardy children of the North who, in their turn,... dissolved away in languor and sickness under the beams of an Italian sun' (Gibbon, *op. cit.*, 1, IX).

Two generations later, Dr. Thomas Arnold (famous not only as the headmaster who shaped, at Rugby, the English public school system but also as a classical historian) described the Alps of the Hannibal era as far colder and more snowladen than in his day (Arnold, undated). Judgement on this specific point is made harder by our still being uncertain through which Alpine pass Hannibal's army entered Italy in 218 BC, thereby taking the Romans utterly by surprise. But a study of tree-ring and glacial data by the late Professor Neumann indicated that around that date median temperatures were, if anything, slightly milder than today's (Neumann, 1992). Maybe, indeed, this is what encouraged Hannibal to make this transit. Its heroic dimension would have lain in how, aided by the novelty of his elephantine armored corps, he prevailed against the fierce opposition of the local hill tribes.

All the same, the wider proposition that the later Roman era was a cold one around the Mediterranean and across Europe Proper has been supported by a UNESCO-sponsored study by Arie Issar. More particularly, this identifies as centuries of severe cold (not least in Eastern Europe) the first century AD and, less acutely, the fifth (Issar, 1992).

If this inference of prolonged interludes of cooling is valid, the likely corollary is that the climatic zones were displaced, however erratically and irregularly, towards the equator. One thing this connotes is that the blocking anticyclones so characteristic of Scandinavia in winter could extend more often southwards, thereby drawing dry easterlies across much of the continent, thus limiting winter snowfall. Evidence of sustained glacial recession across Scandinavia during the first four centuries of the Christian era may corroborate this interpretation (ibid., Fig. 11).

It may correlate, too, with archaeological observation of farming practice in Denmark. Here one alludes specifically to the notion of keeping cattle in enclosed stalls for much of the year. Around the beginning of the Iron Age (*ca.* 500 BC) the balance of the argument had switched towards this labor-intensive routine by dint of the advent of iron scythes to assist fodder collection but also because the climate was turning colder and wetter. During the first five centuries of the Christian era, things were drier and, for much of the time, warmer. Yet, contrary to what one might expect, the stalls continue in service and 'seem to get larger' (Hedeager, 1992). More bitter winters may be the explanation.

Another of the earlier students of climate change, C E P Brooks concluded that, in the fourth and fifth centuries, 'many German settlements were established on low ground, now swampy' (Brooks, 1926). In other words, the regime was drier than in the early twentieth century. However, one does have to bear in mind that, although Brooks was a bold and determined researcher, his chronologies were not always accurate in relation to modern practice.

For the British Isles, the evidence of water levels in aquifers may tend, on balance, to indicate wettish conditions (Horne, 1993). But is has also been suggested that the siting of Roman villas indicates not too high a frequency of south-westerly gales (Schneider and Temkin, 1978). Those bits of tentative evidence might be reconciled with Brooks by postulating a high frequency of blocking anticyclones over Scandanavia, at any rate in winter.

All the same, one does have to admit that the evidence about aridity in that era has remained incomplete so far as Western Europe is concerned (Lamb, 1995). For Italy there are said to be few indications of really wet years between 200 and 550 AD. However, Italy's situation would be intermediate in relation to the aridity hypothesis. More significant may be the finding that Lake Van in Turkey tended to be 'abnormally' high between the years 250 and 550 (Issar, *op. cit.*, Fig. 5). Perhaps this is where the analysis for Europe and the Mediterranean can dovetail with that for what we used to know as the European U.S.S.R. Maybe storms were displaced well to the South most of the time, a typical winter track perhaps being through the Gulf of Tunis and the Turkish Straits to southward of the Caucasus.

What has to be admitted, none the less, is that, for the Roman era, our knowledge of climatic conditions is still insufficiently synoptic, definitive and assured. Yet even now this may not be the main sticking point as regards an assessment of the historical impact of climate change. Perhaps that is instead the plethora of parallel or alternative explanations that has ramified since Gibbon's time for Rome's decline and fall. These explanations have ranged as far as soil depletion; a gold and silver shortage; lead poisoning from cooking pots and water pipes; and, in the fascist era, genetic pollution by inferior races (Bishop, 1959). Professor Lamb even suggested salt shortage induced by climate change as a relevant factor. The connection would presumptively be made by the rise in Mean Sea Level (MSL) associated with the irregular but insistent global warming he believes occurred for half a millennium between *ca.* 100 BC and 400 AD. Such a rise would have overrun coastal salt pans constructed in the wake of an MSL decline in a previous chilly era (Lamb, 1977, *op. cit.*, pp. 256–258).

Among the more persuasive reasons advanced for the waning of Rome is the recurrent spread of epidemics, this sometimes consequent upon contact with infected populations further afield. In that day and age, the Mediterranean region was highly exposed. All else apart, it was the very heart of the Empire. In 165 AD, troops returning from active service in Mesopotamia introduced a disease (probably smallpox) that killed off up to 30% of the population in parts of the Mediterranean basin. From then on, population decline punctuated by renewed epidemic outbreaks was to continue for centuries. The 'Justinian Plague' that hit the Eastern Roman Empire from 541–2 (*i.e.* in the middle of the reign of the Emperor Justinian) seems to have come out of Central Africa. It may well be related to a major dust veil event that occurred in 536. The impact of a comet or meteorite? (Baillie, 1994).

However, susceptibility to disease can be aggravated by chronic malnutrition perhaps consequent on secular climate change. Then, as always, its marginal location—especially as regards rainfall—made the Mediterranean peculiarly prone to climate alteration. Furthermore, the intricate topography of that region would have tended to make any general variation very selective in its local incidence. An overall trend towards dryness, say, could have made some existing farmlands too dry to remain fertile but allowed marshy areas to dry out enough to become workable. So differential an effect had not been lost on Aristotle who in *Meteorologica* had noted that 'In the time of the Trojan Wars, the Argive land was marshy and could only support a small population whereas the land of Mycenae was in good condition... But now... Mycenae has become completely dry and barren, while the Argive land that was formerly barren owing to the water has become more fruitful' (Webster, 1923). The significance of such variations in Roman decline will have been the greater given that, while the Empire at its zenith made elaborate arrangements to ensure the grain supply of the metropolis itself, it left the management of staple foodstuffs elsewhere largely to local oligarchies. The result was that, though famines were rare, local food shortages were common (Garnsey, 1988).

Over and beyond which, the southern side of the Mediterranean has, for well over a century, been recognized as a region that affords salient evidence of the advance, historically speaking, of the desert (Shaw, Sir Napier, 1932). So a question that constantly comes up is how far 'desertification' was due to climate change as opposed to war or plain mismanagement.

About Roman North Africa, this much may be said. The Carthaginians and then the Romans did achieve remarkably extensive margins of cultivation against a climatic background that apparently was broadly similar to today's. Some aspects of the 'granary of Empire' image that has been handed down may be spurious or unproven (Shaw, B., 1981). But this probably means merely that this received image is overdone, not that it is baseless. Yet if one does allow that the region did considerably have a granary role, it is not clear how far the Empire achieved this by high yield agriculture as opposed to sheer deprivation of the local peasantry. At all events, the agriculture of Egypt, at least, did not much depend on local climate variations. Then as now, the Nile rose far to the South. Besides which, one Roman innovation, especially in the Egyptian oases, was the deep and heavy tapping of fossil groundwater (Murphey, 1951).

That the Romans did inflict ecological damage on North Africa seems undeniable. Evidently, too, the margins of cultivation and settlement did recede markedly in later centuries. No doubt, also, climate fluctuations figured in this process. Yet none of this proves that around the turn of the fifth century, the most critical time in the decline of Rome, the desert fringes advanced significantly further north in Africa.

Nor would such an advance necessarily have accelerated the fall of the Empire as a whole. The importance of North Africa had largely lain in grain supply to a city of Rome that may have numbered up to a million free people and slaves. Nevertheless, the waning of this metropolis may have long preceded any contraction of its bread basket. Rome never was a city of peace and harmony. It was always riven with ugly tensions. Emperor Augustus claimed he had found Rome a city of brick and left it one of marble. But the brilliant engineering that shaped this urban experiment in 'hydraulic civilization' impacted only tangentially on the squalid poor. To the free Romans within this underclass, Augustus did offer *panem et circenses* as a palliative. But in the ensuing centuries, the circuses became ever more ugly, with the massive sacrifice for sadistic pleasure of wild animals largely from North Africa and slaves from anywhere. If the rulers of Rome eventually lost pride and confidence in their imperial role, this continual savagery may well have been a reason. By the third century, the cultural leadership of the Roman world was shifting towards the Aegean and Asia Minor, back to its Greek roots as one might say. In 330, the Emperor Constantine capped this tendency by founding Constantinople to be the new imperial capital.

But there still remains the question of whether the fall of the Empire might be due to its collapse under the weight of tribal invasion induced by climate change. By early in the third century, the military threat from the barbarian North was already looming large. Soon the bounties being paid to barbarian chieftains were matching the payroll of the imperial army. Meanwhile, a fundamental weakness in Roman military science was being more fully exposed. That weakness was an undue neglect of cavalry, the arm that Alexander the Great had valued so highly and which had likewise made sure that Hannibal left 60,000 Roman dead on the field of

Cannae. Infanteers marching along well-made roads to sustain an essentially linear defense of an imperial perimeter 16,000 km in extent and of very mixed terrain were always exposed to being locally overwhelmed. None the less, in the second half of the third century the imperial army was expanded in total size from 350,000 to perhaps half a million; and the cavalry proportion within the legions was raised (Ferrill, 1986). At the same time, the political factiousness of the army was curbed. All this allowed some restoration of fortunes. Campaigns by the Emperor Aurelian (270–5) were successful in striking a new balance in Danubia and in Asia Minor.

Soon, too, greater stability was being achieved on another, no less important, plane. Constantine became an instant Christian convert on the field of battle in 312. He then promoted his new-found faith as the Empire's established religion in all but name, thereby leaving it free to launch wave after wave of rather coercive conversion. This elevated and extended profile did nothing for the principled concern of the Early Church with the underdog, with the end of war and such like causes. Nor did it help the position within it of women. Nevertheless, it did promise to turn Christianity into a powerful agent for social cohesion, especially in the East.

Yet here we become witnesses to the final race against time. Ramsay MacMullen, himself a skeptic about climate change as a factor in decline, notes how violent attacks on other religions (already endemic in the more exposed provinces to north and east) grew more prevalent after 380 (MacMullen, 1984). It was a sure sign of deepening insecurity as the barbarian pressure resumed. The Visigoths who had been allowed to enter the imperial domain in 376 very soon turned on their hosts; and their horde annihilated at Adrianople in 378 the main legionary field army. From 395 until his execution for treason in 408, the Roman general Flavius Stilicho struggled to contain the depredations of Alaric, the Visigoth king. But from 395 the Eastern Empire and the Western were formally separated politically. In 407, the Huns crossed the Danube. In 410, Alaric sacked Rome itself. From then on, the survival of the Western Empire, in anything like its pristine form, was surely next to impossible. Its formal end came in 476. Those who would explain this outcome essentially in military terms would look still to the cavalry factor.

However, military history has been too much influenced by cavalry enthusiasts. Among their legends is that it was the 'powerful mounted forces' of the Goths that overwhelmed the Roman legions at Adrianople (Lawford, 1976), a flattish valley location surrounded by broken Balkan terrain. At the same time, many writers have followed Gibbon in arguing that the disinclination of 'enervated' Roman citizens to help defend their realm ineluctably led to even greater reliance on barbarians to protect them: as co-belligerents, cavalry auxiliaries, or even legionaries proper. On that argument, Stilicho, who himself was of Vandal origin, had no chance to be a great general. Above all, the imperial army remained deficient in authentic Roman cavalry.

But in his highly incisive review of the military explanation, Arther Ferrill firmly dissents. He points out to start with that it is 'now commonly recognized by professional historians that the Germans and even the Huns of the fifth century' generally 'fought on foot'. Both armies at Adrianople were largely composed of foot soldiers; and the troops on the Roman sides fought as toughly as ever, not least in adversity. The difference between the two sides was that the Goths were handled far better tactically. The Roman Empire 'on the eve of Adrianople was not obviously on a downhill course'(Ferrill *op. cit.*, 8 and 164). He also implies that personal errors of judgement by the hesitant and ambiguous Stilicho (including perhaps a decision freely taken to rely too much on barbarian troops) contributed very materially to the dramatic military decline of the Western Empire after 395.

Yet as always, if the performance of individual leaders has the power to make things worse, it also has the power to make things better. Had Stilicho been a Caesar, Hannibal or Alexander, he might have imposed himself upon the social and institutional weaknesses of army and empire. So he might then have prevailed against the added barbarian pressure apparently induced by the renewed swing to a colder, and presumably drier, climate then taking place. By much the same token, a case can be made for saying that, regardless of judgements about barbarian practice, the Roman army should have been reorganized further in the fourth century. What surely was needed at that stage was more recourse to mobility in depth. What that in its turn would have involved is the creation of theater cavalry reserves, akin to those led with some success in the sixth century by Ambrosius and Arthur, the Romano-British *duci bellori.*

No doubt Ferrill is right to cast doubt on the 'downhill all the way' notion of continual decline from the Antonines onwards, be this attributed to lead pollution or climate change or whatever. All the same, his argument would look the stronger if he did not celebrate so the material standards reached by the Empire at its zenith, not least as a 'hydraulic civilization'. He asserts that unlike 'most people before or since, until very recent times, *the inhabitants of the Roman Empire* (my italics) had ample supplies of fresh water for drinking and bathing, often transported across hundreds of miles in the famous systems of aqueducts'. He further asserts that there 'were public facilities for the elimination of bodily wastes that were unmatched anywhere until the nineteenth century' (*ibid.*, 12). Surely, such perceptions fail to take proper account of the lot of the slaves and the peasantry or, indeed, the urban poor. Nor, one might add, proper account of the world beyond Rome.

4.4. The Dark Ages?

A key question for our purposes remains what exactly was happening to Europe's climate during that hinge of fate, 375 to 410 AD. Broadly speaking, it does, indeed, seem to have been getting colder. But to what extent was this in fits and starts; and how may these relate to rainfall fluctuations? Then how may both the climate parameters relate to what we know of the chronology of the barbarian migrations?

High definition answers to such questions are just the kind of thing the new phase of palaeoclimatological research is starting to seek. Pending its findings, a measure of speculative reflection may be in order about the apparent impact of environmental pressure on the behavior of the peoples caught up in what was, by any reckoning, a situation of remarkably high and prolonged drama. It has sardonically been observed that the one thing everybody knows about the Roman Empire is that it declined and fell. The last 30 years or so, however, historians have laid more stress on the continuities between the Empire and its legatees in the West. The barbarians conquered only to meld with. There was not that much destruction. There was not so general and catastrophic a fall in population totals or living standards; and the decline that did occur was not just the direct result of genocidal rampage. In short, the 'Dark Ages' were not that dark.

This revised view does not, in itself, prove anything conclusive about climatic cause and historical effect. Nevertheless, it would be consistent with climate deteriorating through the turn of the fifth century, yet never so precipitately as to induce a frenzy of conflict as a kind of Malthusian 'natural check'. As much can be well corroborated by reference to the territory that was in the process of becoming England. The idea used to be that its population in Roman times reached somewhere between two and six million; and that whatever the actual total reached then, it had almost halved by the year 600 (Hoskins, 1988). Now both sides of that scenario are being actively reassessed (Hills, 1990). Take, for instance, the famous *De Excidio Britanniae* ('The Ruins of Britain') written in the early sixth century by Gildas, a Romano-British scholar writing from somewhere near Salisbury Plain. He talks apocalyptically of the divine retribution being unleashed on his decadent fellow-countrymen, possibly an allusion to a strike directly on the British Isles by a stream of meteorites (Clube and Napier, 1990). At all events, the images he otherwise generates of diverse and stable agriculture undermine received notions of 'massive population decline, abandonment of entire landscapes and massive discontinuity' (Higham, 1991).

Ironically enough, the polity that did come close to eclipse in the 6th and 7th centuries was what we have come to regard as the great survivor: the Eastern Roman Empire, alias Byzantium. From 500 it was smitten by 'a remarkable succession of droughts, plagues of locusts, earthquakes and other calamities'. The Justinian 'bubonic plague of 541–2, the first of its kind in history, was by all accounts a disaster of unprecedented magnitude' (Mango, 1980). It was to rage intermittently around the Eastern Mediterranean until 750. In Byzantium as elsewhere in the region, there was a widespread decline of urban life and culture. Take Constantinople itself. Its population was close to half a million around 550. But within 200 years, it had collapsed into a series of enclaves surrounded by urban desert. The worst of the sieges or blockades it successively endured in this era were that by the Persians, Avars and Slavs in 626 and the Arab one of 674–8.

Yet by the ninth century, Constantinople was leading the urban civilization that was Byzantium out of this long recession. But in terms of the shifts of climatic zones to be anticipated as global mean temperature alters, the timing of both recession and recovery is hard to explain. One might expect the North-East Mediterranean to have been helpfully moist in the sixth century. One might further expect it to have been drier by the ninth century as the phase of early medieval warming got under way. The evidence already cited from Lake Van would support this interpretation. Freeze-ups of the Nile in 829 and 1010 tend also to favor it. They do so because they indicate the Azores anticyclone extending regularly into central Europe even in winter, thereby drawing north-easterlies across Egypt.

Besides which, any attempt to explain Byzantium's misfortunes in terms of climate determinism ought concurrently to take account of the glories of nearby Persia under the Sassanids. From its foundation around 226 until it was overrun by the Arabs in 637, that dynasty was stable and registered many achievements: philosophical, artistic and military.

4.5. Mohammed and Charlemagne

The next element to introduce into this analysis has to be the Islamic Arab expansion which effectively begins with the death of Mohammed in 632 and peaks out militarily when Charles Martel defeats a Muslim army at Tours, near Poitiers in central France, just 100 years later. Since early this century, there has been a broad consensus that the initial thrust out of Arabia was fired by drought rather than plenitude (Becker, 1913).

Some recent research on the floods that come down the Nile each summer indirectly lends further support to this interpretation. It indicates that these floods were weak in 28% of years between 622 and 999, as against only 8% between 1000 and 1290 (Guinn, 1992). Moreover, the seventh and early eighth centuries were very typical within the former tendency (*ibid.*, Table 6.6). Here the critical variable is in the flow of the Blue Nile down from the Ethiopian highlands. This is mainly because fluctuations in the White Nile are moderated by the swamps of El Sudd.

By 632, Islam already held sway over the west side of the Arabian peninsula. Twelve years later, it had extended across all Arabia, Mesopotamia, the Levant and the Lower Nile (Matthew, 1983). The backbone of its 632 compass had been the mountain range that extends from the Yemen northwards to the Gulf of Aqaba. Across that upland region, annual rainfall averages are these days between 100 and 500 mm. They are that high because in summer the region is peripherally under the influence of the Indian monsoonal low. So, too, are those Ethiopian highlands that feed the Blue Nile. If in a given season, that low is weakened or displaced eastwards, both locations will suffer drought. In Mohammed's time, it would have been the same.

The new Islamic Arab zone achieved and retained a remarkable degree of cultural unity. Yet this was never matched by political cohesion. The absence of a theory of universal statehood may help explain this. The theological split between Sunni and Shi'ite, already emergent by the year 650, does more specifically. The disparate nature of the zonal geography does very fundamentally.

Nevertheless, the Abbasid caliphs in Mesopotamia were briefly to establish (in effect from the foundation of Baghdad in 762) a civilization acknowledged to have been among the most scintillating in the ancient world. Under the fifth Caliph, Harun ar-Rashid (ruled 786 to 809), Baghdad 'was the richest city in the world... Arab merchants did business in China, Indonesia, India and East Africa. Their ships were by far the largest and the best appointed in Chinese waters or in the Indian Ocean. Under their highly developed banking system, an Arab businessman could cash a cheque in Canton on his bank account in Baghdad. In Baghdad everything was plastered with gold. Not only was it used to adorn the women but also the pillars and the roof-beams... Intellectual, and even theological discussions were among the recreations of the educated classes' (Glubb, 1969). It is an efflorescence that closely coincided with the recovery of Byzantium and may be equally hard to explain in terms of a trite climatic determinism. After all, the twin rivers of Mesopotamia, the Tigris and Euphrates, rise in the Taurus-Zagros mountain zone on the fringes of the Byzantine heartland.

Another theme worth pursuing, in order to test how much this may reveal about the importance of climate change relative to other extrinsic influences on history, is one generated shortly before his death in 1935 by Henri Pirenne, a scholar who still stands out titanically in the economic and social historiography of medieval Europe. A measure of the standing he achieved within his lifetime is that, although himself Belgian, he was repeatedly invited to become the first editor of *Annales* (Burke *op. cit.*, 21).

The relevant Pirenne thesis is that the Mediterranean remained all through the Dark Ages even for the Germanic tribes, 'the very center of Europe', especially as a *mare nostrum* across which international commerce was preserved and within which it regenerated. Then came the great expansion of the Arabs. By 720 they held all the southern coastline of the Mediterranean from the Levant round to the Pyrenees. By *ca.* 850, they also held Crete, Sicily, Sardinia and Corsica. With the Mediterranean thus becoming 'a Moslem lake', so Pirenne's argument ran, the center of gravity of Christian Europe switched abruptly northwards. It did so most conspicuously with the creation of a Frankish Empire that in 800 (the year its ruler, Charlemagne, was crowned in Rome by the Pope as 'Holy Roman Emperor') stretched from the Elbe to the Gulf of Genoa and the Pyrenees: 'Without Islam the Frankish Empire would never have existed and Charlemagne without Mohammed would have been inconceivable' (Pirenne, 1976).

As the evidence accumulates, however, it is pointing rather firmly towards the conclusion that the material and moral decline of the western basin of the Mediterranean, in particular, was well under way by the time of Mohammed's birth in 570 or thereabouts. Agricultural performance was patchy (Brown, P., 1967). The urban crisis was acute. Thus the decay yet again of Carthage 'to a shadow of its former self... appears to be typical of cities, large and small, all over the Mediterranean' (Hodges and Whitehouse, 1989). Lending his authority to a similar interpretation, Cyril Mango noted the 'ease with which walled cities fell to an enemy who was often neither very numerous nor very skilled in siege warfare, and the absence of any urban resurgence after the enemy had withdrawn shows... that military hostilities were merely the last shock that brought down a tottering edifice' (Mango, *op. cit.*, 69). Perhaps the most bathetic individual example was the supineness of the citizenry of Alexandria in the face of successive nomadic intrusions. In particular, the sacking of their library by Christian zealots in 389 AD had been a disaster of incalculable proportions. More generally, one has to say that such enervation much facilitated the great offensive pushes by Allah's Bedouin: from Alexandria to Tripoli in two years, from Gibraltar to Toulouse in under ten....

The ascendancy of the Frankish North was soon confirmed by a multiple breakthrough in agricultural practice, 'By the early ninth century, all the major interlocking elements of this revolution had been developed: the heavy plough, the open fields, the modern harness, the triennial rotation—everything except the nailed horseshoe which appears a hundred years later... The agricultural revolution of the early Middle Ages was limited to the Northern plains where the heavy plough was appropriate to the high deep soils, where the summer rains permitted a large spring planting and where the oats of the summer crop supported the horses to pull the heavy plough' (White, 1976).

In France, the divide between the old light-plough agriculture and the new became identifiable quite exactly along a line trending eastwards from l'Ile d'Oléron. Also that divide corresponded closely with two others: the linguistic and the legal. The *langue d'oil* to the north was more Celtic and Frankish (and hence less classical) than the *langue d'oc* to the south. Likewise, in law the *droit coûtoumier* to the north used customary precedent more than the more codified *loi écrit* to the south. The persistence of this cultural contrast into recent times, in its linguistic aspect, affords some indication of its initial depth. Writing in 1967, Peter Brown, now a doyen among the scholars of late Antiquity, saw the legatees of the governing class of the Western Empire as entrapped in the rigidities of their Mediterranean life style as their world collapsed around them (Brown, P., *op. cit.*). Twenty-five years earlier, the Empire's 'failure to develop a *Northern* agrarian society, as an alternative to that of the Mediterranean' had been remarked in the Cambridge Economic History (Runciman, 1952).

Yet there is a linkage between the Carolingian or Frankish Empire and Islam, that other great beneficiary from the fall of Rome. It is a link that relates to one of the three strands in the strategy Charlemagne evolved for consolidating his empire at least for his lifetime. The other two strands, both of them psychological rather than material, were (a) to secure the endorsement of Rome through what was a Christmas coronation and (b) to build up the charisma of the court by making it the leading center of revived learning. But this third was the import via the Baltic of large amounts of Abbasid silver for the minting of coins (Hodges and Whitehouse, *op. cit.*, Chapters 5 and 8). The establishment thus of a currency that could serve to lubricate trade and, at the same time, extend the ruler's writ was very much in line with what the English kingdoms were doing by the late eighth century. Offa, king of Mercia, had something like ten million pennies in circulation (Metcalf, 1967).

However, the Frankish Empire had grown too big to survive Charlemagne long; and was, in the event, to be partitioned through negotiation in 843 and again in 870. None the less, it had already served as an institutional conduit for the transmission of much of Rome's political culture into the European future. Moreover, there had been throughout the Dark Ages a massive diffusion of Roman culture in a broader sense. Various of the distinct languages of Europe, as we know them today, were forged as vulgar Latin interacted separately with the respective native tongues of the different barbarian societies. The resurgence of Christianity across Western Europe, especially vigorous in the seventh century, both furthered and benefited from these lingual fusions.

That this acculturation could take place so readily and comprehensively, and so very largely in a peaceable manner, confirms to me at least that whatever part climatic stress may have played to start with in the barbarian migrations, it did not long persist as a major factor. Certainly, it was to be followed relatively soon by the phase we know as early medieval warming, a broadly salubrious regime particularly characterized by drier summers. In 1983, Robert Gottfried set the take-off at around 750, using the recession of Alpine glaciers as a key indicator (Gottfried, 1983). Most analysts today would accept that dating as being as valid as any.

From the Mediterranean comes some evidence, tentative at any rate, of a consequent displacement of rainfall patterns. The sophisticated systems for agricultural irrigation established in the last centuries BC by the Nabatean civilization (centered around Petra in present-day Jordan) survived the Arab conquest only to collapse finally a century or two later for some quite different reason (Bintliff, 1982). More conspicuously, the Abbasid dynasty broke down through the ninth century and beyond.

Rather more towards what was becoming the favored north, however, the Byzantine heartland of the Aegean basin plus the Anatolian plateau enjoyed another resurgence in the ninth and tenth centuries, this being especially manifest in territorial expansion northwestwards across Serbia and also into Danubia. The driving force behind this expansion was a new-found strength and vigor on the part of peasant proprietors and tenants as more of Anatolia was brought under cultivation (Harvey, 1989). Nevertheless, the overall effect of the onset of medieval warming had to be to advantage the new core area of North-West Europe as against the Mediterranean. The development in the former of relatively sophisticated statehoods can be seen as an expression of this.

4.6. Arctic America

As the medieval warm phase progressed, the ice-albedo feedback ensured that the relative warming was particularly strong on the margins of shrinking ice fields. The coast of the Alaska north slope was one locality affected. According to observations made some 30 years ago, the stubby Barrow peninsula then just clipped the northern margin of the whale population. Also it was less than 200 kilometers south of the median edge of the permanent ice which there reached 1000 kilometers further south than it did in the 'gulf of winter warmth' near Spitzbergen (Bartholomew *op. cit.*, 26). The Eskimo inhabitants of somewhere so finely poised would have stood immediately to gain from even a modest rise in the regional mean temperature. It was true then. It will have been true historically.

As much appears dramatically to have been the case, in fact, by 900. Several centuries of cultural evolution in the Barrow region had produced a distinctive Birnick way of life which then metamorphosed into the Thule culture. Subject to local modifications, this was soon to spread throughout Eskimo Alaska. But its furthest extension was to be across the more open and whale-inhabited seas to the east and on into Greenland. The outcome was a comprehensive cultural revolution and therefore a geopolitical one as well. As the pre-existing Dorset Eskimo culture was submerged, all Eskimos from Alaska through Thule then around the Greenland coast came to speak, subject only to variations in dialectic, the same Thule language. Correspondingly, many of the more material elements in what 'every schoolboy' used to know as 'Eskimo society' were superimposed by the victorious Thule: kayaks and umiaks; bows and arrows; built-up dog sleds; whale-oil lamps; stone igloos.... (Coe *et al.*, 1986).

The significance of all this in terms of European history is partly that the Thule culture thereby became the one that confronted (and in the 15th century triumphed over) the Vikings in Greenland. But it is no less that Europe was directly influenced by the tendency, clearly evidenced in the Thule experience, for warming to be more pronounced at high latitudes. However, any more exact correlation must take account of the seminal field work of Willi Dansgaard and his colleagues at the Geophysical Isotope Laboratory of the University of Copenhagen. They have confirmed that allowance needs here be made for a marked longitudinal time lag in the unfolding of climate change. When core analysis from the Greenland ice cap is set alongside evidence from England from around 850 to 1700, an uncannily neat match is found in the profile of temperature trends so long as you allow England to lag 240 years behind.

4.7. The Vikings

In fact, the period from the mid-sixth century to the end of the tenth appears to have been predominantly warmish over Greenland, the chief cool intermissions being 660 to 710 and 820 to 870 (Dansgaard *et al.*, 1975). Maybe a relatively salubrious climate helped the sixth-century Irish monks to sail in their primitive open boats to Iceland, then probably to Greenland and down the American seaboard (Ashe *et al.*, 1971). More remarked, however, is the subsequent expansion (as warriors, traders and colonists) of the Vikings. In 789 came the first dated raid on what by then was quite a tranquil Anglo-Saxon England: a Norwegian raid on the Channel coast which, however, heralded the eventual domination of much of the northeast of the country, this largely by the Danes (see below). In 836, the Norwegians founded Dublin; and, eight years later, were to appear before Lisbon and Seville. Around 865, the Norse farmer, Floke Vilgerdson, tried to settle in Iceland. He gave up, having lost his cattle and seen 'a fjord filled up by sea ice. Therefore he called the country Iceland' (The Landman Saga, *ca.* 1200 AD). Then in 874, Ingolf Arnarson arrived and did manage to establish himself. By 930, some 20,000 Norwegians had migrated to Iceland (Logan, 1985).

The frontier society thus created was to lead the early medieval European world as a model of parliamentary democracy (the Athling being formed in 930), of the rule of law, and, in due course, of inspirational literature. Slaves—nearly all of them Irish—are frequently mentioned in the Icelandic sagas. At the same time, however, the Nordic precept of according slaves certain rights seems to have been singularly well upheld in Iceland; and, in any case, the whole institution of slavery had died out there by 1100, a century or two ahead of the rest of Scandinavia (Kirkby, 1977). Nor, of course, was Iceland ever involved in a transcontinental slave trade. So the overall impression retained is of a polity that for quite a while could remain agreeably free and open since there was no indigenous population to subdue but also because the climate was congenial enough and the general resource base adequate.

In 982, the pugnacious Erik the Red discovered a new land to the west of Iceland. He called it 'Greenland'. According to the Greenlander Saga, he flattered that territory thus in order to persuade others to follow upon his founding of the Østerbygd settlement (near Cape Farewell) in 984–5. No doubt he was keen for the added security that a strong presence would afford. In a brilliant study of the Viking centuries published in 1911, Fridtjof Nansen (explorer, zoologist, oceanographer and international statesman) discerned a physical brutality towards the Thule Eskimo that must soon have left the Norsemen themselves fearful of revenge. But that cycle of insecurity is archetypically a function of having toeholds on an alien coast. It does not prove acute climatic adversity. What is clear is that Greenland was warmer then than in recent times. Nor is her coast entirely barren today.

Around the turn of the millennium, the Vikings reconnoitered the Labrador coast and established a settlement (though maybe only for 20 or 30 years) at the now well-excavated site at L'Anse aux Meadows on the northern tip of Newfoundland. Among the territories named in the sagas as having been reached in that era are *Helluland* (*i.e.* the slate or stone land) and *Markland* (*i.e.* the woodland). Nansen well identified the former as Labrador and the latter as Newfoundland. But he discounted as mythical the most famous of several other supposed landfalls, Vinland. He pointed out that the image thus conjured of a frost-free American land endowed so implausibly with wild grapes and self-sown wheat was entirely in line with a paradisal tradition in mythology going back through Homer to Ancient Egypt and Babylonia (Nansen, 1911). The fact remains, however, that nowadays at least two black grape varieties of American wild vine, *vitis labrusca* and *vitis riparia*, do grow as far north as the New England-St Lawrence region (Robinson, 1994). On the other hand, wheat is not indigenous to North America. So for 'wheat' read 'maize'? Suffice to note that the Icelandic sagas do reveal a commitment to historical truth high for their generation. Nor was classicism, some Aristotle excepted, at all in vogue in the early Middle Ages when the sagas were written.

The Norsemen also pushed northwards, mainly in pursuit of whales, seals and walruses. They appear to have visited Jan Mayen. By 875, according to Nansen's interpretation, they had reached the White Sea (Nansen, *op. cit.*, 2, Chapter XII); and always they were ready to interact either through trade or through combat, with the native inhabitants of the Far North of the Fenno-Scandinavian mainland. In due course (in 1194, the sagas tell us), they also reached the Svalbard archipelago.

Even so, the said push continued to be tentative. Not so long after its discovery, Svalbard was forgotten; and was to be rather ostentatiously rediscovered by William Barentz, the Dutch navigator, in 1596. Still more to the point, the Norsemen did not actually settle in the territory of Finnmark (the furthermost province of modern Norway) until the thirteenth century (Stagg, 1952). Nor did these Norwegian Vikings urbanize very far up their coast. Only eight settlements in medieval Norway were ever accorded the legal status of townships. Of these the northernmost (at latitude 63° 20') was Trondheim (Lunde, 1985), founded in 997.

To identify a climatic root cause, one may have to look towards a variation on the theme of medieval warming. In September 1993, Keith Briffa of the Climatic Research Unit at the University of East Anglia gave a paper at the Durham meeting of Britain's Quaternary Research Association. Taking quasi-fossil tree rings as his evidence, he inferred that the period 1100 to 1300 was actually rather cold in Northern Scandinavia. Most probably, summer temperature would have been the most critical parameter as regards growth. But it is possible also to envisage a significant effect from the longer, colder and drier winters that could have resulted across those latitudes from a displacement towards the Pole of the Eurasian anticyclones as the atmospheric circulation of the Northern Hemisphere reorganized in response to global warming. That in its turn would be consistent with what was said above about Khazaria and also Mongolia. It would, in due course, circumscribe the 'gulf of winter warmth'.

In some ways, however, the most amazing dimension in the whole phenomenon of Viking expansion was that afforded by the Swedes. Having built up their nautical propensities on the rich herring grounds of the Baltic, these Eastern Vikings (alias, 'the Rus') projected themselves along the system of natural waterways that, more than any other single factor, was to afford the basic unity of what we know as Russia. Early in the ninth century, the Rus established a settlement by the shores of Lake Ladoga. Around 840, a proto-state emerged around their new southern city of Kiev; and surviving manuscripts show them to have been raiding already the southern shore of the Black Sea. In 860, a Russian fleet of 200 small ships menaced Constantinople; and the city was only saved, so its Greek inhabitants averred, by a storm conjured up by the Virgin Mary. The Russian Nestor Chronicle recorded Novgorod as being founded in 862. Within twenty years, Russian pirates were abroad on the Caspian. About that time, too, the legendary Oleg made Kiev the capital of a broad Rus confederacy. In 907, this Grand Prince of Kiev confronted Constantinople with an armada comprising, so it was said, 2,000 ships and 80,000 men; and supported overland by cavalry (Kendrick, 1930).

Inside this overarching spread of the Norsemen and the Rus, the Danes, too, were pushing outwards. By the year 830, they had begun to raid Frisia and the English Channel coast. Then so far as England was concerned, this sporadic raiding gave way (between 855 and 874) to a massive take-over of the east and center of the country. Moving in hosts that may rarely have exceeded a thousand warriors and their families (Kirkby, *op. cit.*, 16), these Vikings came not so much to plunder as to settle and to rule. Early in the tenth century Vikings, largely of Danish extraction, laid the foundations of the Duchy of Normandy. Not merely did this vibrant martial state conquer England in 1066. Between 1060 and 1091, Norman troops wrested Sicily from the Arabs thereby completing, in effect, the Viking encompassing of Europe.

It is tempting to think that the Viking expansion may have been the outcome of some fortuitous breakthrough in technology. But there is no evidence of this in respect of weaponry. Nor is it easy to make a case along these lines in respect of maritime science. Take the two Kvalsund boats from Norway (the larger of them 50 feet long) that are usually, albeit tentatively, dated as from the late seventh or early eighth century (Wilson, 1991). It was to this kind of vessel, 'built of unpainted oak and clinched with iron nails, broad and shallow, of easy entrance and run, with a deep rockered keel and a steep sheer, that the Viking shipbuilder always remained faithful...' (Kendrick, *op. cit.*, 24). As things progressed, however, bigger ocean-going versions began to appear. Take the 'long ship' (*i.e.* man-of-war) from the late ninth century discovered at Gokstad in Sandefjord in 1880. Her length is all but 80 feet; and her displacement, 30 tons.

One Viking innovation, so far as the Scandinavian world was concerned, was the sail. It was during the eighth century that sails came in. They did so originally to supplement propulsion by oar though the wide-beam *knörr* or *knarr* boats later used on Atlantic crossings were to depend on them almost completely. But all the Vikings were thus doing was adopting belatedly something utilized in the southern North Sea since Roman times (Foote and Wilson, 1980). Similarly in their use of heavenly bodies (above all, the Sun and the Pole Star) to obtain navigational fixes, they were borrowing and refining science first developed elsewhere (Marcus, 1956). A pre-existing technique adopted and adapted when the time is ripe in more general terms (climate perhaps included) can hardly rank as a prime mover. It is therefore hard to gainsay the conclusion that 'better ships and improved methods of navigation were absolutely fundamental to the success of Viking activity ... but they were one of its instruments, not its cause' (Kirkby, *op. cit.*, 56). Nor does any of this nautical aspect bear much on the exploits of the Rus.

So again one looks for causation to the broader context of economic and social change. The key to this may lie in material advance at a time when western Europe proper was barely emergent from migratory turmoil: 'Although never free from internal trouble—as many a violently destroyed farm dramatically relates in the archaeological record—the Scandinavians were building up a self-confident civilization of their own between 400 and 800 ... In Sweden the Migration period and the period of two hundred years before the Viking adventure began (the Vendel period) has been described as an age of gold' (Wilson, *op. cit.*, 46). All of which could be held to confirm that the Nordic heartland areas of southern Scandinavia experienced the onset of medieval warming quite early on.

At all events, there will have been a close and positive association in that kind of society between rising living standards and population growth; and the latter factor is one that historians of the Viking period have stressed for a century past as inducing the overseas expansion. The widespread polygamy has been noted. So, too, has a cultural propensity to measure manhood in terms of the number of sons sired. A lack of wheeled ploughs would have exacerbated the demographic pressure (*ibid.*, 31–32).

To which one might add more specifically that soil analysis in Denmark has indicated that, in the ninth century, the amount of land under tillage markedly increased along with an added emphasis on protein-rich crops (Logan, *op. cit.*). It all points towards a classically Malthusian crisis of success, a crisis to which climate improvement may well have contributed. No doubt a strong tradition of primogeniture in rural society was one factor which favored recourse to piracy and emigration before the population pressures brought about a social collapse.

Then there is the political factor, meaning the fractious politics associated with the unsteady emergence of the three proto-states. Denmark would appear from the map to have been the easiest to unify. In fact, however, no Danish kingdom was consolidated until well into the tenth century. A century and a half earlier, its Swedish counterpart had emerged, centered on Uppsala and what would by then have become the very fertile plain around. Where nation building was always going to be toughest was on the forbidding topography of Norway. Harald Fairhair (*ca.* 850 to *ca.* 933) unified the south-west of that country with a ruthlessness that is understood to have much increased the Norse exodus; not least of alienated aristocrats and, above all, to Iceland.

Yet with the Viking expansion as with the earlier *völkerwanderung*, one is struck by how readily aggressive warfare gave way to convergence with the indigenous inhabitants: to intermarriage and to cultural fusion, not least as regards religion. Here, too, the inference may be drawn that the factors that had impelled these people abroad had never been so harsh as to drive them entirely beserk. Soon a potent expression of convergence was trade. In England, the Romano-British city of York waxed strong as a focus of trade across the Danelaw. Likewise through Dublin and Wexford, the Norsemen controlled much of the external trade of Ireland. From the early ninth century, the Abbasid silver flowing to the Franks was being procured by the Rus for the Franks in large quantities in exchange for fur and slaves, the respect in which Charlemagne would have been inconceivable without Mohammed (Hodges and Whitehouse, *op. cit.*, Chapters 5, 6 and 8). G.M. Trevelyan, the last of the great Whig historians, remarked of the Vikings that they 'combined the pride of the merchant with the very different pride of the warrior, as few people have done' (Trevelyan, 1942).

4.8. Europe Resurgent

In the tenth and eleventh centuries, the Viking rampage died down as the defenses against it became more organized; and as the Vikings themselves took more to commerce and Christianity. Europe was thus left free to evolve more fundamentally. The tenth century used to be seen by historians as 'an age of decadence and decay, the ruins of the Carolingian Empire' (Fichtenau, 1991). It was manifestly a time of change. It was so not least as regards an early loosening, notably in the Frankish lands and Italy, of the feudal system: meaning, above all, the way in which the peasant was obliged to render dues and services to his lord of the manor in exchange for personal protection (*ibid.*, 347). In order to make some assessment of whether this loosening betokened social advance or societal dissolution, it is necessary once again to view things in their wider setting. A pertinent question to ask is how far mass alienation found expression in an upsurge of millennarian prophecy as the first Christian millennium drew to a close.

After a special study of this matter, Henri Focillon arrived at the conclusion that a groundswell of belief did surge through Christian Europe that the year 1000 would witness the end of the world: 'Satan will soon be unleashed because the thousand years have been completed' (Focillon, 1971). However, the opinion of G.G. Coulton, the distinguished Cambridge medievalist, was that there was really little more than 'a somewhat heightened expectation' of a Second Coming that, all through the Middle Ages, never seemed far into the future. Nor does one get the sense of millennarian fantasy being endemic and well defined, socially and geographically, the way it was to be in late medieval times. That perception encourages one to view the tenth century as more one of solid progress than of inchoate decline (Coulton, 1965).

4.9. Culmination

Moreover, one is talking here of the past as prelude. For there is a general acceptance that the second half of the eleventh century witnessed the onset of what Coulton called 'a very real revival, comparable to that later revival which we call the Renaissance' *(ibid.)*. Deforestation and population increase were among the most tangible indications. But there was also something more holistic: something to do with the quality of what still remained for most a short life of poverty and constriction. Thus Lord Kenneth Clark, the art historian, wrote metaphorically about a 'Great Thaw that seems to have affected the whole world but its strongest and most dramatic effect was in Western Europe.... In every branch of life—action, philosophy, organization and technology—there was an extraordinary explosion of energy, an intensification of existence' (Clark, 1969).

What is more, this was to go on across North-Western Europe for close to two hundred years, thus coinciding closely with an accelerating 'thaw' in climatic terms: the 'Little Climatic Optimum' of the Middle Ages. A quite pronounced rise in mean temperature was to occur between 900 and 1250. Across England and the adjacent continent, it was close to one degree. By the early 13th century (effectively, the medieval optimum) mean summer temperatures in Western Europe were 0.7 to 1.4°C higher than in the 20th (Lamb, 1995 *op. cit.*, 179). Meanwhile, on certain glacial margins of the North Atlantic and the Norwegian Sea, there may have been early on a rise of several degrees. The tree line across Northern Europe (being very temperature sensitive) characteristically became 80 meters higher by the year 1300 than it is today (Grace, 1989). More generally, the frontiers of agriculture ascended 60 meters or so. Our data base is still nothing like complete enough to weave together a coherent picture of climate progression in which like is consistently being compared with like. Nevertheless, the broad indications are already reasonably clear.

By the fourteenth century, the climate was in decline over much of Europe: cooler, wetter and generally more unsettled. The impact thereof soon becomes apparent. Even in the relatively balmy days of the early thirteenth, Europe had faced a petrifying threat from the Mongols, itself considerably a product of the way rising rainfall reinforced the manpower and the horse power of the Inner Asian steppes. But over the next one or two hundred years, Europe endured decline for more intrinsic reasons. Mercifully, one cannot talk holistically of regression in all respects. But in some salient ones, European society did suffer a loss of cohesion and stability.

References

Arnold, T. (undated) Hannibal's Passage of the Alps. *In* L. Valentine (ed.), *Half Hours with Standard Authors*, The Library Press, London, pp. 133–164.

Ashe, G. *et al.* (1971) *The Quest for America*. Praeger, New York, pp. 271–272.

Baillie, M.G.L. (1994) Dendrochronology raises questions about the nature of the AD 536 dust veil event. *The Holocene*, 4(2), 212–17.

Bartholomew, J.C. (1967) *The Advanced Atlas of Modern Geography*. Oliver and Boyd, London, pp. 16 and 28.

Bartholomew, J.C. *op. cit.*, p. 26.

Becker, C.H. (1913) Chapter XI, *In* H.M. Gwatkin and J.P. Whitney (eds.), *The Expansion of the Saracens—The East, The Cambridge Medieval History*, Cambridge University Press, Cambridge, II, p. 331.

Bintliff, J.L. (1982) Climatic change, archaeology and Quaternary science in the Eastern Mediterranean region. *In* A.F. Harding (ed.), *Climate Change in Later Prehistory*, Edinburgh University Press, Edinburgh, pp. 143–161.

Bishop, M. (1959) *The Horizon Book of the Middle Ages*. Cassell, London, p. 13.

Brooks, C.E.P. (1926) *Climate through the Ages*. Ernest Benn, London, p. 340.

Brown, N.G. (1994) Climate change and human history. Some indications from Europe, AD 400–1400. *Environmental Pollution*, 83(1 & 2) 37–43.

Brown, P. (1967) The Later Roman Empire. *The Economic History Review*, XX, 2, 327-343.

——. *Op. cit.*

Budyko, M.I. and Izrael, Y.I., eds. (1991) *Anthropogenic Climate Change*. The University of Arizona Press, Tuscon, Chapter 10 (translated from the Russian by these editors).

Burke, P. (1990) *The French Historical Revolution*. Polity Press, Cambridge, p. 61.

Burke, P. *The French Historical Revolution, op. cit.*, p. 21.

Burrow, J.W. (1985) *Gibbon*. Oxford University Press, Oxford, Chapter 5.

Chappell, J.E. (1970) Climatic change reconsidered: Another look at 'the Pulse of Asia'. *The Geographical Review*, LX, 3, 347–373.

Chernavskaya, M.M. (1990) Moistening changes on the Russian plain during the historical period. *In* R. Bradzil (ed.), *Climatic Change in the Historical and Instrumental Period*, Masaryk University, Brno, pp. 130–133.

Chin-Fu Hung, (1981) *Harvard Journal of Asiatic Studies*, 41(2), 605.

Clark, K. (1969) *Civilization*. BBC and John Murray, London, 33.

Clube, V. and Napier, W. (1990) *The Cosmic Winter*. Basil Blackwell, Oxford, pp. 107–110.

Coe, M. *et al.* (1986) *Atlas of Ancient America*. Equinox, Oxford, pp. 46–48.

Coulton, G.G. (1965) *The Medieval Scene*. Cambridge University Press, Cambridge, Chapter VIII.

Coulton, *Ibid.*

Dansgaard, W. *et al.* (1975) Climatic changes, Norsemen and modern man. *Nature*, 255(5503), 24–28.

Diesner, H.L. (1978) (C.S.V. Salt, *trans.*) *The Great Migration*. GDR Edition, Leipzig, pp. 71–85.

Ferrill, A. (1986) *The Fall of the Roman Empire, the Military Explanation.* Thames and Hudson, London, pp. 41-42.

Ferrill, *Ibid.*, p. 12.

Ferrill, *op. cit.*, pp. 8 and 164.

Fichtenau, H. (P.J. Geary, *trans.*) (1991) *Living in the Tenth Century.* University of Chicago Press, Chicago.

Focillon, H. (1971) *The Year 1000.* Harper and Row, New York.

Foote, P.C. and Wilson, D.M. (1980) *The Viking Achievement.* Sidgwick and Jackson, London.

Garnsey, P. (1988) *Famine and Food Supply in the Graeco-Roman World.* Cambridge University Press, Cambridge.

Gibbon, E. (1776) *Decline and Fall of the Roman Empire.* W. Strahan and T. Cadell, London.

Glubb, Sir John (1969) *A Short History of the Arab Peoples.* Hodder and Stoughton, London.

Gottfried, R.S. (1983) *In* J.R. Strayer (ed.), *Dictionary of the Middle Ages*, 13 vols,. Charles Scribner's, New York, 3, p. 453.

Goudie, A. (1990) *The Human Impact on the Natural Environment.* Blackwell, Oxford.

Grace, J. (1989) Tree lines. *In* P.G. Travis, *et al.* (ed.), *Forests, Weather and Climate*, The Royal Society, London, pp. 233–245.

Graves, R. (introduced) (1968) *The New Larousse Encyclopedia of Mythology.* Hamlyn, London, 445.

Guinn, W.H. in Diaz, H.F. and Markgraf, V. (eds.) (1992) *El Niño*, Cambridge University Press, Cambridge, Chapter 6.

Guinn, *Ibid.*, Table 6.6.

Harvey, A. (1989) *Economic Expansion in the Byzantine Empire, 900-1200*, Cambridge University Press, Cambridge, pp. 1–13.

Hedeager, L. (John Hines, *trans.*) (1992) *Iron-Age Societies. From Tribe to State in Northern Europe 500 BC to AD* 700. Blackwell, Oxford, p. 208.

Higham, N.J. (October 1991) Old light on the Dark Ages landscape: the description of Britain in the *De Exidio Britanniae* of Gildas, *The Journal of Historical Geography*, 17(4), 363–372.

Hills, C. (October 1990) Roman Britain to Anglo-Saxon England, *History Today*, 40(10), 46–52.

Hodges, R. and Whitehouse, D (1989) *Mohammed, Charlemagne & the Origins of Europe.* Duckworth, London, pp. 169–170.

Hodges and Whitehouse, *op. cit.*, Chapters 5 and 8.

Hodges, R. and Whitehouse, D. *Op. cit.*, Chapters 5, 6 and 8.

Horne, B.J. (March 1993) *Journal of Meteorology* **18**(177), 92.

Hoskins, W.G. (1988) (first published 1955) *The Making of the English Landscape.* Guild Publishing, London, Chapter 2.

Huntington, E. (1907) *The Pulse of Asia.* Archibald Constable, London, p. 5.

Ibid., pp. 13 and 14.

Issar, A.S. (1992) *The Impact of Climate Variations on Water Management Systems.* Ben-Gurion University of the Negev, Sede Boker, p. 44.

——. *Ibid.*, Fig. 11.

——. *Op. cit.*, Fig. 5.

Kendrick, T.D. (1930) *A History of the Vikings.* Methuen, London, Part II, Chapter V.

——. *Op. cit.*, p. 24.

Kirkby, M.H. (1977) *The Vikings.* Phaidon, Oxford, p. 157.

——. *Op. cit.*, p.16.

——. *Op. cit.*, p. 56.

Kolb, A. (1971) *East Asia*, Methuen, London, pp. 70, 47.

Lamb, H.H. (1977) *Climate, Present, Past and Future.* Methuen, London, 2 vols., 2, Chapter 12.

Ibid., 2, pp. 256–8

Lamb, H.H. (1995) *Climate, History and the Modern World.* Routledge, London, Fig. 59.
Ibid, p. 179.
Latouche, R. (1981) *The Birth of the Western Economy.* Methuen, London, p. 189.
Lawford, J. (ed.) (1976) *The Cavalry.* Purnell, Abingdon, p. 44.
Le Roy Ladurie, E. (1972) *Times of Feast, Times of Famine.* George Allen and Unwin, London, pp. 88–89.
Logan, F.D. (1985) Vikings. *In* J.R. Strayer (ed.), *Dictionary of the Middle Ages*, 13 volumes, Charles Scriber's, New York, 12, 422–32.
——. *Op. cit.*
Lunde, O. (1985) Archaeology and the medieval towns of Norway. *Medieval Archaeology* XXIX, 120–135.
MacMullen, R. (1984) *Christianizing the Roman Empire.* Yale University Press, New Haven, p. 119.
Mango, C. (1980) *Byzantium, the Empire of New Rome.* Charles Scribners, New York, pp. 66, 68.
——. *Op. cit.*, p. 69.
Marcus, G.J. (February-October 1956) The Mariner's Compass: Its Influence upon Navigation in the later Middle Ages. *History* XLI, 16–24.
Matthew D. (1983) *Atlas of Medieval Europe.* Equinox, Oxford, pp. 50–51.
Metcalf, D.M. (1967) The Prosperity of North-Western Europe in the Eighth and Ninth Centuries. *The Economic History Review* XX(2), 344–357.
Murphey, R. (March 1951) The decline of North Africa since the Roman occupation: Climatic or human? *Annals of the Association of American Geographers* XLI(I), 116–132.
Nansen, F., (Chater, A.C. *trans.*) (1911) *In Northern Mists.* William Heinemann, London, 2 Volumes, 1, Chapter IX.
——. *Op. cit.*, 2, Chapter XII.
Neumann, J. (1991) Climate of the Black Sea Region around 0 C.E. *Climate Change* 18(4), 453–465.
——. (1992) Climatic conditions in the Alps in the years about the year of Hannibal's crossing (218 BC), *Climatic Change* 22(2), 139–150.
Pirenne, H. (1976) Medieval cities. *In* Havighurst, F. Alfred (ed.) *The Pirenne Thesis*, D.C. Heath, Lexington, pp. 1–26.
Robinson, J. (ed.) (1994) *The Oxford Companion to Wine.* Oxford University Press, Oxford, pp. 28–29.
Runciman, S. (1952) *In* M. Postan and E.E. Rich (eds.), *The Cambridge Economic History of Europe*, The University Press, Cambridge, II, Ch. 3.
Schneider, S.H. and Temkin, R.L. (1978) *In* J. Gribbin (ed.), *Climatic Change*, Cambridge University Press, Cambridge, pp. 228–229.
Shaw, B. (1981) Climate, Environment and History: The case of North Africa. *In* T.M.L. Wigley, *et al.* (eds.), *Climate and History*, Cambridge University Press, Cambridge, pp. 379–403.
Shaw, Sir Napier (1932) *Manual of Meteorology.* Cambridge University Press, Cambridge, 1, *Meteorology in History*, Chapter VI.
Stagg, F.N. (1952) *North Norway. A History.* George Allen and Unwin, London, pp. 51–52.
Trevelyan, G.M. (1942) *A History of England.* Longmans and Green, London, p. 74.
Waldron, A.N. (1983) The problem of the Great Wall of China. *The Harvard Journal of Asiatic Studies* 43(2), 642–663.
Webster, E.W. (trans.) (1923) *The Works of Aristotle, Meteorologica.* Clarendon Press, Oxford, Book I, Chapter 14.
White, L. The northward shift of Europe's focus. *In* A.F. Havighurst (ed.), *op. cit.*, pp. 166–168.
Wilson, D.M. (1991) *The Vikings and their Origins.* Thames and Hudson, London, p. 36.
——. *Ibid.* pp. 31–32.
——. *Op. cit.* p. 46.

Chapter 5

Paleohydrology of the Northern Negev: Comparative Evaluation of Two Catchments

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5.1. Runoff and Human Settlement of the Negev

Although most of the Negev Desert is today extremely arid with <100 mm annual precipitation (Fig. 5.1), it has experienced periods of intensive human habitation. Permanent water sources in the region are scarce, making runoff the dominant source for both domestic use and agriculture (Evenari *et al.*, 1971). The technology of collecting the runoff to cisterns and cultivated plots was practiced in the northern Negev during several periods, climaxing in the Late-Nabatean—Roman—Byzantine period (Kedar, 1957; Kloner, 1975; Mayerson, 1960; Rubin, 1989).

Most runoff farms in the northern Negev were irrigated by runoff from contributing slopes measuring <1 km^2 per farm (Evenari *et al.*, 1971). In some cases flood waters were [partly] diverted from ephemeral streams by dams. Physical properties such as lithology and the gradient of the runoff catchment area have an important effect on water yield (Yair, 1983; Yair and Lavee, 1982).

A secular variation in the magnitude and/or frequency of runoff events during historical times is liable to have had an impact on human settlement in this marginal region. Such is known to have been the case throughout the Holocene elsewhere (Ely *et al.*, 1993; Enzel, 1992; Knox, 1993). As regards the Negev, the possibility of historical climatic change influencing settlement patterns has been debated for almost a century, without much paleohydrological evidence from the Negev itself (Huntington, 1911; Evenari *et al.*, 1971; Issar and Tsoar, 1987; Issar *et al.*, 1991; Rubin, 1989; and further bibliography there).

Reconstructing historical runoff fluctuations, we should preferably use proxy evidence that has a direct, intrinsic relationship with runoff. Fluvial sediments have

A. S. Issar and N. Brown (eds.), Water, Environment and Society in Times of Climatic Change, 97–111.

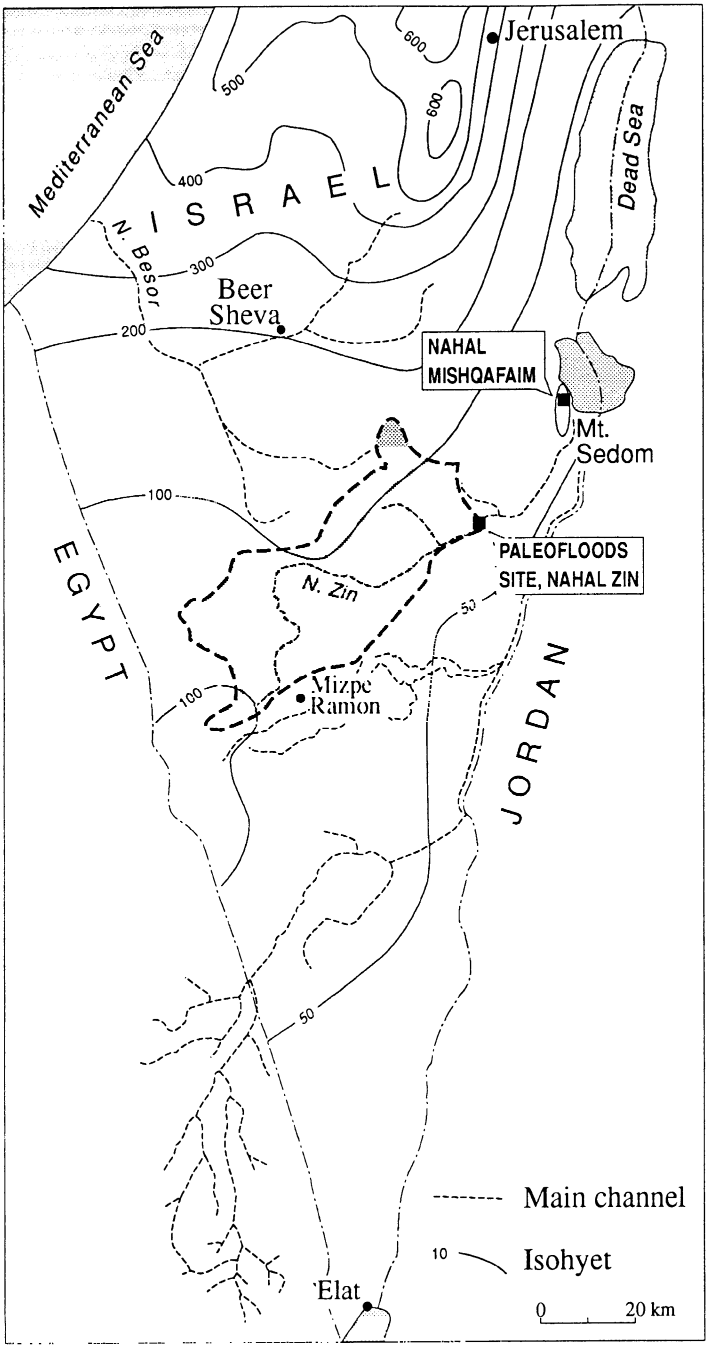

Figure 5.1. Location of study areas in the northern Negev Desert. Average annual rainfall is from the Israel Meteorological Service.

been used previously as paleoclimatic indicators (Bruins, 1986; Goldberg, 1984) but are subject to complex threshold and feedback processes with variable time lags in response to climatic change (Bull, 1991).

The northern Negev has enormous regional diversity. A proper understanding of its runoff processes will require well dated study covering the full range of catchment characteristics. Among these, the effect of variations in drainage basin sizes and lithologies must be investigated (Bull and Schick, 1979). Basins with and without human interference, which may alter natural runoff, present a particularly worthy subject of comparative research. So, too, do the effects of scale.

In this general context, we compare the results of two independent paleo-hydrological studies performed in two catchments differing in area by four orders of magnitude and contrasting markedly in lithology and topography (Fig. 5.2). Major runoff farm assemblages, including those of Avdat and Mamshit, are within the drainage basin of the larger catchment, Nahal Zin. The smaller one, Nahal Mishqafaim, was never affected by human interference. It preserves a rare paleomorphologic record inside a multiple-level salt cave. Temporal control in both study areas has depended on the radiocarbon dating of organic materials deposited by floodwaters.

5.2. Nahal Mishqafaim Study Area, Mount Sedom

Mishqafaim Cave, an underground segment of Nahal Mishqafaim, drains a 0.078 km^2 catchment in the hyperarid Mount Sedom salt diapir (altitude ~250 m below m.s.l.) into the Dead Sea (Fig. 5.2). The fluvial and karst systems of the actively rising diapir have developed since the end of the Pleistocene—Early Holocene, when the top of the diapir became exposed (Frumkin, 1996).

The Mishqafaim catchment consists mainly of a sub-horizontal plateau developed on Late-Pleistocene anhydrite caprock and lacustrine Lisan marls. The lower part of the catchment, which developed over the eastern border faults of the diapir, is steeper. The lithology in this part is anhydrite caprock and sandstone, dolomite and shales of the Neogene Benot-Lot member of the Sedom Formation (Zak, 1967).

East of the Benot-Lot shale member, the Me'arat Sedom salt member crops out, covered by a thin anhydrite caprock. Nahal Mishqafaim has been captured here into Mishqafaim Cave, formed within the extremely soluble rock salt. Four successive swallow holes (AI, BI, CI, DI; Fig. 5.3) developed along the subaerial channel during the Late Holocene, diverting it underground in four stages. During each stage a cave level was formed which is partly connected to the other levels. Cave development has proceeded mainly through gradual downcutting but sometimes also by stream capture. The underground channel surfaces again at the easternmost escarpment of the Mount Sedom diapir, flowing into the southern basin of the Dead Sea. The downward migration of the passages and outlets observed along the escarpment (Fig. 5.3b,c) is associated with Dead Sea level fluctuations (Frumkin *et al.*, 1991) and diapir uplift (Frumkin, 1996).

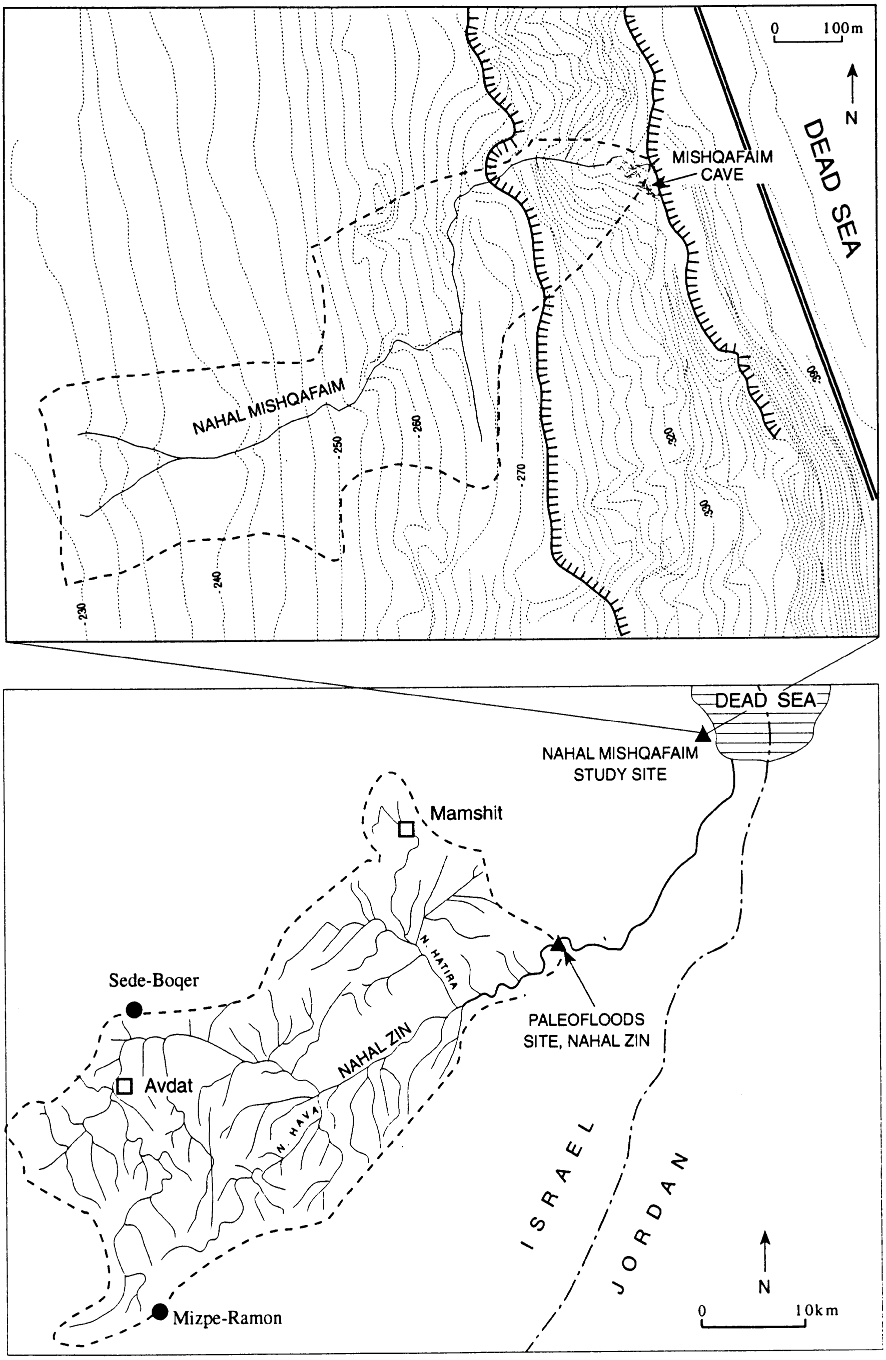

Figure 5.2. Nahal Zin and Mishqafaim studied catchments. Avdat and Mamshit are Nabatean-Roman-Byzantine cities, which practiced runoff harvesting within Nahal Zin catchment.

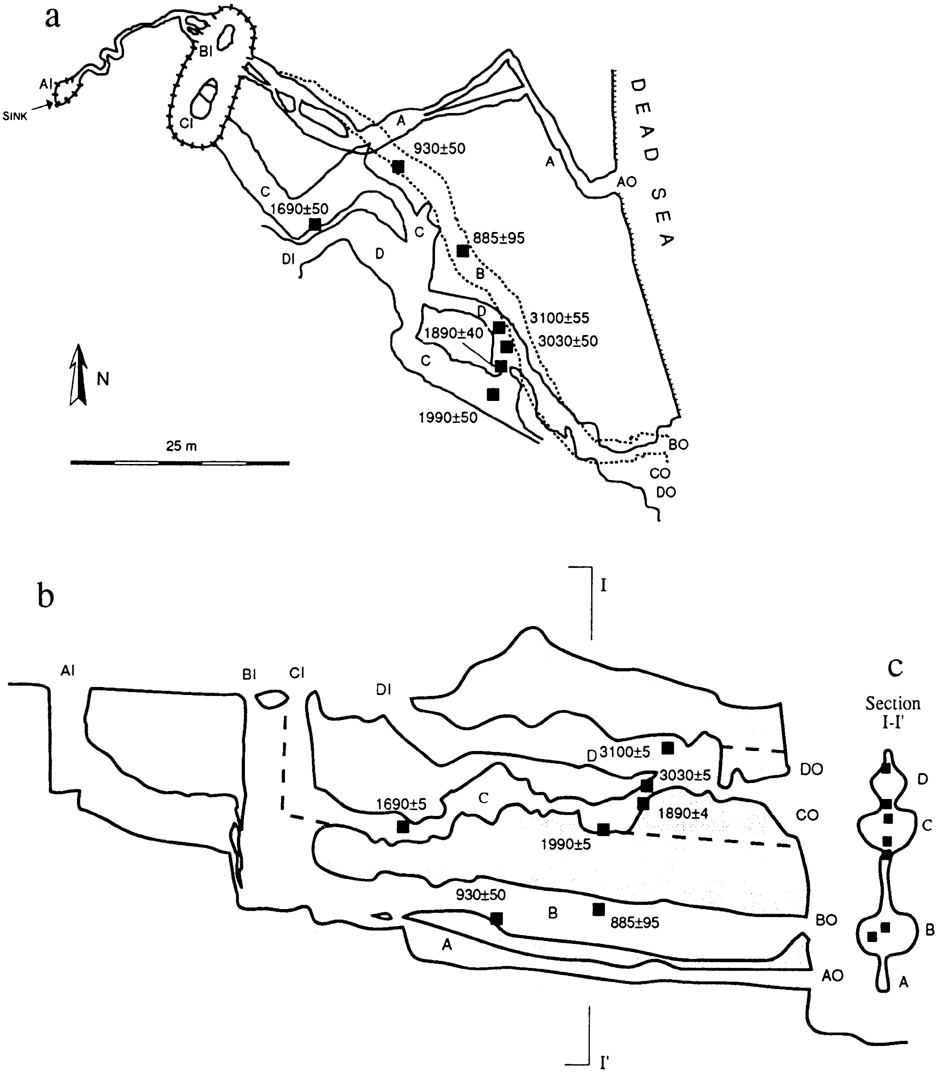

Figure 5.3. Mishqafaim Cave with radiocarbon dates (years BP). A, B, C, D are the four main levels of the cave, where A is the presently active one and D is the oldest. AI, BI, CI, DI are the input sinks of each passage, showing the gradual upstream capture of Nahal Mishqafaim; AO, BO, CO, DO are the outlets of each passage towards the Dead Sea. Note that older passages have been modified by breakdown following their formation by floodwater dissolution, (a) map; (B) profile; (c) cross section.

The small catchment area, with its relatively impermeable caprock and steep gradients, gives rise to very short run-off concentration times. Floodwater entering the cave is highly aggressive to rock salt, allowing rapid dissolution and transformation of cave passages morphology (Frumkin, 1994). Thus the geomorphic response is virtually immediate.

5.3. Nahal Zin Study Area

Nahal Zin has a 1,400 km^2 catchment draining limestone and chalk terrains of the north-eastern flanks of the Negev highland (altitude ~1,000 m a.s.l) (Fig. 5.2). The uppermost 233 km^2 part of the catchment is developed over hard carbonates with a flashy response to rainstorms due to the lack of soils or alluvial cover in the floodplain, although its average slope is only 0.006. Runoff harvesting was practiced here by the inhabitants of the Avdat runoff farm (Evenari *et al.*, 1971).

A sharp knickpoint in the main Nahal Zin channel leads to a wide alluvial channel 25 km long formed in erodible clays, marls and chalks (Bentor and Vroman, 1951). At the end of this reach a large tributary—Nahal Hava (120 km^2)—enters the main channel. This tributary drains a rocky basin which resembles the upper Zin basin in lithology, surface characteristics and specific overland flow generation, but is three times steeper (0.018) and thus has an even shorter concentration time.

In its next 25 km long segment, the main channel is filled with up to 20 m thick alluvium, where rapid infiltration attenuates flood hydrographs. At the end of this segment Nahal Zin is joined by its largest tributary, Nahal Hatira (266 km^2), which drains limestone, dolomite, chalk, conglomerate, and sandstone. Its average slope is between 0.020 and 0.030 and, accordingly, the concentration time is short. The ancient Mamshit inhabitants built a series of dams over a large tributary of this drainage system, directing the floodwater to their farms.

Further downstream Nahal Zin is entrenched in resistant carbonates forming a 20 km long canyon, 150–250 m wide. The canyon is partly filled with highly cemented conglomerates into which the channel is also entrenched, usually at the contact with the rocky wall of the canyon, where an inner mini-canyon 8–10 m deep was formed. On exiting the canyon, one enters the final 15 km long, moderately steep segment of Nahal Zin. It is cut in soft marly lacustrine Lisan sediments (Begin *et al.*, 1980), before flowing into the Dead Sea.

5.4. Present Rainfall and Runoff Regime

Annual rainfall over Mount Sedom currently averages some 50 mm. Over Nahal Zin mean annual rainfall varies from 90 mm in the upper parts of the catchment to 60 mm in the lower reaches (Fig. 5.1). Rainfall originates from two different sources—the Mediterranean and the Red Sea. Winter rain-bearing depressions arriving from the Mediterranean result in regional rainstorms of generally low to moderate intensity and relatively long duration (Katsnelson, 1979). Intensities average less than 15 mm hr^{-1} with peak values up to 45 mm hr^{-1}. The Red Sea Trough, in contrast, is characterized by the development, mainly in autumn and spring, of small convective rain cells from some tens up to a few hundreds km^2 in area. These rainstorms have a short duration but yield high intensities, typically 60 mm hr^{-1} (Sharon and Kutiel, 1986). Peak values recorded in Mount Sedom are 144 mm hr^{-1} over two minutes (Frumkin, 1992) and 102 mm hr^{-1} over thirty minutes (Gerson, 1972). High intensity rainstorms cause the majority of the large floods in the Negev Desert (Schick, 1988; Greenbaum, 1996).

Nahal Mishqafaim has no instrumental flow record. Nahal Zin has an average of 2.3 flow events a year for the 40 year long instrumental record. Only ten floods had a discharge exceeding 100 m^3s^{-1}. In 25% of the hydrological years no flows were observed. The largest floods on record are: 1.1.1945—550 m^3s^{-1}; 13.5.1945—600 m^3s^{-1}; 13.10.1991—550 m^3s^{-1}.

5.5. Paleohydrological Evidence

The different morphology and evolution rates of the two study sites prescribe different methods of study. Mishqafaim Cave channel, formed in rock salt, is extremely unstable with high downcutting rates (Frumkin and Ford, 1995), rapidly changing cross sections, and stream capture events every ~1,000 years (Fig. 5.3). The high evolution rate allows the use of conduit morphology for indicating runoff regime. On the other hand, the canyon of Nahal Zin is incised in highly resistant dolomite; profile and cross sections are assumed to have remained stable during historical times. This allows of reconstructing paleodischarges with confidence by the use of slackwater flood deposits.

Floodwater in both study sites carries floating organic debris which ultimately is deposited at the top of the flood sedimentary unit. This organic material allows accurate radiocarbon dating. We use conventional radiocarbon years BP but calibrated dates (Stuiver *et al.*, 1993) are also presented (Table 5.1).

5.5.1. NAHAL MISHQAFAIM

The underground segment of Nahal Mishqafaim has changed its course four times, forming four distinct cave passages (A–D, Fig. 5.3). Passage A is active today, but passages B, C, and D are inactive. These relict passages are detached from fluvial and karstic processes, as well as from other geomorphic processes which tend to eliminate surface terraces of similar age. Passages B, C, and D are only slightly altered by roof collapse and neotectonic activity and offer an opportunity for paleohydrologic reconstruction. However, identifying the water level, cross section and

Table 5.1. Radiocarbon dates of wood and charcoal

Site	Lab No.	^{14}C date yrs. BP	Calibrated age 1σ range		Passage width ratio (%)
Misqafaim D	982G	3,100±55	1436–1312	BCE	185
Misqafaim D	982E	3,030±55	1393–1258	BCE	111
Misqafaim C	982F	1,990±55	54 BCE–66	CE	259
Misqafaim C	1236	1,890±55	60–180	CE	259
Misqafaim C	1235	1,690±55	259–408	CE	221
Misqafaim B	982D	930±55	1033–1158	CE	117
Misqafaim B	848C	885±55	1043–1217	CE	117
Nahal Zin ZB	RT-1534	1,770±70	180–360	CE	
Nahal Zin ZB	RT-1533	1,450±50	580–650	CE	
Nahal Zin ZA	RT-1405	885±80	1040–1220	CE	
Nahal Zin ZB	Beta-64143	870±70	1060–1260	CE	
Nahal Zin ZA	RT-1532	440±40	1430–1490	CE	

channel profile pertaining to a single flood event is hardly possible because of the rapid changes in passage geometry associated with any such event. Therefore a more general method was adopted: The original width of the relict passages was measured in several places. The width ratio of a relict cave passage was defined as the ratio (in %) of its average width to that of the presently active passage (A).

The three main relict cave levels (B,C,D) have width ratios of ~200%. Intermediate levels are narrower, with a width ratio <100%. Passages B,C, and D were dated to ~900, 1,700–2,000, and 3,000–3,100 ^{14}C years BP, respectively. The age of each level is confirmed by at least two dates (Fig. 5.3). Intermediate narrow levels have no organic deposits, but their age is constrained between the dated levels above and below (Fig. 5.4).

A wider cave passage may be explained by two possible factors. First, increase in discharge or in the frequency of high-magnitude runoff events, caused by a general increase in rainfall intensity and/or the number of rainstorms over Mount Sedom. Second, a decrease of the hydraulic gradient caused by base-level changes, controlled by the diapir rising rate and Dead Sea level fluctuations.

5.5.2. NAHAL ZIN

The peak paleoflood discharges of Nahal Zin were reconstructed using fine-grained sandy and silty slackwater deposits (SWD) and other paleostage indicators (PSI) such as driftwood lines. These fine-grained sediments are rapidly deposited from suspension at sites in which flow velocities are significantly reduced; they represent the high stage of the flood. Such sites are: back-flooded tributary mouths; caves and alcoves in the canyon wall; channel expansions and obstacle shadows where flow separation causes eddies; and overbank deposits (Patton *et al.*, 1979). Preferable paleoflood sites are those that preserve multiple flood stratigraphic records which can be separated into individual flow events, using sedimentological criteria (Baker, 1987). Preservation of flood deposits depends on their elevation and distance from the active channel. Thus smaller floods deposit their sediment lower and closer to the active channel than large floods, leaving it more exposed to subsequent erosion.

Nahal Zin water surface profiles are determined by the HEC-2 application program (O'Connor and Webb, 1988). Comparing the elevations of the SWD and PSI to the computed profiles provides estimations of the peak discharge. Calibration of the paleoflood record was performed using two recent measured floods (Greenbaum *et al.*, in press).

Paleofloods were measured at several sites along the lower Nahal Zin canyon, where the drainage basin area is 1,150 km^2. Figure 5.5 shows the geometry, sedimentology and stratigraphy of two of the sites located in tributary mouths. The stratigraphy of each site was determined by field relations and radiocarbon dates. Four sedimentological markers were used to discriminate between individual flood sedimentation layers (Baker, 1987): (1) Abrupt vertical grain size change; (2) Slope colluvium and tributary alluvium interfingering with SWD from the main channel; (3) Fine silt, clay and organic layers which are deposited from the washload of the flood; (4) Buried soils of a low degree of development, which offers some estimate as to the time of exposure.

Twenty eight paleofloods were identified in Nahal Zin. The peak discharges of the paleoflood record range from 200 m^3s^{-1} to 1,500 m^3s^{-1} (Greenbaum, 1996). Five of these paleofloods were ^{14}C dated, yielding ages back to 1,770 ^{14}C years BP (Table 5.1, Fig. 5.4). These dates constrain the age of other floods which fall stratigraphically between the dated ones. Five floods (30–34, Fig. 5.5) are older than the earliest ^{14}C date (1,770±70 BP). The similarity in texture and lack of paleosols and abrupt contacts indicates that these floods are close in age to each other and to 1,770±70 BP (broken line in Fig. 5.4). The five undated floods prior to 1,770 BP suggest that this period was more visited by floods than the subsequent one, which had two floods only.

The results show two periods rich in high magnitude floods: That shortly before 1,770 years BP, and 1,380–920 years BP. Each of these periods was followed by one of low flood incidence, characterized also by weaker floods (Greenbaum *et al.*, in press).

5.6. Discussion

A preliminary report on the Mount Sedom caves suggested that the temporal variation of passage width ratio can be explained by changes in flood discharge, flood

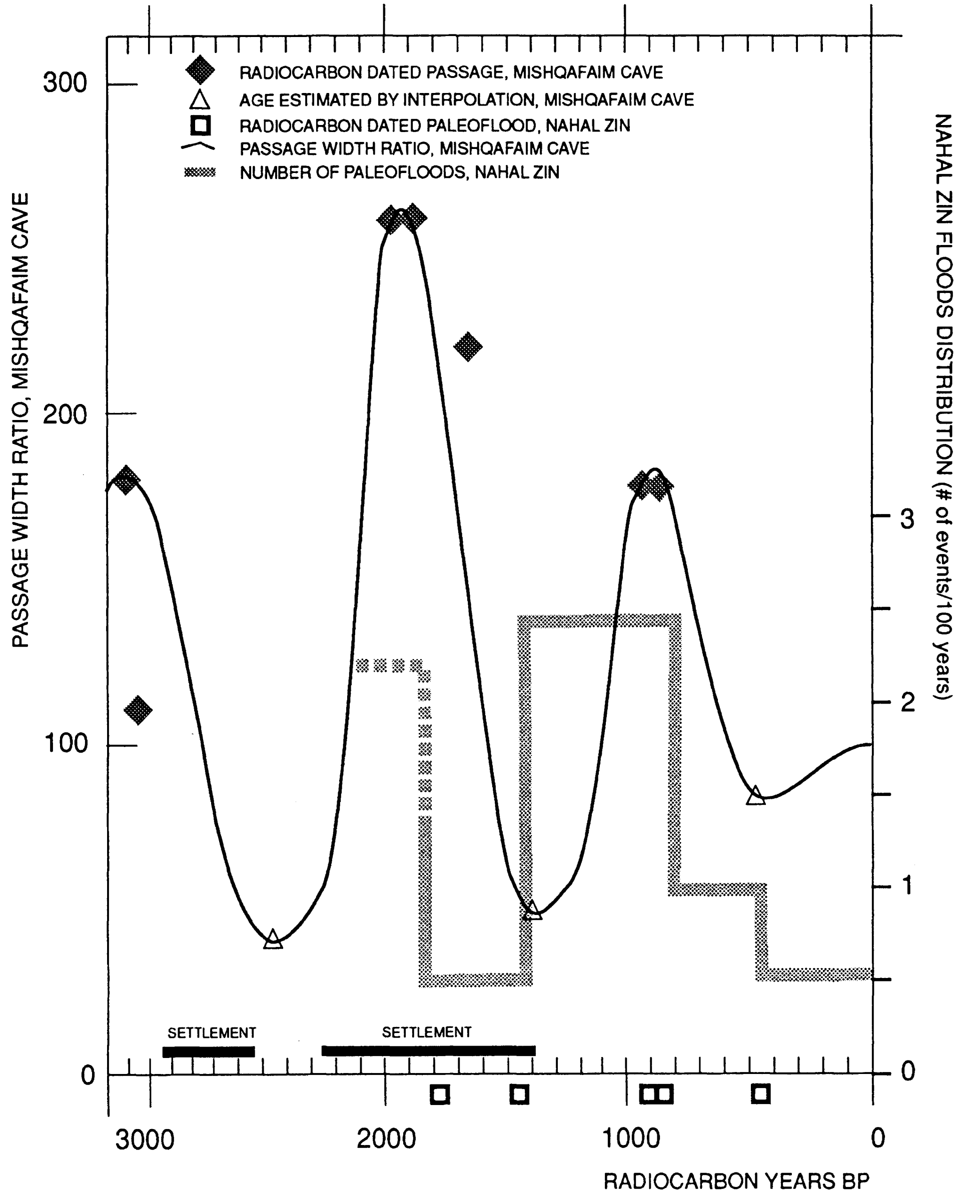

Figure 5.4. Temporal changes of Mishqafaim Cave passage width ratio and Nahal Zin flood distribution. The width ratio of a relict cave passage is defined as the ratio (in %) of its average width to the average width of the presently active passage. Flood distribution was calculated separately for five periods, giving an average value for each period. Cave passages and flood sediments lacking any organic material were dated by interpolation between available radiocarbon dates. Major settlement periods are indicated at the bottom.

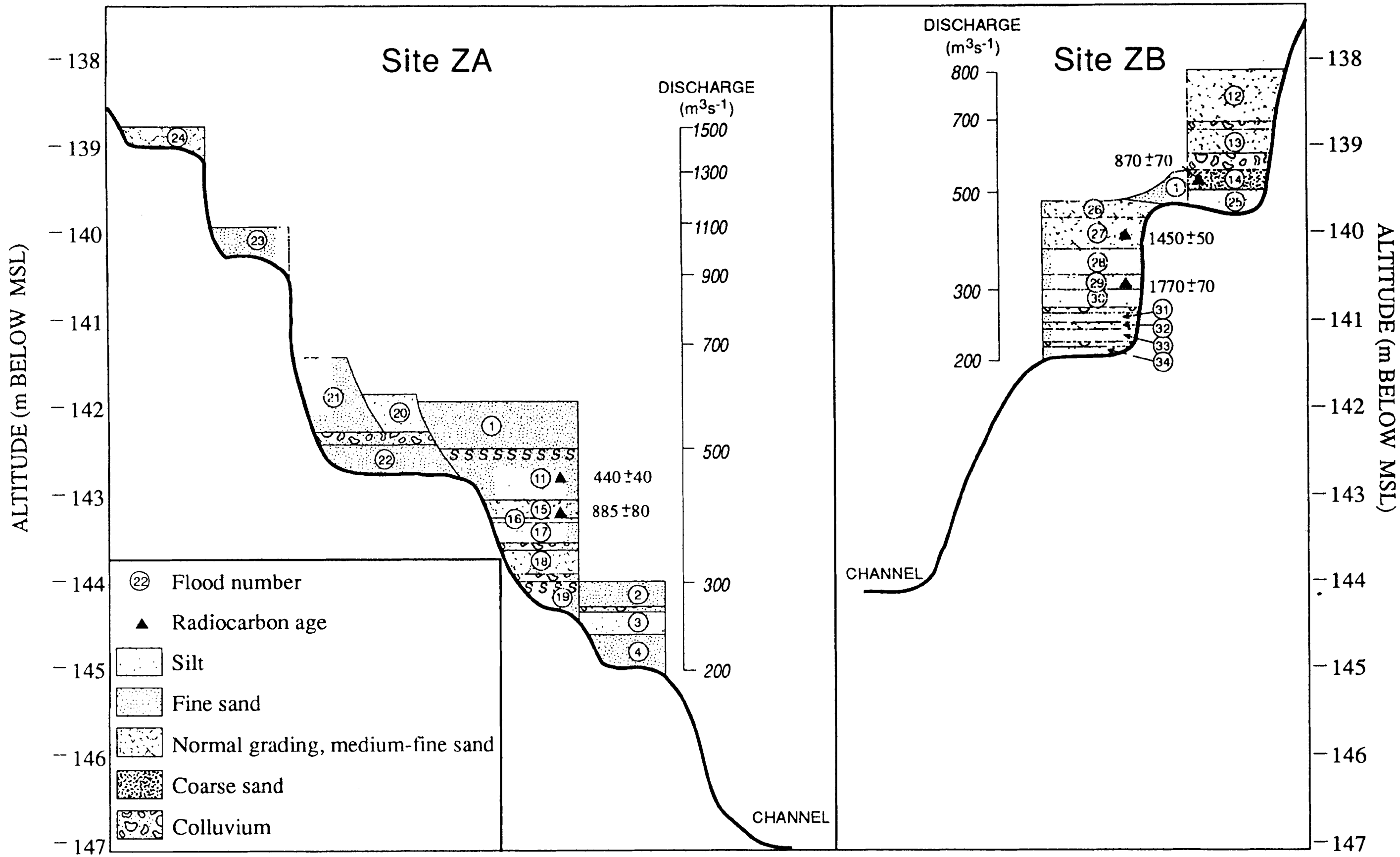

Figure 5.5. Flood sediment stratigraphy of the two main paleoflood sites in Nahal Zin with radiocarbon dates (years BP). Discharge for each level is computed using the HEC-2 prodecure. Flood no. 1 occurred in 1991, during this study. Floods 30–34 occurred prior to the oldest radiocarbon date.

frequency, and Dead Sea level fluctuations (Frumkin *et al.*, 1991). The way cave passage morphology integrates the influence of these factors over time makes it difficult to distinguish between the effect of each. Nevertheless this review of the Nahal Mishqafaim evidence and of the Nahal Zin paleoflood record does indicate that, during the last two millennia, cave passage form was dominantly influenced by local climatic conditions rather than by Dead Sea level.

The temporal distribution of both the Nahal Mishqafaim passage width ratio and the Nahal Zin paleofloods (Fig. 5.4) shows fluctuations of ~1,000 years frequency. The slight differences in timing between the two records may be attributed to uncertainties about the residence time prior to deposition of dated organic matter. This compounds the ^{14}C laboratory error.

The close comparability of records obtained from two basins so contrasting indicates that they share a common hydrological regime during the last 2,000 years. It is therefore suggested that frequency and magnitude of large paleofloods regionally can be associated with the width ratio fluctuations of Mishqafaim Cave. This is further corroborated by continuous measurements of morphological changes in another Mount Sedom cave (Frumkin and Ford, 1995).

Flood clustering episodes characterized by both more frequent and larger magnitude floods occur about 2,000 and 1,000 years BP. These episodes will be associated with a higher probability of synoptic conditions appropriate for high-intensity storms over the northern Negev. This trend could have been caused by a strengthening of either the Mediterranean winter regime or the Red Sea Trough systems. A clearly positive correlation exists in both study areas between the frequency and magnitude of large runoff events and amounts of seasonal precipitation (Gerson, 1972; Shanan, 1975; Greenbaum, 1996).

Evenari, Shanan, and Tadmor (1971), in their agricultural experiments at the reconstructed Avdat runoff farm, have classified the years 1961 to 1970 into three types: two 'wet' years with two to four 'medium to large' flood events and 160–165 mm/y rainfall; three 'drought' years with no 'medium to large' flood events and 25–90 mm/y rainfall; and four 'moderate' years. While the experimental farm generally succeeded in obtaining reasonable crops, the crop of a 'wet' year was on average 14 times larger than that of the 'drought' year (Evenari *et al.*, 1971).

A *runoff* drought year usually coincides with a *rainfall* drought year. All the same, the crop depends more on large runoff events than on annual rainfall because the reconstructed runoff harvesting system was designed to feed the runoff from large areas to small cultivated plots. During drought years in the past, stored water was probably all reserved for human and animal use, rather than for agriculture. The possible occurrence of several successive drought years undoubtedly posed a severe limiting factor for agricultural production. Under the present rainfall regime, years with a shortage of runoff have a mean recurrence interval of 3–4 years (Bruins, 1986).

The paleohydrologic fluctuations reported here evidently influenced the development of runoff farming and human settlement in the northern Negev. The beginning of this settlement during the Nabatean-Roman Period may have been favored by higher runoff. This was followed by a runoff decline in the Nahal Zin area apparently preceding the decline in Mount Sedom. The Roman-Byzantine runoff farming in Nahal Zin basin may have played a role, albeit a slight one, in preventing some of the runoff in this basin from reaching the main Nahal Zin channel. Since no farming was practiced on Mount Sedom, its record may better indicate natural hydrologic conditions without human interference. The decline of settlement around the end of the Justinian period cannot be adequately explained simply on the basis of the evidence presented here. In future discussions, climatic and hydrologic change impact must be considered together with the role of political, economical and sociological factors influencing human settlement in the Negev.

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References

Baker, V.R. (1987) Paleoflood hydrology and extraordinary flood events. *Journal of Hydrology* 96, 79–99.

Begin, Z.B., Nathan, Y. and Ehrlich, A. (1980) Stratigraphy and facies distribution in the Lisan Formation new evidence from the area south of the Dead Sea, Israel. *Israel Journal of Earth Sciences* 29, 182–189.

Bentor, Y.K. and Vroman, A. (1951) *The Geological Map of the Negev, 1:100,000, Sheet 18: Abde (Avdat)*, 1st ed. Geological Survey of Israel, Tel-Aviv (in Hebrew).

Bruins, H.J. (1986) *Desert Environment and Agriculture in the Central Negev and Kadesh-Barnea During Historical Times*. Midbar Foundation, Nijkerk.

Bull, W.B. (1991). *Geomorphic Responses to Climatic Change*. Oxford, New York.

Bull, W.B. and Schick, A.P. (1979) Impact of climatic change on an arid watershed: Nahal Yael, Southern Israel. *Quaternary Research* 11, 153–171.

Ely, L.L., Enzel, E., Baker, V.R., and Cayan, R.C. (1993). A 5000 years record of extreme floods and climate change in the southwestern United States. *Science* 262, 410–412.

Enzel, Y. (1992) Flood frequency of the Mojave River and the formation of late Holocene lakes, southern California, U.S.A. *The Holocene* 2, 11–18.

Evenari, M., Shanan, L. and Tadmor, N. H. (1971). *The Negev the Challenge of a Desert.* Harvard University Press, Cambridge, Mass.

Frumkin, A. (1992) The Karst System of the Mount Sedom Salt Diapir. Ph.D. thesis, The Hebrew University, Jerusalem (in Hebrew, English abstract).

Frumkin, A. (1994) Hydrology and denudation rates of halite karst. *Journal of Hydrology* **162**, 171–189.

Frumkin, A. (1996) Determining the exposure age of a karst landscape. *Quaternary Research* **46**, 99–106.

Frumkin, A. (1996) Uplift rate relative to base level of a salt diapir (Dead Sea, Israel), as indicated by cave levels. *In* I. Alsop, D. Blundell and I. Davison (eds.), *Salt Tectonics,* Special Publication no. 100. Geological Society, London, pp. 41–47.

Frumkin, A. and Ford, D.C. (1995) Rapid entrenchment of stream profiles in the salt caves of Mount Sedom, Israel. *Earth Surface Processes and Landforms* **20**, 139–152.

Frumkin, A., Magaritz, M., Carmi, I. and Zak, I. (1991) The Holocene climatic record of the salt caves of Mount Sedom, Israel. *The Holocene* **1**(3), 191–200.

Gerson, R. (1972) Geomorphic Processes of Mount Sdom. Ph.D. thesis, The Hebrew University, Jerusalem (in Hebrew, English abstract).

Goldberg, P. (1984) Late Quaternary history of Qadesh Barnea, northern Sinai. *Zeitschrift für Geomorphologie* NF28(2), 193–217.

Greenbaum, N. (1996) Paleofloods in the Large Ephemeral Streams of the Central Negev. Ph.D. thesis, The Hebrew University, Jerusalem (in Hebrew, English abstract).

Greenbaum, N., Schick, A.P. and Baker, V.R. (in press) The paleoflood record of a hyperarid catchment, Nahal Zin, Negev Desert, Israel. *Earth Surface Processes and Landforms.*

Huntington, E. (1911) *Palestine and its Transformation.* Houghton Mifflin, Boston.

Issar, A.S. and Tsoar, H. (1987) Who is to blame for the desertification of the Negev, Israel? *In The Influence of Climate Change and Climatic Variability on the Hydrologic Regime and Water Resources,* IAHS, Vancouver, **168**, 577–583.

Issar, A.S., Govrin, Y., Geyh, M.A., Wakshal, E., and Wolf, M. (1991) Climatic changes during the Upper Holocene in Israel. *Israel Journal of Earth Sciences* **40**, 219–223.

Katsnelson, J. (1979) The Arava rainfall. *In* A. Schmueli and J. Gradus (eds.), *The Negev Land,* Ministry of Defence Publication, Tel-Aviv, pp. 51–73 (in Hebrew).

Kedar, Y. (1957) Water and soil from the desert: Some ancient agricultural achievements in the central Negev. *Geographical Journal* **123**, 179–187.

Kloner, A. (1975) Ancient agriculture at Mamshit and the dating of the water-diversion systems in the Negev. *Eretz-Israel, Israel Exploration Society* **12**, 167–170.

Knox, J. (1993) Large increase in flood magnitude in response to modest changes in climate. *Nature* **361**, 430–432.

Mayerson, P. (1960) The ancient agricultural remains of the central Negeb: methodology and dating criteria. *Bulletin of the American Schools of Oriental Research (BASOR)* **160**, 27–37.

O'Connor, J.E. and Webb, R.H. (1988) Hydraulic modeling for paleoflood analysis. *In* V.R. Baker, R.C., Kochel and P.C Patton (eds.), *Flood Geomorphology,* Wiley, New York, pp. 383–402.

Patton, P.C., Baker, V.R. and Kochel, R.C. (1979) Slack-water deposits: A geomorphic technique for the interpretation of fluvial paleohydrology. *In* D.D. Rhodes, and G.P. Williams (eds.), *Adjustment of the Fluvial System,* Kendall/Hunt, Dubuque, Iowa, pp. 225–253.

Rubin, R. (1989) Sources of water supply for Negev settlements in the Nabataean, Roman and Byzantine Period. *In* D. Amit, Y. Hirschfeld and J. Patrich (eds.), *The Aqueducts of Ancient Palestine,* Yad Izhak Ben Zvi, Jerusalem, pp. 313–324.

Schick, A.P. (1988) Hydrologic aspects of floods in extreme arid environments. *In* V.R. Baker, R.C. Kochel and P.C. Patton (eds.), *Flood Geomorphology,* Wiley, New York, pp. 189–203.

Shanan, L. (1975) Rainfall and Runoff Relationships in Small Watersheds in the Avdat Region of the Negev Desert Highlands. Ph.D. thesis, The Hebrew University, Jerusalem (in Hebrew, English abstract).

Sharon, D. and Kutiel, H. (1986) The distribution of rainfall intensity in Israel, its regional and seasonal variations and its climatological evaluation. *Journal of Climatology* 6, 277–291.

Stuiver, M., Long, A. and Kra, R.S. (1993) Calibration 1993. *Radiocarbon* 35(1), 1–244.

Yair, A. (1983) Hillslope hydrology, water harvesting and areal distribution of some ancient agricultural systems in the northern Negev Desert. *Journal of Arid Environments* 6, 283–301.

Yair, A. and Lavee, H. (1982) Factors affecting the spatial variability of runoff generation over arid hillslopes, southern Israel. *Israel Journal of Earth Sciences* 31, 133–143.

Zak, I. (1967) The Geology of Mount Sedom. Ph.D. thesis, The Hebrew University, Jerusalem (in Hebrew, English abstract).

Chapter 6

Climate Change and History during the Holocene in the Eastern Mediterranean Region

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6.1. Introduction

A wide spectrum of data obtained as a result of palynological, hydrological, geological, paleontological and archaeological investigations is enabling us to reconstruct quite comprehensively and with high definition the climate changes which have occurred in the Mediterranean region during the last 10,000 years. One may start with indicators not influenced by human activity, like isotope ratios, or ancient Mediterranean and Dead Sea levels; proceed with such environmental proxy-data, as pollen time series; and then add to this archaeological and human historical data. One may thus obviate the 'danger of arguing in circle' which is the danger of 'inventing climatic change to explain events in human history, which are then held to prove the occurrence of a change of climate' (Lamb, 1982).

This affords also an answer to the central problem posed in the first place with the research presented herewith: namely to explain the appearance and disappearance of agricultural settlements along the borders and inside the deserts of the Middle East. Today it can be concluded that the main factor responsible for the prospering or and decline of cultures during historical times in semi-arid regions was climate. In other words, it was not the gratuitous mismanagement of natural resources by man that destroyed society and culture. Instead climate deterioration was the factor which triggered the deterioration of human socio-economic systems. At other times climatic conditions turned extensive parts of the desert into potentially fruitful land. It is not to say that the human factor was insubstantial.

A. S. Issar and N. Brown (eds.), Water, Environment and Society in Times of Climatic Change, 113–128.

But compared to the natural factor, man's activities were secondary in their impact—a response in order to survive.

Once the inter-relation between climate and the settling of the desert by man was ascertained, the scope of this research was widened in order to evaluate the future impact of the greenhouse effect on the hydrological cycle. This approach can be summarized by the sentence: The past is a key to the future.

A progressive advance from time series which could not have been a result of human activity to anthropogenic proxy-data is the basis of the results presented in this paper. Thus the first stage draws on isotopic data from a core hole in the Sea of Galilee; and then data from speleothemes found in caves in the Galilee. Palynological and sedimentological data, ancient levels of the Mediterranean and Dead Sea were added later. So were archaeological records regarding phases of settlement and desertion. Next were derived the most pronounced co-varia revealed in a preliminary survey involving a division of the Holocene into time stratigraphic units. Where climate and historical changes occurred more or less simultaneously, a causal link could provisionally be inferred.

6.2. Time Stratigraphic Units

6.2.1. The Neolithic Period (10,500 BP–6,200 BP)

6.2.1.1. *Pre-Pottery Neolithic A (PPNA), cold and humid period*

The regional climate during the Pre-Pottery Neolithic A, 10,500–9,700/9,500 BP (PPNA) was probably cool and humid as can be concluded from the depletion in the ^{18}O in the sediments of the rain-fed Lake Van (Schoell, 1978). This most probably facilitated the establishment of sedentary communities organized in small villages in the more humid parts of the eastern Mediterranean, while the semi-arid zones could support only a hunting economy (Bar Yosef, 1986a). Van Zeist (1969) suggested that the Near East suffered from the rise of temperature during the PPNA; and that the climate became too dry to allow the cultivation of barely and wheat. So the grains found there will have been imported. This explanation looks, to the author of the present paper, rather complicated.

6.2.1.2. *Pre-Pottery Neolithic B (PPNB), warm and humid period*

The climate during the PPNB (9,700/9,500–8,000 BP) seems to be warmer, as the oxygen isotopes of Lake Van become heavier (Schoell, 1978). Also Horowitz (1977) pointed out that according to the pollen spectrum found in the sediments of the PPNB (Sde Divshon), this period was more humid than today. The more well known sites from that period of time are Sde-Divshon (from the Avdat area), some sites from Ramat Mitrad, Nahal Lavan and Halutza (Gopher, 1981). There are well known sites from other desert areas of the Levant: Eastern Jordan and Southern Jordan

(Avner *et al.*, manuscript). Also the climate allowed agriculture even on the desert margins. This period was concurrent with the Boreal period of Europe (Horowitz, 1980). It was characterized in the Levant by a transient increase in the quantity of precipitation.

According to Bar Yosef (1986b) the economy of the PPNB sites in fertile areas was based on legumes and cereal cultivation together with hunting and herding. In the drier areas the inhabitants probably lived in established sites during the winter, autumn and Spring with a summer economy based on hunting and gathering. One should consider the possibility that the humidity, during this period, in the southern more arid parts of the Levant was due to an extension of the Indian monsoon

6.2.1.3. *Pottery Neolithic Period (8,000–7,500 BP) PNA, mainly cool and humid*
Evidenced by the depletion of ^{18}O of Lake Van (Schoell, 1978), a higher level of the Dead Sea (Frumkin *et al.*, 1991), and a low level of the Mediterranean (Raban and Galili, 1985) the climate during the lower and mid parts of the Pottery Neolithic period (8,000–7,500 BP) seems to have been cooler as well as more humid. This climate may have caused severe flooding of Neolithic sites which were located close to river beds. Considering that the Neolithic inhabitants, before the 8th millennium BP, appear to have been a peaceful society, Bar Yosef (1986b) suggests that the Neolithic walls of Jericho were for flood protection. Very few sites from the Pottery Neolithic were found in the northern part of the Negev or Southern Sinai. On the other hand, six sites were discovered in the Southern Negev. Most of the Pottery Neolithic sites in Israel are found in the Jordan Valley and the coastal plain. Bar Yosef (1986b) explained this phenomena of scarce sites not as a consequence of climatic condition but as a consequence of social and economic systems. Towards the end of the Neolithic, namely 7,500–7,000 BP the climate seems to become warmer once again as the isotopes of Lake Van become heavier (Schoell, 1978).

Thus except for two rather short intervals, one at the beginning and the other in the middle, the Neolithic climate was warm. This may be correlated with the results reported by Rossignol-Strick *et al.* (1982) who found that the ratio of $^{18}O/^{16}O$ of the epipelagic foraminifera Globigerinoides rubes reached minimal values between 8,000 and 6,000 BP; and the results reported by Luz and Perelis-Grossowicz (1980) of a ^{18}O minimum at *ca.* 8,700–6,500 BP. Lowe *et al.* (1980, cited in Rossignol-Strick *et al.*, 1982 and Luz, B. and L. Perelis-Grossowicz, 1980) explained part of the negative $d^{18}O$ by a fresh water influx which formed low salinity surface water (which also accounts for the sapropelic layer). The sharp depletion signals a general lowering in salinity and probably connotes heavy Nile floods which also correlate with the period from 8,000 BP to 7,000 BP. Yet the Nile water originating in low latitudes ought to be isotopically heavy. In a more recent work, however, Luz (1991) advanced an alternative explanation for the isotopic depletion: namely, a high influx of low salinity water entered the Mediterranean Sea from the Black Sea as rising sea surface reached above the threshold of the Bosporus.

6.2.1.4. *Pottery Neolithic Period B (7,500–6,200 BP) PNB, mainly warm and dry*
The heavier isotopic composition of the sediments of Lake Van (Schoell, 1978) suggest that the climate during this period was warmer and drier.

6.2.2. The Chalcolithic Period (~6,200–5,100/5,200 BP)

6.2.2.1. *The Lower and Middle Chalcolithic (6,200–5,500?), cold and humid*
Most of the Chalcolithic period was colder than the Neolithic. As much as can be concluded from the depletion in the ^{18}O isotopes of Lake Van (Schoell, *op. cit.*). Also the oxygen isotope curve of the Sea of Galilee (Stiller *et al.*, 1984) shows a depletion around 5,500 BP. From archeological data which will be further discussed, it seems that climate was more humid in the Levant. No ancient high level of the Dead Sea was reported by Frumkin (*op. cit.*). But this may be explained by the fact that its being obliterated by the posterior higher lake level of the Early Bronze.

During the Chalcolithic period, there was a sequence of settlements in the Negev from the Beer-Sheva Valley, the Arad Valley and Nahal Besor towards the south—to the Negev mountains. Towards the east, frontier of settlements reached the margins of the Arava Valley up to Ein Yahav. In the south, the settlement reached Timna area. Most of the settlements were located on the top of low hills close to river valleys (Cohen, 1989). The population of the Chalcolithic period settled in Israel in farming communities in the Jordan Valley, the coastal plain, the Judean Desert and the northern and eastern Negev (Ussishkin, 1986).

Sedentary village life had been established during the Neolithic period but the Chalcolithic communities were in larger and more advanced farming villages. A famous Chalcolithic site was at Tel el-Ghasul in the eastern Jordan Valley near Jericho (Hennessy, 1967). The center of the Chalcolithic settlement in the Negev was the Beer-Sheva Valley. The archaeological remains from the Beer-Sheva area point to the importance of agriculture to Chalcolithic society. Cohen (1989) claimed that the settlements in the Negev mountains were temporary and semi-nomadic settlements, based on pastoral grazing and on the transportation of copper from the Feiran area and the Timna Valley. The discovery of mining and smelting sites in the Feiran area, the Timna Valley and the Elat area suggests the importance of mining and metal production.

6.2.2.2. *The Upper Chalcolithic (5,500–5,200 BP), warm and dry*
The magnificent Chalcolithic culture, with its artistic tradition and technical knowledge, lasted but a few hundred years. It suddenly disappeared in an unknown way like it had started. All the sites were abandoned without any signs of violence. There are several hypotheses concerning this disappearance of the Chalcolithic culture. Hennessy (1969) suggested that they had to leave their settlements because a migration by a new wave of people—the ones who established the Early Bronze Age culture. But archaeological remains do not reveal any transition of cultures

within the same sites. Others have explored the circumstantial connection between the disappearance of the Chalcolithic population and the expansion of the first Egyptian Kingdom of Naarmer at about 5,000 years BP. But there are no indications of violence during the abandonment of the settlements.

The ^{18}O fraction becoming heavier can be seen in the curve of Lake Van and Sea of Galilee towards 5,000 BP. This suggests a warming and therefore drying of the climate as a reason for the collapse of the whole system. The fact that the settlements in the area of Beer-Sheva and of the Judean Desert were the last to be abandoned (Gonen, 1986) can be explained by the fact that by this time the Chalcolithic people had found a way to dig wells into the shallow gravel aquifer in the Beer-Sheva river bed, while in the Judean Desert the people maintained a pastoral way of life, and moved from one flowing spring, which still continued to flow, to another (Govrin, 1991).

Information from deep sea cores in the Eastern Mediterranean (Luz, 1979; Luz *et al.*, 1980) have indicated a depletion in the ratio of $^{18}O/^{16}O$ in planktonic foraminifera between *ca.* 6,200–5,300 BP. As discussed earlier, it is not clear yet what may have determined the isotopic composition of the water of the eastern Mediterranean, whether the dynamics of the sea itself or changes in the supply from the Nile or the rivers draining to the Black Sea. It is best to refrain from judging the evidence until additional data is provided—for example, from sediments characteristic of the water of the Nile.

6.2.3. The Bronze Age (5,200–3,200 BP)

6.2.3.1 *Early Bronze Period (EB) (5,200/5,100–4,000 BP)*

The depletion of the ^{18}O and ^{13}C values of carbonates deposited in the Sea of Galilee (Stiller *et al., op. cit.*) of the ^{18}O of the sediments of Lake Van together with a rise in the water level of the Dead Sea (Frumkin *et al., op. cit.*) while the level of the Mediterranean dropped (Raban and Galili, *op. cit.*) all show that, during the EB (5,100–4,000 BP), the climate was colder and more humid.

This change of climate coming as it did after the warm and dry period of the Upper Chalcolithic, caused a severe impact on the socio-economic state of affairs in the region. At the beginning of the 5th millennium BP agricultural society went through cultural, structural and organizational changes which led to the emergence of the walled city. Soon the city achieved a central role as a religious and economic center. The Early Bronze Age started about 100 years before the rule of the first king of the first dynasty in Egypt by the name of Naamer. Based on the intermediate chronology system of Egypt, it is suggested, that the beginning of the EB is 5,200/5,100 BP (Wright, 1971; Ben Tor, 1989).

However, there are some questions concerning the beginning of the EB. It is not yet clear whether it started immediately after the end of the Chalcolithic period or if there was an overlap between those two cultures or maybe a gap in time existed between the end of the Chalcolithic and the beginning of the EB I. Until the middle 1970s, the transition from the Chalcolithic period to the EB I was explained by diffusionist theories (Kenyon, 1979; Kenyon *et al.*, 1971; Lapp, 1970; de Vaux, 1971). Today, there is evidence for the claim that the transition from the Chalcolithic period to the EB I was a process of local evolution, rather than a diffusion from the outside (Levy, 1986; Amiran and Kochavi, 1985; Ben Tor, 1989).

The settlement size of the EB I was almost the same as during the Chalcolithic period but the density in the average settlement increased. This is an indication of an increase in population (Ben Tor, 1989). However, not all the settlements during the EB I are defined as cities. Some were small cities' such as Meggido, Lachish and Jericho. Others were considered 'big cities' such as Yarmut, Gezer, Afek, Beit Yerach, Tel Erani. Agricultural villages were located alongside the cities.

There is also a transition of EB settlements to the mountainous areas, the Shefelah and the valleys of Israel. New sites locations sites meant a transition to a Mediterranean mode of agriculture. Ben Tor (1989) and Broshi and Gophna (1984) claimed that the people of the EB preferred mainly areas in which the annual amount of rainfall was more than 300 mm. But more than 600 sites from the EB I and II were discovered in the Negev Highlands and Uvda Valley (Avner *et al.*, manuscript).

All the EB settlements of the Negev were unwalled settlements and most of them were located on hills, near valleys which had permanent water resources (Cohen, 1989). The same author claimed that the population growth of the Negev Highlands was due to the transition of settlers who could not adapt themselves to the emergent urban life style in the northern parts of Israel. The economy of the EB settlements in the Negev was based on agriculture, pastoralism and hunting.

As can be seen from the isotope data as well as from sea and lake levels (Fig. 6.1), towards the end of the 5th millennium, the climate became warmer and drier. This brought Issar (1990) to suggest that it was not the ancient Mesopotamian people who caused the desertification of their land at *ca.* 4,000 BP—but rather global warming, which caused the flow of the Tigris-Euphrates to decline. The lack of water did not allow proper flushing of the soils. This brought the Sumerian people to grow barley, which is more tolerant to dry conditions than wheat. Moreover a considerable effort had to be invested in digging long canals. Weiss *et al.* (1993) came to a similar conclusion concerning the end of the Akkadian empire at the end of the 5th millennium BP.

With this gradual warming and drying of the climate came a gradual transition, observed by the archaeologist through EB II (4,900/4,950–4,700 BP) to the EB III (4,700–4,350 BP). This was manifested by transition from unwalled settlements to fortified cities (Ben Tor, 1989; Broshi and Gophna, 1984), as well as the desertion of settlements in the more arid parts, like the city of Arad, which was abandoned at the end of the EB II (4,650 BP) and was not settled again until the Iron Age (Amiran, 1986). The same is true for the settlements in northern Sinai along the road to Egypt, (Richard, 1980). Thus by the EB III no settlements remained in the Negev or Sinai.

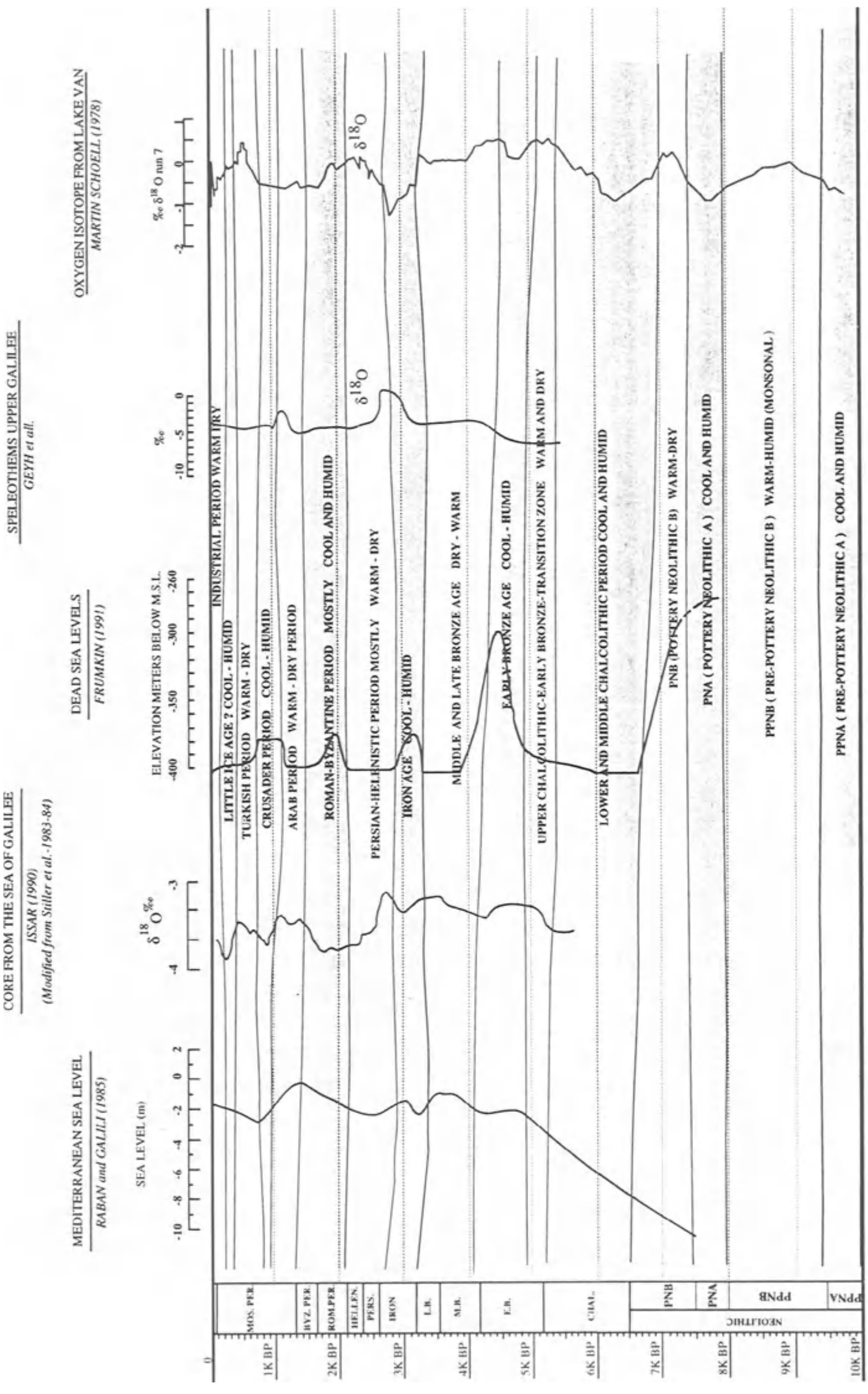

Figure 6.1 Correlation diagram of time-series of proxy-data in the Eastern Mediterranean basin.

6.2.3.2. *Middle Bronze Period (MB) (4,100/4,000–3500 BP)*
As can be seen from Figure 6.1, the ^{18}O of Lake Van as well as the ^{18}O and ^{13}C of the speleothemes of the caves of the Galilee show a trend of becoming heavier from about 4,500 BP. The same can be said for the ^{18}O of the Sea of Galilee at about 4,000 BP. The level of the Dead Sea dropped, until its southern shallow part totally dried up (Frumkin *et al., op. cit.*), while the level of the Mediterranean was rising. All this points to the conclusion that the warming and drying trend which started in the EB III and affected first the communities, like Arad, along the border of the desert came to its peak some time between 4,300 to 3,500 BP. It is suggested it be called the Middle Bronze Warm Period. Cultural considerations have brought the archaeologists to start the Middle Bronze Age at *ca.* 4,100 BP. Yet from the climatic point of view, it is better to include the EB IV (4,350–4,100 BP) in the Middle Bronze.

During the EB IV (4,350–4,000 BP) the abandonment of cities came to a climax. At the end of this stage all the urban centers, which had existed for several hundreds of years, disappeared (Dever, 1980, 1985). The same is true throughout the Middle East (Butzer 1958:116–118; Harlan 1982). Dever (1987) called this period of time the 'dark age' of the EB.

The change towards a warmer and less humid climate can explain also the migration of the Amorites, a semi-nomadic people who overran the EB III cities causing deurbanization. However archaeologists see this migration as caused by the socio-economic changes in the Middle East (Albright, 1951; Kenyon, 1971). It is argued that the onslaught of the Amorites was another result of a severe climate change, globally.

As the isotope curve of Lake Van indicates, the climate after 4,000 BP may well have started to become a bit cooler and probably more humid, This had a positive influence wherever the average precipitation was on or above the threshold of aridity (300 mm yearly rainfall). Settlements soon flourished in these parts of the Levant, though the desert border areas remained unsettled. An additional reason for the rejuvenation of the MB cities was the development of sophisticated rainwater management and water storage systems. The technology of excavating wells in the rock was probably first known in the Middle Bronze Age (Miller, 1980).

The MB I (4,000–3,800 BP) is therefore considered an urban period. Walled cities characterized that period of time, though the peasants and agriculture systems were also very important components in the social structure. The economy was based on agriculture with some local crafts, domestic industry and trade in luxury items. Meanwhile the political organization of the MB I did not yet favor centralization or the domination of one city. It seems that a tribal-patriarchal system was still dominant. The political urban regime, based on one ruler in each urban center, crystallized only during the MB II (Mazar, 1967). On the one hand, the MB II city was characterized by social complexity and integration. But on the other, it was still far from the highly political developed city that characterized Mesopotamia and Egypt just before 5,000 BP (Dever, 1987).

Magaritz and Kaufman (1973) found signs of excess of evaporation over precipitation in the water of the Eastern Mediterranean from about 4,900 BP to 2,800 BP (2900–800 BC). This evidence mainly concerns the ^{18}O composition of the marine sediments. These authors explained this as a result of a change in wind direction or due to a tectonic change which caused a rise in the bed of the Sicilian strait and impaired contact between the Eastern Mediterranean and the Western Mediterranean. It is the opinion of the present author that applying Occam's razor, one would prefer an explanation which gives a rise in temperature as the reason to this excess. Meanwhile the deposition of sands and sandy silts in the river bed of Lachish during the MB I and MB II may be interpreted as 'rapidly fluctuating rainfall patterns interspersed with drought leading to soil stripping from the hill slopes' (Rosen, 1986:56–57).

Some evidence from pollen diagrams from northern Turkey shows that, at about 4,000 BP, Pinus pollen percentages increased (van Zeist and Bottema, 1982). If this change in vegetation was brought about by a natural factor, this would be drier climate conditions after 4,000 BP. However, van Zeist and Bottema (1982) cited anthropogenic factors as the reason for the changes in the pollen. Also marked decline in deciduous oak pollen percentages at about 4,500–3,500 BP in Gahb valley-northwestern Syria (Niklewski and van Zeist, 1967, cited in van Zeist and Bottema, 1982) is said by van Zeist and Bottema (1982) to be due to the activity of man. But a decline in arboreal pollen is shown also in a diagram from a core in the Hula basin (Tsukasa, cited in van Zeist and Bottema, 1982). After 4,500 BP there was a decline in Olea, Pistachio and Quercus. It is interesting to note that the same change is observed in the core of the Sea of Galilee during the same period of time (Baruch, 1986). One should therefore consider the possibility that the prime factor was climate change: namely, that a more humid climate encouraged the cutting of the natural vegetation and the planting of Olea. Later, warmth and dryness caused the decline in the income from the olive orchards in the drier regions and the desertion of the plantations. This brought the oak and pistachea habitat to renew itself. Neumann and Sigrist (1978) undertook a survey of references to barley harvest dates in the clay tablets of Ancient Babylon. They concluded that the period 3,800–3,650 BP was warmer than the present; and that this allowed the harvest to start 10–20 days earlier.

6.2.3.3. *The Late Bronze Age (3500–3200 BP) and Early Iron Age (3,200–3,000 BP), mainly cold and humid*

From *ca.* 3,500 BP the start of a new cold and humid climate can be inferred from the depletion in ^{18}O composition in the sediments of Lake Van, the Sea of Galilee and the speleothemes of the caves in the Galilee, where the ^{13}C follows suite. In the same time the Mediterranean sea level receded to about 1.0 m lower than the present, while the level of the Dead Sea rose to about 370 m below MSL (Fig. 6.1). It seems from the proxy-data presented in Figure 6.1 that this cold phase continued until about 3,000 BP, namely the middle of the Iron Age. This leads one leads to propose combining the Late Bronze and first half of the Iron Age into one time-stratigraphic unit.

The comparatively cool and more humid climatic conditions of the Late Bronze Age allowed a recovery of Canaanite urban life. The spread of the Late Bronze Canaanite cities was along the coastal plain from Sinai to the Carmel, the upper Jordan valley, Jizrael valley and Beit Shean valley. The mountainous areas were little settled. According to Gonen (1986) only a few settlements existed in Judea and Samaria mountains. In the Galilee almost no settlements were resumed. The few in the mountains were populated by Semite tribes (Amaru), Hurian, Hittite, Jevusite, *etc.* The nomads of that period settled in the mountains.

Most of the cities were located on artificial hills (Tel) on the ruins of the Middle Bronze Age cities. A city was usually close to a reliable source of water (Aharoni, 1986, Issar, 1990) to enable water supply to the city. The springs were usually outside the city but the water supply system for peaceful days and war time shows a very sophisticated technical ability.

6.2.3.4. *The Late Iron Period (3,000–2,200 BP), warm and dry*

Towards 3,000 BP a warm and dry trend was reestablished. Neumann and Parpola (1987) reviewed textual and nontextual environmental evidence in order to ascertain the magnitude of the change. According to them 3,200–2,900 BP was a warming and drying period. What is not yet clear is whether it was only this climate change that caused the widespread migration of population, which occurred throughout the eastern Mediterranean or whether this was a combined effect of a cold period then a warm one.

It is the author's opinion that this movement of nations began already during the cold period of Late Bronze and Early Iron Age, which pushed towards the south the people of the northern plains of Central Asia causing a 'domino effect' which reached the countries of the Eastern Mediterranean two centuries later. Another possibility is that here two factors combined: namely, that the socio-economic tumult was triggered by Hellenic tribes, who started their move south-westward, due to the cold spell of the Late Bronze, beginning of Iron Age. Then as they started to penetrate into the Levant, this region started to go through an economic crisis, due to the warming of the climate. This may explain the destruction of the great Mycenaean culture in the Aegean Sea, where there is no evidence of newcomers and violence but there are signs of abandonment and migration There exist reasons to think that the Dorians entered the Peloponnese after its desertion (Carpenter, 1966; Weiss, 1982).

This movement of people (leaving their places due to factors environmental, socio-economic, military or all combined together) brought a tremendous change to the Eastern Mediterranean region. Among the nations which catastrophically suffered were the Hittites, whose empire collapsed and disappeared (Carpenter, 1966; Wiess, 1982). The Sea People moved down the eastern coast of the Mediterranean and fought the Egyptians and later settled down in Phillastea. Big city states along the Syrian coast such as Ugarit and Alalach were destroyed and never reoccupied. Cyprus was devastated around the year 1200 BC. Assyria underwent several revolts which weakened the empire.

The Semite tribes (the Apiru) moved to Canaan; and the Hittites to northern Syria below the Taurus mountain rampart (Carpenter, 1966; Weiss, 1982). Very important cities in Canaan such as Hazor, Lachish, Beit Shean, Megiddo, Afek, Beit Shemesh, Gezer, Tel Beit Mirsim and others were destroyed (Mazar, 1990). Some of these cities were never rebuilt. Others never regained their former importance (Hazor, for example). Some of them recovered only at the beginning of the 12th century BC.

Climate began to improve once more about 2,800 BP (800 BC). Neuman (1985) by quoting references concerning climatic change in the classical Greek and Roman literature came to the conclusion that northern Italy (and probably the whole Mediterranean area) was affected by a relatively cool and wet climate between 2,800 and 2,400 BP. As can be seen from the proxy-data (Fig. 6.1), these moderate conditions, persisted with very little change until the beginning of the Roman period, *ca.* 2,200 BP.

6.2.4. The Roman-Byzantine period (2,200–1,300 BP), mostly cold and humid

6.2.4.1. *Roman Period (2,200–1,700 BP), cold and humid*

From the ^{18}O isotope curve of the core from the Sea of Galilee and the rise of the Dead Sea level, it can be deduced that around 2,300 BP the climate became colder and more humid. The ^{18}O curve of Lake Van shows a short depletion trend at the same time. On the same curve, an opposite trend followed at *ca.* 1,800 BP, most probably due to a warm period of about a century. At about 1,500 BP, there is evidence of warming up, which, according to the isotope curves of the Sea of Galilee and the stalagmites in the caves of Galilee, peaked at *ca.* 1,200 BP. Thus it may be concluded that during most of the period in which the Roman and Byzantine empires ruled the Mediterranean basin the climate was colder and thus more humid. This looks to have been an optimal climate for the spread of agricultural settlements into the desert margins of the Mediterranean region.

Liphschitz *et al.* (1981) studied the assemblage of trees from the Roman siege rampart in Masada (*ca.* 70 AD) and found that acacia are rare relatively to tamarisk, in comparison with the present balance. They saw this as an indication that the climate was wetter during the 1st century AD, as tamarix thrive above shallow water tables. Yakir *et al.* (1994) investigated the isotopic ratios ($^{13}C/^{12}C$ and $^{18}O/^{16}O$) of the cellulose of the tamarix trees from this siege rampart, and compared it with ratios measured in present day tamarix trees growing in the Masada region. It was found that the ancient tamarix cellulose is more depleted in both ^{13}C and ^{18}O. This points to the conclusion that the ancient trees enjoyed more humid conditions during their growth compared to contemporary trees in this arid region.

Palynological data taken by Baruch (1986) from a core in the Sea of Galilee show a peak in the olive pollen and a fall in that of oak and pistachea during the Roman period. Baruch *(op. cit.)* concluded that the economy flourished during the Roman period and thus the farmer cut the forests and grew olives. Conversely the decline in Olea pollen at *ca.* 500 AD was due to economic crisis in which inflation increased and caused the export of oil product to be too expensive. But, as already discussed, the abundance of olive and decline of oak pollen, in the same core, during the Early Bronze period was concurrent with an expansion of settlements in the Negev. The conclusion to draw is that both the expansion of the olive groves and the cutting down of the oaks is due to better climatic conditions and vice versa.

A noticeable expansion of settlements was that by the Nabateans. Negev (1979) views the history of the settlement of the Negev and Transjordan by the Nabateans, from the 4th century BC until the 7th century AD as one chapter. It is interesting to note that their expansion came from inside the desert, towards its margins; and this caused the collision with the Hashmonean kingdom of Judea, which tried to expand into the same region. In 106 AD, the Romans annexed the Nabatean kingdom to their empire as Provincia Arabia.

6.2.4.2. *Roman-Byzantine Transition Period (1,700–1,800 BP), warm*

An abrupt change in the ^{13}C curve of the Sea of Galilee curve at about 1700–1800 BP indicates a short phase of warm climate. This same fluctuation caused the level of the sea to rise, and cover the floors of Roman buildings along the English coast.

6.2.4.3. *Byzantine Period (1,800–1,500 BP), cool and humid*

The proxy-data presented in Figure 6.1 suggests that the positive climatic conditions, which prevailed during the period of Rome's hegemony continued during that through which Constantinople ruled the eastern Mediterranean.

6.2.5. The Moslem Period (1,300 BP–500 BP), mainly warm and dry

6.2.5.1. *The Moslem-Arab Period (1,300–1,000 BP), warm and dry*

As already mentioned, a warmer climate started at *ca.* 1,500 BP and, with it, the gradual abandonment of the Nabatean cities of the Negev and the farms surrounding them. A very interesting description of the physical environment is found in a letter from the sophist philosopher Procopius (*ca.* 450–526 AD) to his friend Jerome in Egypt: "There will be a day when you will see Elusa again and you will weep at the sand being shifted by the wind stripping the vines naked to their roots" (cited in Mayerson, 1983:251–253).

The desertion of cities and farms accelerated after the Arab conquest (636 AD) and the reign of the Caliphs from Umayyad House (636 AD till the middle of the 8th century AD). However, the greatest crisis broke during the establishment of the House of Abas (the middle of the 8th century AD until the middle of the 9th century AD). The Umayyad rulers had retained the Byzantine settlement pattern and preserved its Christian administration. Nor did the Arab conquest cause the disappearance of settlements. According to the Papyri of Nitzana the settlement existed unchanged at least to the very end of the 7th century AD. At the first half of the 8th century AD Nitzana and other villages in the Negev were abandoned. Only some pieces of pottery and nine Umayyad coins before 750 AD were found (Lewis, 1948). Also none of the Greek Papyri is dated after 700 AD. According to an archaeological survey of the Negev (Cohen, 1985), only six Arab sites were discovered from the Umayyad period in comparison to 44 Byzantine sites. There is also a doubt whether the Arab sites continued to exist during the domain of the Abassids.

It is also important to note that the settlements which were totally deserted were those located south of the present 200 mm/year precipitation line, including the town of Beer-Sheva. Settlements north of this line, such as Gaza and Hebron, continued to exist though they did not flourish as in the past. Raban and Galili (1985) indicated that towards the end of the Byzantine period and during the Arab period the level of the Mediterranean Sea came up, indicating an increase in the global temperatures.

Carpenter (1966) maintains that the desertion of Greek settlements in the Peloponnese was necessitated by a severe long drought during the 7th and 8th century AD, and not by the Slav invasions. This drought came to an end by the 9th century AD and then 'southern Greece was once more fit to support an increasing population' (*ibid.*, 79).

6.2.5.2. *The Crusader Period (1,000–800 BP), cold and humid*

The proxy-data presented in Figure 6.1 show that around 1,000 BP there is a further colder period with more rain in the eastern Mediterranean basin. It caused the level of the Dead Sea to rise. This cold phase disrupts the so called 'Mediaeval optimum'.

6.2.5.3. *The Moslem-Turkish Period (800–500 BP), warm and dry*

There is a pronounced trend of the isotopes of the Sea of Galilee and of Lake Van to become heavier, while the level of the Dead Sea drops. Once again, the conclusion to draw is that this period was drier and warmer.

6.2.6. The Little Ice Age (500–100 BP), cold and humid

A depletion trend in the ^{18}O curves of the Sea of Galilee and Lake Van cores, show that between *ca.* 500–100 BP, the climate was colder and most probably more humid in the Mediterranean region.

6.2.7. The Industrial Period (100 BP to present)

The proxy-data presented in Figure 6.1 suggests a warmer phase since 100 BP.

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References

Aharoni, Y. (1986) The Land of Israel in the Late Canaanite period and the Israelite settlement period. *In* Y. Ripel (ed.), *The History of the Land of Israel,* Vol. 1, Israeli Ministry Publishing, Tel-Aviv, pp. 87–178 (in Hebrew).

Albright, W.F. (1951) Palestine during the Middle Bronze Age and the Late Bronze Age. *In The Archaeology of Palestine.* DAAT, Tel-Aviv, pp. 77–102 (in Hebrew).

Amiran, R. (1986) The fall of the Early Bronze Age II City of Arad. *Israel Exploration Journal* 36, 74–76.

Amiran, R. and Kochavi, M., (1985) The Land of Israel towards the end of the 3rd millennium B.C.—A separate culture or the last stage of Early Bronze Age? *The Land of Israel* 18, 361–365 (in Hebrew).

Avner, U., Carmi, I. and Segal, D. (manuscript). Settlement in the southern Negev from Neolithic to Bronze Age in light of radiocarbon dating.

Baruch, U. (1986) The late Holocene vegetational history of Lake Kinneret (Sea of Galilee). *Israel. Paleorient.* 12(2), 37–48.

Bar Yosef, O. (1986a) The prehistoric period. *In* Y. Ripel (ed.), *The History of the Land of Israel,* Vol. 1, Israeli Defense Ministry Publishing, Tel-Aviv, pp. 1–26 (in Hebrew).

——. (1986b) The walls of Jericho: An alternative interpretation. *Current Anthropology* 27(2), 157–162.

Ben Tor, A. (1989) The Early Bronze Age. *In Introduction to the Archaeology of the Land of Israel in the Biblical Times,* The Open University, Tel-Aviv, pp. 11–71 (in Hebrew).

Broshi, M. and Gophna, R. (1984) The settlements of Israel and its population during the Early Bronze Age. *In The Land of Israel* (Braver Book), Vol. 17, pp. 147–157 (in Hebrew).

Butzer, K.W. (1958) *Quaternary Stratigraphy and Climate in the Near East.* Dummler, Bonn.

Carpenter, R. (1966) *Discontinuity in Greek Civilization.* Cambridge, The University Press.

Cohen, R. (1985) *An Archaeological Survey of Israel. Sde Boker-East Map (168).* Antiquity Division Publishing, Jerusalem (in Hebrew).

——. (1989) The settlements in the Negev mountain from the 4th millennium B.C. until the 4th century B.C. *Kadmoniot* **83–84**, 62–81 (in Hebrew).

de Vaux, R. (1971) Palestine in the Early Bronze Age. *In* I.E.S. Edwards, C.J. Gadd and N.G.L. Hammond (eds.), *The Cambridge Ancient History of the Middle East,* Vol. 1(2), Cambridge: University Press, pp. 208–237.

Dever, W.G. (1980) New vistas on the 'EB IV-MB I' horizon in Syria-Palestine. *Bulletin of the American Schools of Oriental Research* 237, 31–59.

——. (1985) Relation between Syria-Palestine and Egypt in the 'Hyksos' period. *In* J.N. Tubb (ed.), *Palestine in the Bronze and Iron Ages. Papers in Honour of Olga Tufnell,* London: Institute of Archaeology, pp. 69–87.

——. (1987) The Middle Bronze Age. The zenith of the urban Canaanite Era. *Biblical Archaeology* **50,** 149–177.

Frumkin, A., Magaritz, M., Carmi, I, and Zak, I. (1991) The Holocene climatic record of the salt caves of Mount Sedom, Israel. *The Holocene* 1(3), 191–200.

Gonen, R. (1986) The Late Bronze Age. *In Introduction to the Archaeology of the Land of Israel in the Biblical Times.* Unit 7. The Open University, Tel-Aviv, pp. 90–169.

Gopher, A. (1981) *The Prehistory of the Central Negev.* The Society for Nature Protection, Sde Boker Field School (in Hebrew).

Govrin, Y. (1991) The question of the origin of early Arad's population. *In The Land of Israel. Researches in Geography and Archaeology,* Vol. 21, Ruth Amiran Book. The Society for Exploring the Land of Israel and its Antiquities Publishing, pp. 107–110 (in Hebrew).

Harlan, J.R. (1982) The Garden of the Lord: A plausible reconstruction of natural resources of Southern Jordan in Early Bronze Age. *Paleorient.* 8(1), 71–78.

Hennessy, B. (1967) Preliminary report on the first season of excavations at Teleilat Ghassul. *Levant* 1, 1–24.

Horowitz, A. (1977) The climate and the settlement in the Negev mountain during the last Quaternary. *In* E. Sohar (ed.), *The Desert: Past, Present, Future,* Reshafin Publishing, Tel-Aviv, pp. 52–59 (in Hebrew).

——. (1980) Palynology-climate and distribution of settlements in Israel. *Kadmoniot* 13(3–4), 51–52 (in Hebrew).

Issar, A. (1990) *Water Shall Flow from the Rock.* Springer Verlag, Heidelberg.

Kenyon, K.M. (1971) *Archaeology in the Holy Land.* New York, Norton.

Kenyon, K.M., Bottero, J. and Posener, G. (1971) Syria and Palestine *c.* 2160–1780 B.C. *In Cambridge Ancient History,* 3rd revised ed., Vol. I(2), Cambridge, Cambridge University, pp. 532–594.

Lamb, H.H. (1982) *Climate, History and the Modern World.* London, Methuen.

Lapp, P. (1970. Palestine in the Early Bronze Age. *In* J.A. Sanders (ed.), *Near Eastern Archaeology in the 20th Century—Essays in Honor of Nelson Glueck,* New York, Doubleday, pp. 101–131.

Levy, T.E. (1986) The Chalcolithic period. *Biblical Archaeologist* June, 83–108.

Lewis, N. (1948. New light on the Negev in ancient times. *Palestine Exploration Quaternary* **80,** 102–117.

Liphschitz, N., Lev-Yadun, S. and Waisel, Y. (1981) Dendroarchaeological investigations in Israel (Masada). *Israel Exploration Journal* 31, 230–234.

Lowe, J.J., Gray, L.M. and Robinson, L.E. (1980) *Studies in the Late Glacial of Northwest Europe.* Pergamon, Oxford.

Luz, B. (1979) Palaeo-oceanography of the post-glacial eastern Mediterranean. *Nature* 278(5707), 847–848.

——. (1991) Post-glacial Paleo-oceanography of the Eastern Mediterranean. International Workshop on Regional Implications of Future Climate Change. Apr. 28–May 2 (1991) Ministry of Environment, Jerusalem, Israel.

Luz, B. and Perelis-Grossowicz, L. (1980) Oxygen isotopes, biostratifigraphy and recent rates of sedimentation in the Eastern Mediterranean off Israel. *Israel Journal of Earth Sciences* **29,** 140–146.

Magaritz, M. and Kaufman, A. (1973) Changes in the isotopic composition of the East Mediterranean seawater during the Holocene. *Nature* 243, 462–464.

Mayerson, P. (1983) The city of Elusa in the literary sources of the fourth-sixth centuries. *Israel Exploration Journal* 33, 247–253.

Mazar, B. (1967) The Middle Bronze Age in the Land of Israel. *Eretz Israel* **8**, 216–230 (in Hebrew).

Mazar, E. (1990) The Iron Age I. *In Introduction to the Archaeology of Ancient Israel in the Biblical Period,* The Open University of Israel. Unit 8, pp. 11–82 (in Hebrew).

Miller, R. (1980) Water use in Syria and Palestine from the Neolithic to the Bronze Age. *World Archaeology* **11(3)**, 331–341.

Negev, A. (1979) The Nabateans in the Negev. *In* A. Shmueli, A. and Y. Gradus (eds.), *The Land of the Negev—Man and Desert,* 1, Israeli Defense Ministry Publishing, Tel-Aviv, pp. 226–269 (in Hebrew).

Neuman, J. (1985) Climatic change as a topic in the classical Greek and Roman Literature. *Climatic Change* **7**, 441–454.

Neuman, J. and Parpola, S. (1987) Climatic change and the eleventh-tenth century eclipse of Assyria and Babylonia. *Journal of Near Eastern Studies* **46**, 161–182.

Neumann, J. and Sigrit, R.M. (1978) Harvest dates in ancient Mesopotamia as possible indicators of climatic variations. *Climatic Change* **1**, 239–252.

Raban, A. and Galili, E. (1985) Recent maritime archaeological research in Israel—A preliminary report. *The International J. of Nautical Archaeology and Underwater Exploration* **14**, 4321–4356.

Richard, S. (1980) Toward a consensus of opinion on the end of the Early Bronze Age in Palestine-Transjordan. *Bulletin of American Schools of Oriental Research* **237**, 5–34.

Rosen, A.M. (1986) Environmental changes and settlement at Tel Lachish, Israel. *Bulletin of American Schools of Oriental Research* **263**, 55–60.

Rossignol-Strick, M., Nesteroff, W., Olive, P., and Vergnaud-Grazzono, C. (1982) After the Deluge: Mediterranean stagnation and sapropel formation. *Nature* **295**, 105–110.

Schoell M. (1978) Oxygen isotope analysis on authigenic carbonates from Lake Van sediments and their possible bearing on the climate of the past 10,000 years. *In* E.T. Degens and Kurtman (eds.), *The Geology of Lake Van,* The Mineral Research and Exploration Institute of Turkey, Ankara, pp. 92–97.

Stiller, M., Ehrlich, A., Pollingher, U., Baruch, U., and Kaufman, A. (1984) The Late Holocene sediments of Lake Kinneret (Israel)—A multidisciplinary study of a 5 m core. *In Geological Survey of Israel,* Ministry of Energy and Infrastructure, Jerusalem, pp. 83–88.

Ussishkin, D. (1986) The land of Israel during the Chalcolithic period. *In* Y. Ripel (ed.), *The History of the Land of Israel,* Vol. 1, Israel Defense Ministry Publishing, Tel-Aviv, pp. 47–60 (in Hebrew).

van Zeist, W. (1969) Reflections on prehistoric environments in the Near East. *In* P.J. Ucko, and G.W. Dimbleby (eds.), *The Domestication and Exploitation of Plants and Animals,* London: Duckworth, pp. 35–46.

van Ziest, W. and Bottema, S. (1982) Vegetational history of the Eastern Mediterranean and Near East during the last 20,000 years. In *Palaeoclimates, Palaeoenvironments and Human Communities in Eastern Mediterranean Region in Later Prehistory,* Oxford: British Archeological Report, International Series **133**, pp. 277–321.

Weiss, B. (1982) The decline of Late Bronze age civilization as a possible response to climatic change. *Climatic Change* **4**, 173–198.

Wiess, H., Coutry, M.A., Wetterstrom, W., Guichaard, F., Senior, L., Meadow, R., and Curnow, A. (1993) The genesis and collapse of third millennium north Mesopotamian civilization. *Science* **261**, 995–1004.

Wright, G.E. (1971) The archeology of Palestine from the Neolithic through the Middle Bronze Age. *Journal of the American Oriental Society* **91**, 276–293.

Yakir, D., Issar, A., Gat, J., Adar, E., Trimborn, P., and Lipp, J. (1994) ^{13}C and ^{18}O of wood from the Roman siege rampart in Masada, Israel (AD 70–73): Evidence for a less arid climate for the region. *Geochimica et Cosmochimica Acta* **58(16)**, 3535–3539.

Chapter 7

Population Growth and Decline in the Northern Part of Eretz-Israel during the Historical Period as Related to Climatic Changes

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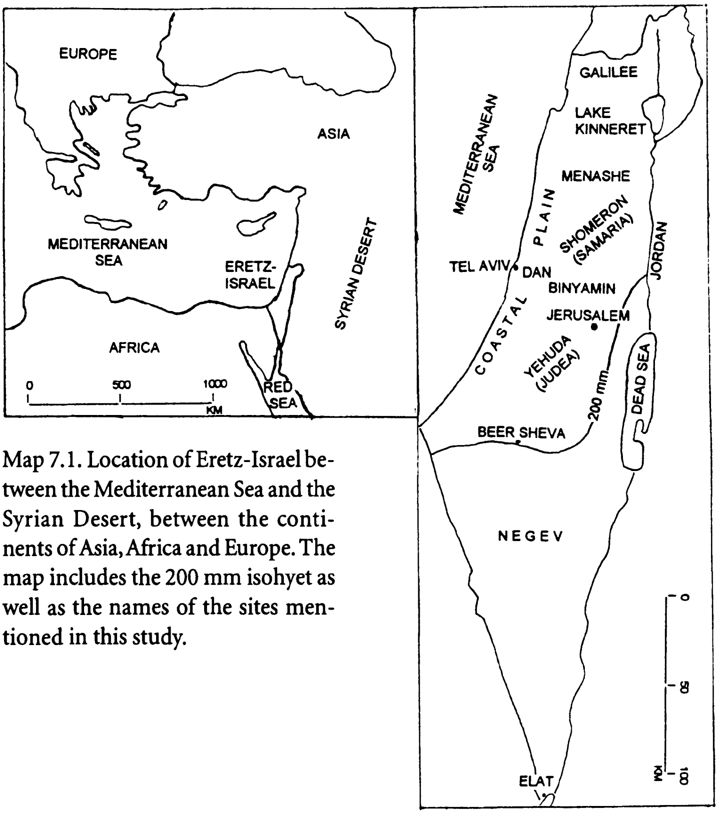
7.1. Introduction

Eretz-Israel (The Land of Israel) is located at the south-eastern corner of the Mediterranean Sea, between the Sea and the Syrian Desert. It is the transition zone between the Euroasian and African continents, and between the Mediterranean and the arid climatic zones (Map 7.1). The historical-geographical name Eretz-Israel applies to the region between the Mediterranean Sea to the Jordan River and the Dead Sea Rift Valley, which includes the State of Israel and the Palestinian Authority and was known until 1948 as the British Mandate of Palestine. This land has been settled by Man since the early stages of his history upon earth, since the Early Pleistocene. Archaeological findings and historical records present the history of the Land and its inhabitants during the historical and pre-historical periods.

Settlement and administration as well as the deportation of people through wars and conquests, changed the size of the population of Eretz-Israel during the historical periods. Thus, the population changes were generally related to political developments.

During the last two decades much evidence of climatic change have been detected in this Land, triggering the idea that the population changes are linked to the changes of climate. A clue to this is found in the Bible, as the Lord said to the People of Israel in the Sinai Desert after their Exodus from Egypt: "For the Land, whither thou goest to possess it, ... drinketh water of the rain of Heaven. ... I will give you the rain of your Land in his due season ... that you mayest gather in thy corn, and

A. S. Issar and N. Brown (eds.), Water, Environment and Society in Times of Climatic Change, 129–145.



Map 7.1. Location of Eretz-Israel between the Mediterranean Sea and the Syrian Desert, between the continents of Asia, Africa and Europe. The map includes the 200 mm isohyet as well as the names of the sites mentioned in this study.

thy wine. and thy oil. And I will send grass in thy fields for thy cattle, that thou mayest eat and be full" (Bible, Deuteronimium 11, 11–15).

Twelve stages of population change, with six maxima and six minima, can be discerned in historical and archaeological data in three locations in the north of Eretz-Israel during the last six millennia. The same number of stages was detected for the rainfall changes in this Land during the same time-span. The synchronization between the climatic and the settlement stages is marked. The impact of climate on Human settlement in this geographical zone is evident.

7.2. Man and Climate

Human settlement depends upon four major factors: drinking water, food supplies, security and roads. Two of these factors, water and food, are directly linked to rainfall. Several regions of the world depend upon exotic water, *i.e.* water coming from external sources. Egypt, a neighboring desert land which gets rain-water by the Nile from Equatorial East Africa, is a fine example. Other regions depend upon direct rainfall for drinking water and growing crops. Eretz-Israel is one of these regions, as detailed below.

Until modern times, Man had no facilities to store water from one year to another for agricultural irrigation. Cisterns, which were dug in several regions of this Land since some 3,000 years ago, were only used for drinking and horticulture. The whole life cycle was completely dependent upon rain. This factor could not be enhanced in order to support population increase during arid climatic stages, except for the Modern Period (Netser and Gvirtzman, 1995). The collapse of the well established Akkadian Empire, due to aridity during the Intermediate Bronze Period, is an example in a nearby region (Weiss *et al.*, 1993). On the other hand, human behavior, such as bad organization, negligence or wars, caused population decrease even during wet climatic stages, when rainwater supplies were sufficient (Broshi and Finklestein, 1992; Netser and Gvirtzman, 1996).

7.3. Global Climatic Changes

Long-term climatic changes during the Quaternary were evidenced two centuries ago by glacial and inter-glacial cycles. The changes were retrospectively predicted during the early 20th century in relation to the Earth astronomical location and orbital movement around the sun (Milankovich, 1941). The climatic stages were later proven by oxygen isotopes in deep sea cores (Emiliani, 1955, 1966). Twenty one isotopic stages were defined across the last 800,000 years (Imbrie *et al.*, 1984). Each stage lasted at least several millennia, too long to allow close correlation with human history. Within the last two decades however, much proof has been found of short-term climatic stages all over the world: lakes in Africa (Rognon, 1987), global sea level fluctuations (Pirazzoli *et al.*, 1989), an ice core in Antarctica (Jouzel *et al.*, 1989), a glacier in the Andes (Thompson *et al.*, 1987), *etc.* The short-term climatic stages may last only a few decades or centuries and so can be compared closely with human historical evolution.

Many studies of climatic stages were looking for the factors controlling the global climates, such as orbital changes (Hays *et al.*, 1976; Imbrie and Imbrie, 1980) and sun spots (Foukal, 1990; Nesme-Ribes *et al.*, 1996)). Others were looking for the Human factors being controlled by the climatic stages in the world, generally, (Rotberg and Rabb, 1981; Berger and Labeyrie, 1987) and in Israel, which this study is related to, (Issar *et al.*, 1987, 1989 and 1991; Horovitz and Weistein-Evron, 1986; Israel Acad. Sci. Hum. 1991). Numerous studies already done on climatic changes during the Holocene all around the world were recently collected and interpreted, with special reference to the Levant. The interpretation for Israel, as a part of the Levant, has been summarized to a base section for the last 10 thousand years in terms of cold and warm periods (Issar, 1995).

The short-term climatic stages in Eretz-Israel were reinterpreted in terms of rainfall from proxy-climatic data and they establish one of the bases for this study.

7.4. Climate of Eretz-Israel

Eretz-Israel is located on the border between two hemispheric climatic zones, the Mediterranean and the arid. The climate of this region is affected by the subtropical high pressure during the summer, causing warm and dry weather conditions. During the winter the region is affected by Mediterranean depressions, which bring cold weather and rainfall (Issar, 1995). The borderline between the climatic zones is taken to be the average annual isohyet of 200 mm, which passes through the Beer-Sheva Plain and divides the Land into two quite equal areas (map 7.1). This isohyet is considered as the lower limit for rain-grown crops. The southern part is the Negev, an arid land. The average annual rainfall in the Negev drops from 200 mm around Beer-Sheva to 50 mm at Elat (Shachar *et al.*, 1995). The Negev agriculture depends on floods rather than on direct rainfall. In the northern part of Eretz-Israel rain-growing crops is possible.

Both parts of the Land get winter rain. The Negev rains are mainly controlled by the Red-Sea trough, a part of the African low barometric pressure. The rains in the north of Eretz-Israel, as mentioned, are controlled by the Mediterranean and Southern Europe low pressures. The 200 mm isohyet typically shifts annually about twenty km northward or southward, thus constantly changing the border between the climatic zones (Shachar *et al.*, 1995). However, fluctuations of the Inter-Tropical Convergence Zone can move the 200 isohyet far beyond the annual shifts, thus changing the sizes of the desert on one hand and the rain-watered land on the other (Nicholson and Flohn, 1980; Magaritz and Goodfriend, 1987).

7.4.1. CLIMATIC CHANGES IN ERETZ-ISRAEL DURING THE HOLOCENE

Climatic changes in Eretz-Israel have been studied more intensively these last two decades (Horovitz and Weistein-Evron, 1986; Issar, 1995; Issar *et al.*, 1991). Sequences of instrumental climatic records in this Land exist only since the end of last century. To detect past climatic conditions, proxy-climatic factors had to be used. Three such factors were used for this study, as related to the northern part of the Land: The Dead Sea levels, palynology in a core from Lake Kinneret (Sea of Galilee) and the movement of dunes in the Coastal Plain of Israel (Netser, 1994). This study refers to climatic changes in terms of rainfall, arid and wet periods, rather than terms of temperature.

7.4.1.1. *Dead Sea Levels*

The Dead Sea is located at the south of the Jordan Valley, 400 meters below the Mediterranean Sea level. Its area is about 1,000 km^2 and its maximum depth is approximately 350 meters. The Dead Sea is a terminal lake of the Jordan River and its tributaries, which had supplied about 1,000 million cubic meters annually. This quantity was halved in the early 1960s due to water pumping from Lake Kinneret

(Sea of Galilee) by the State of Israel, and from the Yarmuk River, a tributary of the Jordan, by the Kingdom of Jordan. Several other smaller size rivers and wadis supply about 300 million cubic meters annually. The Dead Sea level depends mainly upon water supply from the Mediterranean climatic zone of the Land through the Jordan on one hand. The high evaporation rate in its valley due to high temperatures causes a two meters annual evaporation potential on the other hand. The Dead Sea level changes annually due to the annual rainfall changes in the north of Eretz-Israel. Any longer term change of level reflects a climatic change in the Mediterranean climatic zone.

The Dead Sea levels during past periods are being examined through archaeological researches and historical records, by shore terraces and alluvial fans, as well as the caves in Mount Sedom. These caves supply data for the last 8,000 years (Frumkin *et al.*, 1991). History and archaeology supply data for the last two millennia, supporting the Mount Sedom findings for that period (Klein, 1986). As the past Dead Sea levels represent past climatic conditions, mainly rainfall, the reconsideration of the Dead Sea levels data indicated twelve climatic stages during the last 6,000 years. Within this period, six wet stages and six arid ones have been identified. The climatic stages were numbered 1 to 12 from the present to the past (Netser, 1994, Fig. 1).

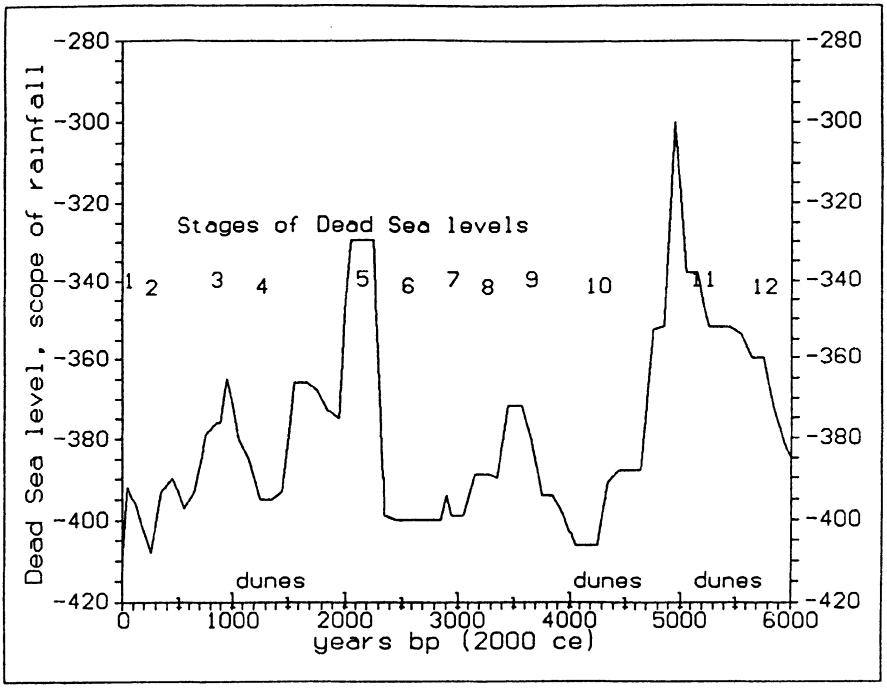


Figure 7.1. The changes of the Dead Sea levels reflect the rainfall changes in Eretz-Israel. Sand dune invasions are marked as well.

All stages are quite noticeable in the graph, except two, which need particular clarification. Stage 7 seems to be insignificant and stage 5 has two peaks. The exceptions are being explained in chapter 7.5 (discussion).

7.4.1.2. *Lake Kinneret*

Lake Kinneret (Sea of Galilee) is located in the Central Jordan Valley, in the north of Eretz-Israel. The area of the Lake is about 160 km^2 and its maximum depth is some forty meters. The Jordan River flows into the lake from the north and exits southwards to the Dead Sea. The bottom of the lake is covered by sediments, including pollen of the Mediterranean vegetation type, which arrives by wind, by the flow of the Jordan and by floods in wadis. A core (numbered KIND4), which was drilled from the sediments at the bottom of the lake, has been studied by various disciplines (Stiller *et al.*, 1986). The palynological study of the core displayed cyclic changes of the AP/NAP (arboreal plants to nonarboreal plants) ratio from the top of the core to a depth of five meters (Baruch, 1986). About twenty five samples were taken from the core for the palynological study, at an average interval of twenty centimeters. Only four samples were carbon dated. The rates of oxygen and carbon isotopes in the core sediments as well as the AP/NAP ratios were interpreted to indicate climatic changes in Eretz-Israel. A rise in the arboreal proportion represents a wet climate, while the drop is evidence of an arid one (Issar *et al.*, 1991).

The palynological data were originally presented on the depth scale and were interpreted later through interpolation into a time scale. The data of Lake Kinneret palynology were reconsidered and calculated to include the compaction rate of the sediments at the bottom of the Lake under the pressure of the water and the mud layers. This reconsideration provided a better dating interpolation (within the four measured dates) to demonstrate the climatic changes in the Mediterranean climatic zone of Eretz-Israel (Netser, 1994).

Eleven climatic stages were detected during the last 5,500 years, six wet stages and five arid ones. These stages of Lake Kinneret correlate well with the corresponding climatic stages of the Dead Sea and support them (Netser, 1994; Fig. 7.2).

7.4.1.3. *Dunes in the Coastal Plain*

The sand that today comprise dunes of the Coastal Plain of Israel was carried by the Nile from Africa to the Mediterranean Sea, transported by sea currents to the Israeli coast, carried by waves to the sea-shore and driven landwards by the winds (Nir, 1989).

The rate of dune movement into the continent depends upon shore topography, upon wind velocity consistency and upon sand humidity. Thus, during wet climates the dunes are stabilized, while during arid climates the dunes are free to move with the winds. Three phases of dune development during the last 6,000 years were detected. The oldest occurred about 5,400 years ago and the second about 4,600 years ago, while the latest started about 1,300 years ago (Issar *et al.*, 1991; Netser, 1994) and continued intermittently until recently. The dating of the dunes was decided according to archaeological findings in the Coastal Plain, which were covered by the dunes, and Carbon dating of land-snails buried in the sand. The last phase, for example, covered Byzantine farms, dated 5–6th centuries AD, as well as Eucalyptus trees planted by Jewish settlers some 100 years ago (photographs 1 and 2). These dune movements are well correlated with stages 12, 10 and 4, the arid ones of the Dead Sea and Lake Kinneret. Dune stabilities about 5,200 and 3,800 years BP are correlated with wet stages 11 and 9, respectively (Netser, 1994, Fig. 2).

7.4.1.4. *Climatic Stages in Eretz-Israel*

From data from the Dead Sea, Lake Kinneret and the Coastal Plain it can be inferred that twelve climatic stages occurred in Eretz-Israel during the last 6000 years, six arid and six wet. The arid stages occurred around 5,600, 4,200, 3,400, 2,600, 1,300 and 400 years ago, and the wet ones around 5,000, 3,800, 3,000, 2,000, 900 and 100 years before the present. Note that all dating in this study is by BP (before present; the present for this study is 2000 CE) years as used in paleoclimate studies, rather than BC or AD (CE), as used in historical and archaeological accounts.

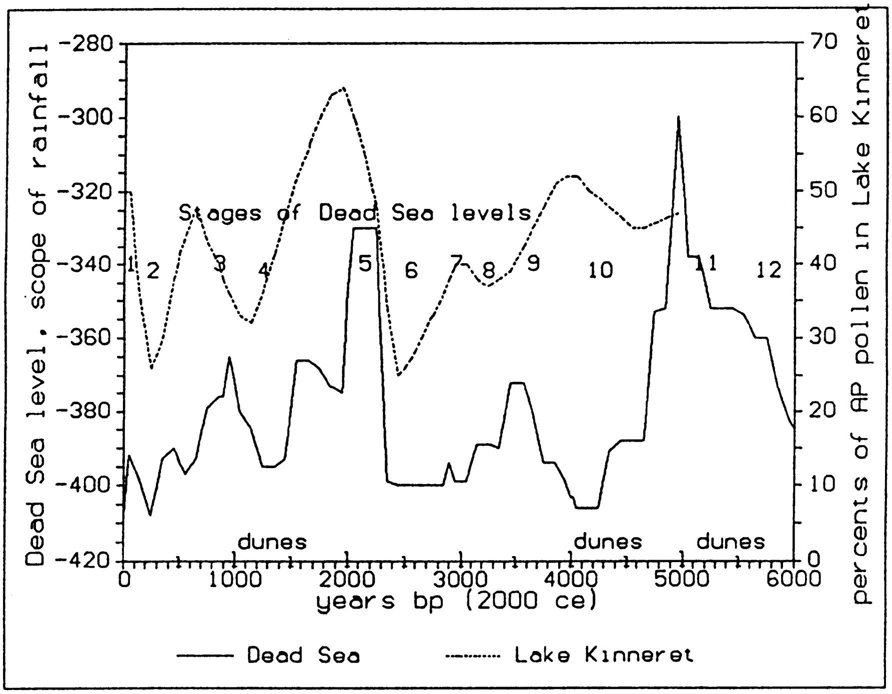
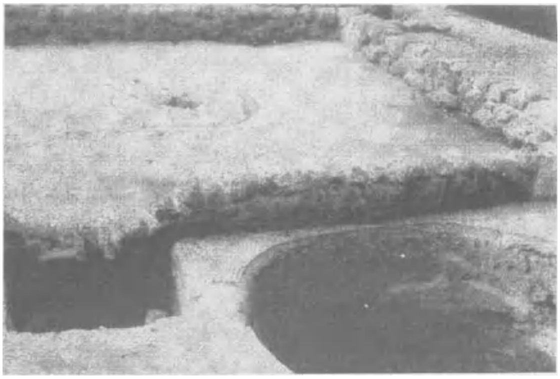


Figure 7.2. The changes of the AP/NAP ratio in Lake Kinneret core compared with the changes of the Dead Sea levels. Sand dune invasions are marked as well.

Photograph 7.1. Sand dune Rishon Le-Ziyyon (the Coastal Plain) covered trees which were planted some 100 years ago in order to stop the dune movement. The burial of the trees indicates dune movement during the last century.



Photograph 7.2. A Byzantine winepress was buried by the invading dunes in the Coastal Plain of Israel and was discovered recently in Rishon Le-Ziyyon (same location as Photo 7.1).
This ancient vine-press indicates the earliest dating of the last dune invasion. (Both photos: M. Netser, 1991.)

The climatic stages as compared to the historical periods demonstrate the following, (from the past to the present): stage 12 is the Late Chalcolithic Period; 11–Early Bronze 1–2; 10–Early Bronze 3 and Intermediate Bronze; 9–Middle Bronze; Byzantine; 4–Arabic; 3–Crusaders and Mamlukian; 2–Turkish-Ottoman; and 1 is the Modern Period. The term 'Period' in this study is used both for cultural ages (Bronze Age, Iron Age) and historical periods (Israelite Period, Roman Period, *etc.*).
The historical timetable of Eretz Israel appears below for reference (Table 7.1).

7.5. Settlement in Eretz-Israel

The settlement in Eretz-Israel, for the purpose of this study, was examined in three regions in the northern part of the Land: Gush-Dan (Region of Dan), Eretz Binyamin (Land of Benjamin) and the Hill Country of Menashe, all named after the Israelite tribes who settled there 3,200 years ago. The restoration of the number of settlements and their spatial dispersion in the past was made by reprocessing of archaeological and historical data, as will be detailed regionally later. Several stages of settlement were found in each region (§7.5.1–3). The outstanding stages of demographic

Table 7.1. Historical periods in Eretz-Israel (after: Stern, 1992; Gophna *et al.* 1988). Yerrs BP are the years before 2000 CE

Years BP	Period	Years BP	Period
Present		2,587	
	Modern		Israelite (Iron) 3
120		2,722	
	Ottoman		Israelite (Iron) 2
484		3,000	
	Mamlukian		Israelite (Iron) 1
708		3,200	
	Crusaders		Late Bronze
900		3,550	
	Arabic		Middle Bronze 2
1,364		4,000	
	Byzantine		Middle Bronze 1 (Intermediate Br.)
1,670		4,200	
	Roman 3		Late Bronze 3
1,867		4,650	
	Roman 2		Late Bronze 2
1,930		5,050	
	Roman 1		Late Bronze 1
2,063		5,300	
	Hellenistic		Chalcolithic
2,332		7,000	
	Persian		
2,587			

rise were, for example, during the settling of the People of Israel (the First Temple), the Roman Period (the Second Temple) and the Byzantine period. Stages of decline occurred, for example, at the end of Early Bronze, at the destruction of the Temples and during the Arabic Period. Several previous studies claim that the rising number of population and settlements occurred during relative peace and organized economy (Broshi and Finklestein, 1992).

Several archaeological methods are used for the estimation of the population size at any location within a given period: 1. The estimated population density per dunam (1,000 m^2.), or hectare (100 dunams), allowing for the differences between urban and rural settlements, 2. The dwelling based estimation, including the number of the rooms at a site and the average family size. 3. The necessary natural resources for the existence of people (Zorn, 1994). One source of inaccuracy is that not all ancient sites were found in archaeological surveys, or through random findings. Another is that the population density varied from one period to another. Calculations of this type were made for all regions in Eretz-Israel for several historical periods, such as the Early Bronze (Broshi and Gophna, 1984a; Finklestein and Gophna, 1993), the Middle Bronze 2 (Broshi and Gophna, 1984b), the Iron (Israelite) 2 Period (Broshi and Finklestein, 1990 and 1992) and the Roman-Byzantine (Broshi, 1979) Periods. All these studies were focused on periods of population growth. A particular calculation was made for one region, the Hills of Menashe, for three successive periods: the Middle Bronze 2, the Late Bronze and the Israelite Periods. The results were 30,000 inhabitants in 116 settlements during the Middle Bronze 2; 7,000 inhabitants in 31 settlements during the Late Bronze; and 27,000 inhabitants in 96 settlements during the Israelite 1 Period (Zertal, 1996). So, here there is a close correlation between the number of the settlements and the size of the total population.

However, none of the above mentioned methods is applicable for the three studied regions. So, for the purposes of this study, the number of settlements during each historical period is used, rather than the size of the population, as the data for calculating the population is insufficient for most regions and periods. This method is being supported by the close relations which were found between the number of settlements, the size of the settled area and the calculated total population during 2,400 years, from the Chalcolithic to the Middle Bronze Periods (Gophna and Portugali, 1988). It is mandatory to consider that the rise or the decline of settlements are relative for each period between its two adjacent ones. Thus, the population in a certain rise period may be lower than the minimum of another one several millennia later. Such an example exists for the rise during the Early Bronze Period

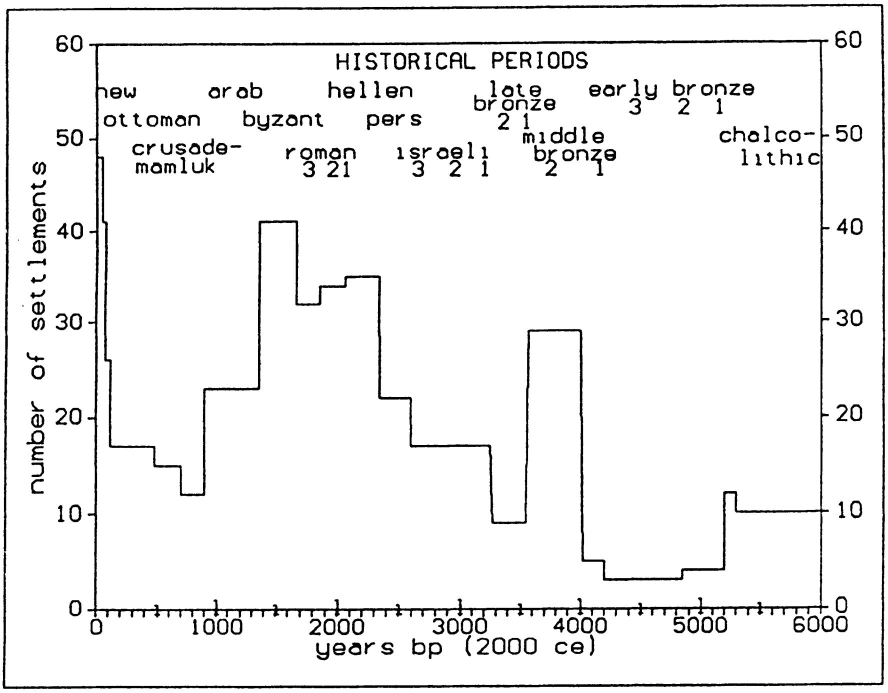


Figure 7.3. The settlement changes in Gush-Dan during the last 6,000 years.

in Gush-Dan, which is lower than the decline during the Arabic Period in the same region (Fig. 7.3).

There is a continual rise of the maxima, for each region, from the past to the present. This rise is related to the development of human culture and technology. The rise of settlements at the present Modern Period is entirely a result of human behavior.

Due to the different methods of calculating population and the different methods of archaeological surveys, the term 'settlement' in this study will refer to a site which was indicated to be settled by man at any given period, either a town or a village, a farm, a hamlet or a cemetery.

7.5.1. GUSH-DAN

Gush-Dan is located at the center of the Coastal Plain of Israel, including the town of Tel-Aviv and its surroundings, a total area of 350 km^2. Gush-Dan is the most populated region in Israel, now. The modern development of this region has been started at the beginning of this century. Due to the fast construction of the region no thorough archaeological survey was made there, except for a few major sites (Yafo, Qasila, Afeq, Azor), and most archaeological findings were random through development activity. Nevertheless, some 180 sites were found in Gush-Dan, mainly during the last 50 years, and a few additional ones are found annually. The archaeological data of Gush-Dan were geographically reworked into 90 settlements defined by their historical periods (Netser, 1994). In Gush-Dan, as in the other regions, very few locations were settled along all historical periods; most locations were inhabited during some periods only.

A growing number of settlements in Gush-Dan occurred during the Early Bronze 1, Middle Bronze 2, Israelite, Roman-Byzantine, Mamlukian and the Modern Periods. The outstanding reductions of settlements happened at the end of the Chalcolithic and during the Early Bronze 2–3, Intermediate Bronze, Late Bronze, and the Arabic Periods (Fig. 7.3).

7.5.2. THE HILL COUNTRY OF MENASHE

The Hills of Menashe are located at the northern part of the Shomeron (Samaria), as a part of the Central Mountain of Eretz-Israel. A thorough archaeological survey was carried out at the western half of this region on an area of about 600 km^2 (Zertal, 1986). All the finds in this survey were organized into 164 settlements, defined by their historical periods (Fig. 7.4). In addition to archaeology, the settlements in the Hills of Menashe were described geographically as related to topography, geology and water resources. The data state that the decline of the settlements during the Late Bronze Period was synchronous with the drying of springs in the region (Zertal, 1986).

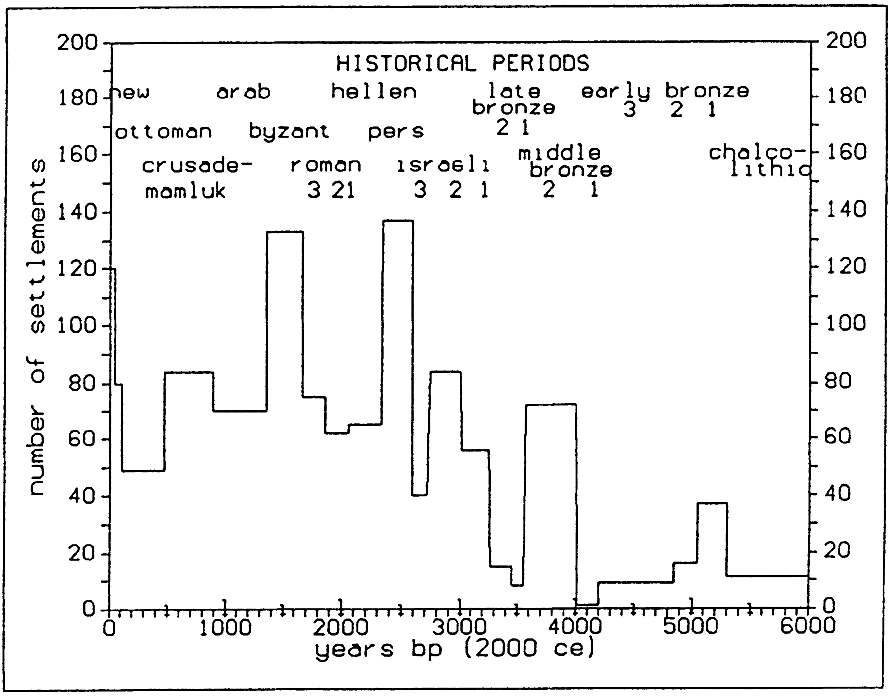


Figure 7.4. The settlement changes in the Hills of Menashe during the last 6,000 years.

The growing number of settlements in the Hills of Menashe occurred during the Early Bronze 1, Middle Bronze 2, Israelite 1–2, Persian, Byzantine, Crusader-Mamlukian and Modern Periods. Outstanding declines of settlements happened at the end of the Chalcolithic and during the Early Bronze 2–3, Intermediate Bronze, Late Bronze, Israelite 3, Roman, Arabic and the Turkish-Ottoman Periods (Netser and Gvirtzman, 1996, Fig. 4).

7.5.3. ERETZ BINYAMIN

Eretz Binyamin (the Land of Benjamin) is located north of Jerusalem, at the northern part of the Mount of Yehuda (Judea), an area of about 500 km^2. An archaeological survey of this region was recently accomplished (Magen and Finkelstein, 1993). The summary of the survey presented the archaeological sites by their historical periods, but did not define the sites into settlements. Studies of other regions demonstrate that the number of settlements is relative to the total number of archaeological sites. Thus, for this region, the number of sites, instead of settlements, was used in this study.

The growing number of settlements in Eretz Binyamin occurred during the Middle Bronze 2, Israelite 1–2, Persian, Hellenistic, Roman, Byzantine and the Modern Periods. Outstanding declines of settlements happened since the end of

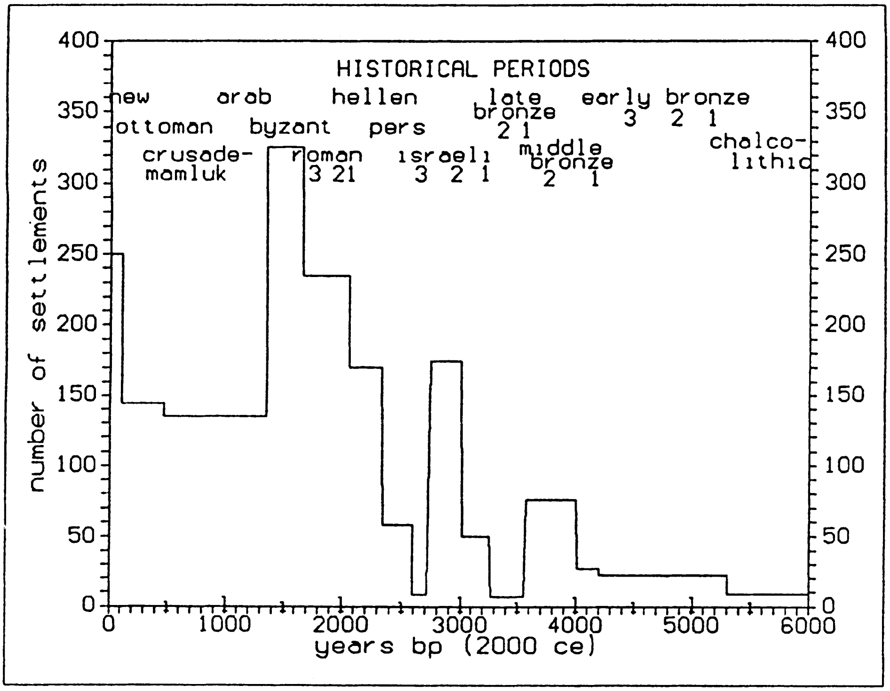


Figure 7.5. The settlement changes in the Land of Binyamin during the last 6,000 years.

the Chalcolithic until the Early Bronze 2–3, Intermediate Bronze, Late Bronze, Israelite 3 and the Arabic Periods (Fig. 7.5).

7.6. Discussion

As seen in the above figures there were twelve stages of settlement in the three studied regions during the last 6,000 years, six stages of settlement boom and six stages of decay. Twelve stages of the Dead Sea level were listed. In general, the synchronization between the rainfall quantities and the number of settlements is very clear along the history of Mankind in this Land (Fig. 7.6). The growing of settlements occurred during the Early Bronze 1, climatic stage 11, Middle Bronze 2, stage 9, Israelite 1–2, stage 7, Roman-Byzantine, stage 5, Crusaders, stage 3, and the Modern Periods, climatic stage 1. The settlements declined during the end of the Chalcolithic, stage 12, Early Bronze 2–3, stage 10, Late Bronze, stage 8, Israelite 3, stage 6, Arabic, stage 4, and the Turkish-Ottoman Periods, climatic stage 2.

In this correlation there are several exceptions. Several exceptions are explained as the results of human behavior. The rise of the settlement in Gush-Dan during the Israelite Period (§7.5.1) is not as outstanding as those in the other two studied regions (§7.5.2–3). This fact is explained by its marginal location at that period and

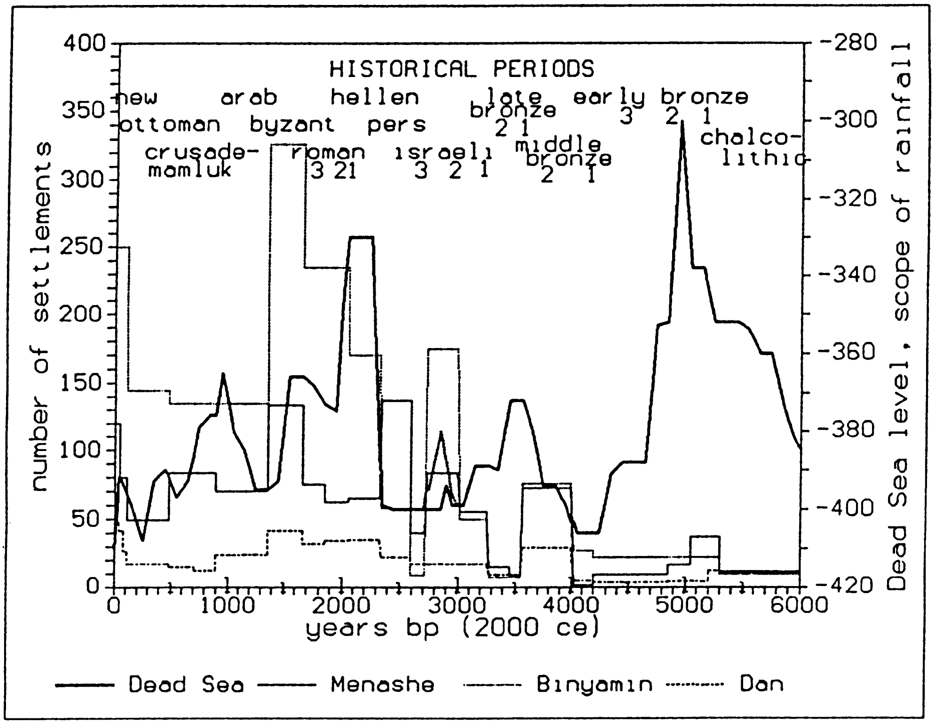


Figure 7.6. Comparison between the rainfall changes in the Mediterranean climatic zone in Eretz-Israel and the settlement changes in Gush-Dan, the Hills of Menashe and the Land of Binyamin during the last 6,000 years. A suggestion for the reinterpretation of climatic stage 7 is shown.

by the constant struggle between the Israelite Tribes and the Philistines in this region. The large drop of the settlements in the Menashe region during the Roman Period (§7.5.2), as compared to the small drop in Gush Dan and to the heavily populated Binyamin at the same period, is explained through the Samaritan rebellion against the Roman Empire.

Other exceptions apply to the climatic stages. During the Israelite Period, the Dead Sea level at stage 7 appears to be low, which seems to contradict the general concept. This is explained by the lack of data from the Dead Sea for the period between 3,000 and 2,000 years BP (Fig. 7.1). But, the data from Lake Kinneret present an increase of the AP pollen percentage during the same time span (Fig. 7.2) to support the idea of a higher Dead Sea level during the third millennium BP, as a result of the rainfall increase. A correction to stage 7 is suggested in Figure 7.6.

Climatic stage 5 is double-peaked. This stage 5 is considered as one even though it has two sub-stages, since the intermediate drop was to the level of –380 meters, which is a high stand of the Sea. The level drop is too small to justify a division into two separate stages.

7.7. Conclusions

The synchronization between the climatic stages and the settlement trends in the north of Eretz-Israel during the last six millennia, as studied in three adjacent regions, is remarkable. Each one of the twelve settlement stages is well fitted into a climatic stage, except for a few explicable exceptions. This synchronization emphasizes the impact of climate on human settlement in this Land. Human life depends upon water. Rain is the only source of water for drinking and for food supplies, either field crops and gardens or pastures. People could exploit the gift of Heaven according to their relevant culture and technological development. In ancient times, Man could not increase, by any means, the available amount of water for the growing of staple food and, thus, the population size in this part of the globe was fully dependent upon climatic conditions, rainfall mainly.

References

Baruch, U. (1986) The Late Holocene vegetational history of Lake Kinneret (Sea of Galilee), Israel. *Paleorient* 12(2), 37–48.

Berger, W.H. and Labeyrie, L.D., eds. (1987) *Abrupt Climatic Changes, Evidence and Implications.* Reidel Publishing Co., Dordrecht.

Bible, Deuteronimium 11.

Broshi, M. (1979) The population of Western Palestine in the Roman-Byzantine Period. *Bulletin of the American School of Oriental Researches* 236, 1–10.

Broshi, M. and Finklestein, I. (1990) The number of the population in Eretz-Israel at the year 734 BC. *Qatedra* 58, 3–24 (in Hebrew).

Broshi, M. and Finklestein, I. (1992) The population of Palestine in Iron Age 2. *Bulletin of the American School of Oriental Researches* 287, 47–60.

Broshi, M. and Gophna, R. (1984a) The settlement of Eretz-Israel and its population during the early Bronze Period. *Eretz-Israel* 17, 147–157 (in Hebrew).

——. (1984b) Eretz-Israel during the Middle Bronze 2 Period: Settlements and Population. *Qatedra* 31, 3–26 (in Hebrew).

Emiliani, C. (1955) Pleistocene temperature variations in the Mediterranean. *Quatenaria* 2, 87–93.

——. (1966) Isotopic Temperatures. *Science* 154, 851–857.

Finklestein, I. and Gophna, R. (1993) Settlement, demographic, and economic patterns in the Highlands of Palestine in the Chalcolithic and Early Bronze and the beginning of Urbanism. *Bulletin of the American School of Oriental Researches* 289, 1–22.

Foukal, P.V. (1990) The variable sun. *Scientific American* 262, 26–33.

Frumkin, A., Magaritz, M., Carmi, I., and Zak, I. (1991) The Holocene climatic record of the salt caves of Mount Sedom, Israel. *The Holocene* 1, 191–200.

Gophna, R. and Portugali, J. (1988) Settlement and demographic processes in Israel's Coastal Plain from the Chalcolithic to the Middle Bronze Age. *Bulletin of the American School of Oriental Researches* 269, 11–28.

Hays, J.D., Imbrie, J. and Shackleton, N.J. (1976) Variations in the Earth's orbit: A pacemaker of the Ice Ages. *Science* 194, 1121–1132.

Horovitz, A. and Weistein-Evron, M. (1986) The Late Pleistocene Climate of Israel. *Bull. Assoc. Française Étude Quaternaire* 1–2, 84–90.

Imbrie, J. and Imbrie Z.J. (1980) Modeling the climate response to orbital variations. *Science* **207**, 943–953.

Imbrie, J., Hays, J.D., Martinson, D.G., McIntyre, A., Mix, A.C., Morley, J.J., Pisias, N.G., Prell, W.L., and Shackleton, N.J. (1984) The orbital theory of Pleistocene climate: Support from a revised Chronology of the marine (180 record). *In* A. Berger, J. Imbrie., J. Hays, G. Kukla and B. Salzman (eds.), *Milankovitch and Climate, part 1*, Reidel Publ. Co., Dordrecht, pp. 269–305.

Israel Academy of Science and Humanities (1991) *Abstracts.* International Workshop on Regional Implications of Future Climate Change, Weizman Institute of Science, Rehovot, Israel.

Issar, A.S. (1995) *Impacts of Climate Variations on Water Management and Related Socio-Economic Systems.* UNESCO, Paris.

Issar, A. and Tsoar, H. (1987) Who is to blame for the desertification of the Negev, Israel? *Proceedings, IAHS Symposium*, Vancouver, Canada, IAHS Publication no. 168, pp. 577–583.

Issar, A., Tsoar, H. and Levin, D., (1989) Climatic changes in Israel during historical times and their impact on hydrological, pedological and socio-economic systems. *In* M. Lenen and M. Sarnstein (eds.), *Paloeclimatology and PaleoMeteorology: Modern and Past Patterns of Global Atmospheric Transport*, Kluwer Academic Publishing, Dordrecht, pp. 525–541.

Issar, A.S., Govrin, Y., Geyh, M.H., Wakshal, E., and Wolf, M., (1991) Climate changes during the Upper Holocene in Israel. *Israel Journal of Earth Sciences* **40**, 219–223.

Jouzel, J., Barkov, N.I., Barnola, J.H., Genton, C., Korotkevitz, Y.S., Kotlyakov, V.M, Kegrand, M., Lorius, C. Petit, J.P., Petrov, V.N., Raysbeck, G., Ritz, C., and Yiou, F. (1989) Global change over the last climatic cycle from the Vostok Ice Core Record (Antarctica). *Quaternary International* **2**, 15–24.

Klein, Z. (1986) The Dead Sea Level Changes and Climatic Fluctuations in Eretz-Israel During the Historical Periods. Ph.D. thesis, The Hebrew University, Jerusalem (in Hebrew with English abstract).

Magaritz, M., and Goodfriend, G.A. (1987) Movement of the desert boundary in the Levant from Later Pleistocene to Early Holocene. *In* W.H. Berger and L.D. Labeyrie, (eds.), *Abrupt Climatic Changes, Evidence and Implications*, Reidel Publishing Co., Dordrecht, pp. 173–183.

Magen, Y. and Finkelstein, I., eds. (1993) *Archaeological Survey of the Hill Country of Benjamin*, Israel Antiquities Authority and the Civil Administration of Judea and Samaria, Jerusalem, Israel (in Hebrew with English abstract).

Milankovitch, M. (1941) *Kanon der Erdbestrahlung. Koning. Serb. Akadem. Geogr. Special Publication* 132, (English translation by the Israel program for scientific translations: Canon of Insulation and the Ice Age Problem, Jerusalem, 1969).

Nesme-Ribes, E., Baliunas, S.L. and Sokoloff, D. (1996) The stellar dynamo. *Scientific American* **275**, 30–36.

Netser, M. (1994) The Climatic Changes During the Holocene Stage and their Effect on the Formation of the Landscape in Gush-Dan (Israel) and on the Human Settlement in this Region. Ph.D. thesis, Bar-Ilan University, Ramat-Gan, Israel (in Hebrew with English abstract).

Netser, M. and Gvirtzman, G. (1995) Climatic changes in Israel during the last 6,000 years and their implications to the future. *Abstracts, INQUA XIV Congress,* Berlin, p. 199.

——. (1996) Settlement changes in Menashe Mountains (Shechem syncline) in scope of climatic changes in Eretz-Israel during the last 6,000 years. *In* Y. Eshel (ed.), *Judea and Samaria Research Studies*, College of Judea and Samaria, Kedumim-Ariel, pp. 293–299 (in Hebrew, with English abstract).

Nicholson, S.E. and Flohn, H. (1980) African environmental and climatic changes and the general circulation in the Late Pleistocene and Holocene. *Climatic Change* **2**, 315–348.

Nir, Y. (1989) Sedimentological aspects of the Israel and Sinai Mediterranean coasts. Report GSI/39/88, Geological Survey of Israel, Jerusalem (in Hebrew).

Pirazzoli, P.A., Grant, D.R. and Woodworth, P. (1989) Trends of relative sea-level change: Past, present and future. *Quaternary International* 2, 63–71.

Rognon, P. (1987) Aridification and abrupt climatic events on the Saharan northern and southern margins, 20,000 years to present. *In* W.H. Berger and L.D. Labeyrie (eds.), *Abrupt Climatic Change*, Reidel Publ. Co., Dordrecht, pp. 209–220.

Rotberg R.I. and Rabb, T.K., eds. (1981) *Climate and History*. Princeton University Press, N.J.

Shachar, A., Adler, R., Amiran, D., Zameret, R., Heinberger, I., Sofer, A., Kadmon, N., Nir, D., Sofer, T., Kantor, M., Rapp, E., Ron, A., and Strug, Z., eds. (1995) *The Atlas of Israel*. Tel-Aviv, Israel.

Stern, E., ed. (1992) *The New Encyclopaedia of Archaeological Excavations in The Holy Land*. 3 vols., Israel Exploration Society, Jerusalem (in Hebrew).

Stiller, M., Ehrlich, A., Pullinsher, U., Baruch, U. and Kaufman, A. (1984) The Late Holocene sediments of Lake Kinneret (Israel)—Multidisciplinary study of a % meter core. *Geological Survey of Israel Current Research, 1983–84,* pp. 53–68.

Thompson, L.G. and Mosley-Thompson, E. (1987) Evidence of abrupt climatic change during the last 1,500 years recorded in ice cores from Tropical Quelccaya Ice Cap, Peru. *In* W.H. Berger and L.D. Labeyrie (eds.), *Abrupt Climatic Change*, Reidel Publ. Co., Dordrecht, pp. 99–110.

Weiss, N., Courty, M.A., Wetterstrom, W., Guichard, F., Senior, L., Meadow, R., and Curnow, A. (1993) The genesis and collapse of Third Millennium North Mesopotamian Civilization. *Science* 262, 995–1004.

Zertal, A. (1986) The Israelite Settlement in the Hill-Country of Menashe. Ph.D. thesis, Tel-Aviv University, Israel (in Hebrew).

Zorn, J.R. (1994) Estimating the population size of ancient settlements: Methods, problems, solutions and a case study. *Bulletin of the American School of Oriental Researches* 295, 31–48.

Chapter 8

Settlement, Agriculture and Paleoclimate in ᶜUvda Valley, Southern Negev Desert, 6th–3rd Millennia BC

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Figure 8.1. ᶜUvda Valley before construction of the air base, from west. In the background are the Edomite Mountains of Jordan.

8.1. Introduction

ᶜUvda Valley, in the southern Negev, Israel, is an extreme desert area. It is characterized by summer temperatures above 40°C, low precipitation of 28 mm annual average, with an annual evaporation rate of 4,000 mm. The significance of these figures is more apparent when compared with those of the Negev Highland desert, 100 km to the north, with 100 mm average annual precipitation and 2,000 mm evaporation potential. Obviously, the water balance of the southern Negev is very negative. Since vegetation is restricted to wadi beds, the carrying capacity of the

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area is low for vegetation, for animals, and for man who subsists on both. Accordingly, only limited archaeological remains would be expected, *i.e.* the markers for human presence and activity in the past. Indeed, most of the southern Negev is characterized by a low density of ancient sites. Yet, contrary to expectations, archaeological remains in the southern-most Negev are exceptionally abundant. From ʿUvda Valley in the north to Eilat in the south, some 1,200 km^2, 1,600 sites are known today, even though only 16% of the area has been surveyed in detail.

ʿUvda Valley, located 40 km north of Eilat and 10 km west of the Arabah Valley, is 15 km long, up to 5 km across, and is surrounded by ridges 50–150 m high. The eastern ridges are built of the Turonian 'Grofit' formation of limestone, while the western side is comprised of the Senonian 'Sayarim' formation of chalk and flint. The valley itself is filled with Pleistocene and Holocene alluvium, at least 52 m deep, made up of silt, clay, lime-sand and fine gravel (Ginat & Zilberman, 1992). The drainage system of the valley covers an area of 400 km^2, mainly in the mountainous area to the south, close to 900 m above sea level. The gradient of the valley is moderate, from 500 m above sea level in the south to 415 m in the north. The gradient in the south is 1%, while in the north it is only 0.3%. Since the eastern side is lower than the western by 20 m, it is better irrigated by flood water.

ʿUvda Valley (*Wadi ʿUqfi* in Arabic) was first briefly described by A. Musil (1907:180–182, 1926:85) and by F. Frank (1934:263–265). Both recognized ancient remains, while Musil noted that the eastern side was cultivated by the Haiwat Bedouin, and citizens of Aqaba, who rented plots from them. The first archaeological survey was made by Rothenberg (1967a:138, 1967b:303–307) who documented 15 sites while citing more sites distinguished from the air.

In December 1987, I began a renewed survey of the area, leading a team which was part of the Negev Emergency Survey, under the auspices of the Israel Department of Antiquities and the Israel Archaeological Survey (today the Israel Antiquities Authority). The emergency survey was intended to precede the redeployment of the Israel army from Sinai, while ʿUvda Valley itself was selected for a new air base. The survey lasted several years but was never completed. The western side of the valley was only briefly surveyed and revealed a small number of sites. However, one third of the area on the eastern side was meticulously surveyed, resulting in the documentation of approximately 400 sites in an area of 50 km^2 (Fig. 8.2). This site density was unexpectedly high considering the present environment (Avner, 1979, 1982a,b, 1990b, 1993). The remains in the valley present a complete sequence of settlement from the Pre-Pottery Neolithic B (PPNB) to the present, 10,000 years. Most sites are dated to the 6th–3rd millennia BC, *i.e.* the Pottery Neolithic, Chalcolithic and the Early Bronze Age. Of these, 186 are stone built habitation sites, the others include tent camps, agricultural installations, and cult and burial locations.

During June 1979, Early Bronze habitation sites were excavated for the first time by Amiran, Arnon and Avner (1980). In February 1980, an excavation operation took place, directed by Eitan and Cohen, in which 22 sites were excavated by 20

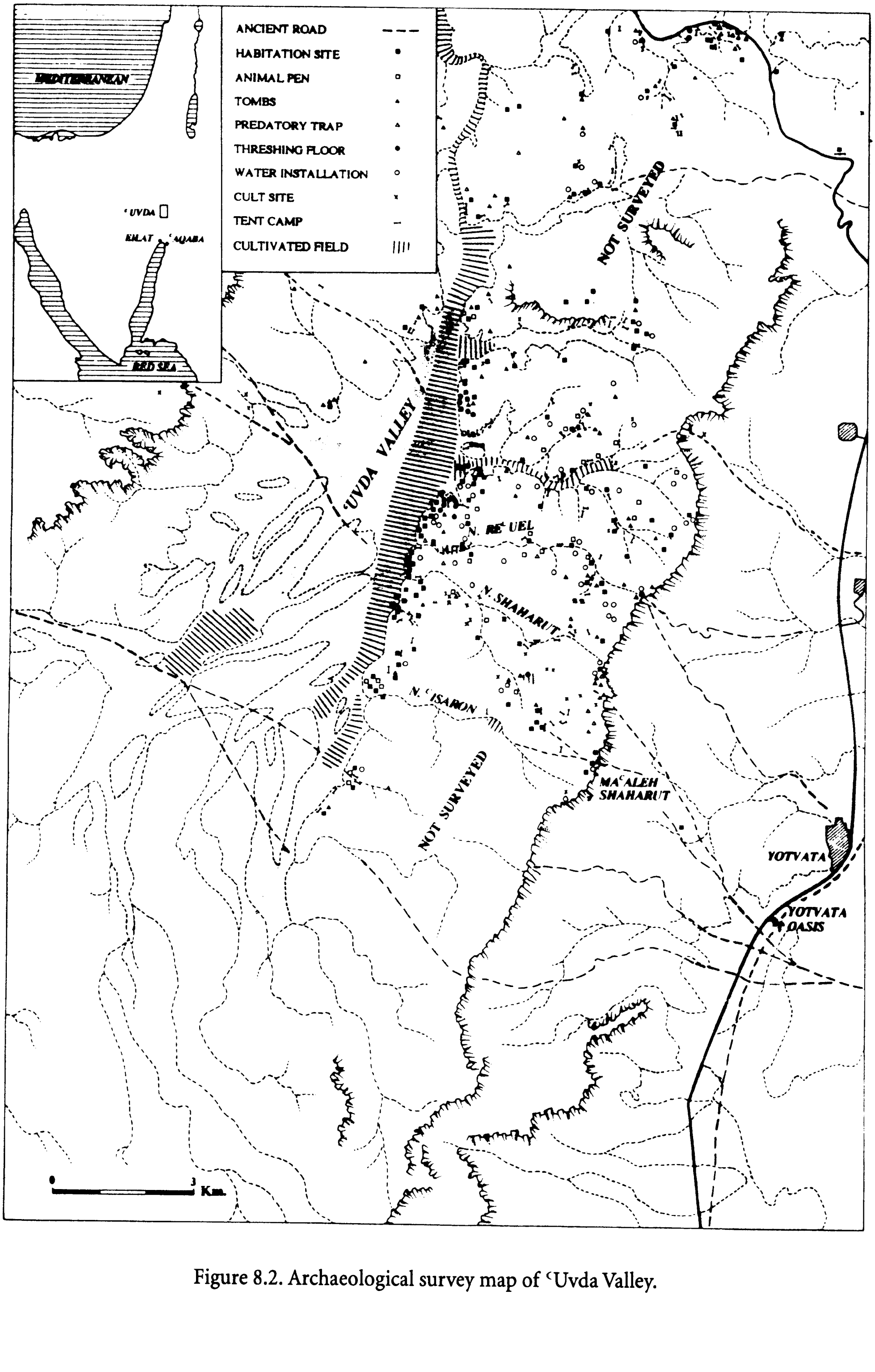

Figure 8.2. Archaeological survey map of ʿUvda Valley.

archaeologists and 180 volunteers (*Hadashot Archaeologiot* 64–65, 1980:35–49). Later, excavations and preservation of additional sites were undertaken (Avner, 1982b, 1982d, 1983, 1986, 1989a).

Initial plans for the air base required destruction of 104 sites. However, after long negotiations, plans were re-adjusted so that all but one site remained outside the base perimeter. The U.S. Army Corp. of Engineers constructing the base also displayed a high sensitivity to the ancient sites and avoided damage.[1] Today, the sites are accessible and an archaeological park is in an advanced stage of planning.

8.2. Dwelling Sites and the Settlement Scenario

8.2.1. NEOLITHIC SITES

Seven sites of the PPNB period (8th–7th millennia BC)[2] were discovered on the eastern margin of the valley and in the wadis to the east. Three are stone-built dwelling sites, the rest, camping or flint industry sites. Two were excavated: the site in Nahal Reʿuel only partially (Ronen, 1980), and the one in the mouth of Nahal ʿIssaron, almost completely (Goring Morris & Gopher, 1983; Goring Morris, 1993; Gopher *et al.*, 1995). The site of Nahal ʿIssaron covers an area of approximately 500 m^2 and consists of circular spaces, 2–3.5 m in diameter, with almost no open spaces (Fig. 8.3). It is similar to contemporary habitation sites excavated in the southern Sinai (*e.g.* Bar Yosef, 1981b, 1984), though somewhat larger. Arrowheads are the most prominent find (Fig. 8.4) comprising 20% of the flint assemblage and indicating the importance of hunting to the economy of the inhabitants. Correspondingly, the animal bones collected include ibex, gazelle, wild ass, wild cattle, hare, several bird species, fish and ostrich egg shells. No remains of domesticated animals were found. This was similar to the situation in Southern Sinai sites,[3] and contrary to contemporary sites in the Edomite Mountains, Beidha and Basta, *ca.* 70 km northeast of ʿUvda Valley (Becker, in Nissen *et al.*, 1991:29–32). Many grinding stones were also found, as well as possible granaries (Fig. 8.5), indicating utilization of cereals for preparing porridge and baking bread. Again, no sign of plant domestication has been discovered. The architectural remains and artifacts imply the presence of one extended family in the site, from autumn to spring, returning to the site repeatedly. The PPNB stage (Stratum C) was dated by nearly 20 radiometric dates beginning in the early 8th millennium BC, and ending in the mid-7th millennium BC (Table 8.1).

The next stage in N. ʿIssaron (Stratum B) was dated to the Pottery Neolithic (PN), first on the basis of arrowhead types (Gopher, 1994:118–119, Gopher *et al.*, 1995, *cf.* Fig. 8.6) and later by 15 radiometric analyses, spanning the early 6th–early 4th millennium BC (Table 8.1). Here a new aspect appeared, an open space surrounded by rooms, heralding an important architectural innovation in desert sites.[4]

Figure 8.3. Aerial photograph of N. ʿIssaron site, from west, on the left side are remains of Stratum A of the 4th–3rd millennia BC.

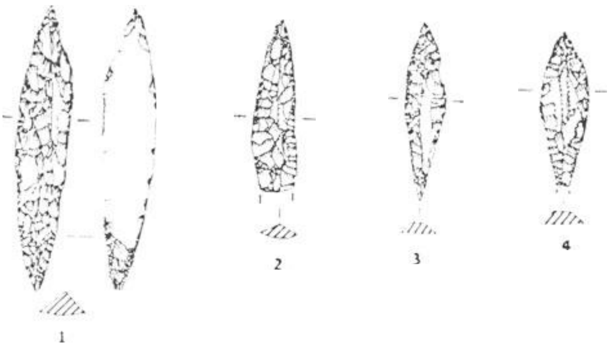

Figure 8.4. PPNB *ʿAmuq* arrowheads from N. ʿIssaron site (surface collection) scale 1:2.

Figure 8.5. N. ʿIssaron site, a granary or oven, with fragments of grinding stones on the top left (excavated by Goring-Morris & Gopher).

Courtyards or open spaces are absent in the PPNB sites of the Negev and Sinai. However, they are characteristic of many hundreds of 5th–3rd millennia BC sites in these regions, as well as in other desert areas of the Near East. In ʿUvda Valley site 9, a level of goat dung mixed with dirt was dated to the 6th–5th millennia BC based on small winged and transversal arrowheads, and the radiometric date of an adjacent cult installation (Table 8.2, #8). In two additional sites, courtyards were dated to the 2nd half of the 6th and 5 millennia BC (Table 8.2, #15,17). Accordingly, it seems that stone-built courtyards first appeared in the 6th millennium BC, and indeed signify the emergence of herding in the Negev and Sinai. This innovation had an enormous influence on the desert population, which will be addressed below.[5]

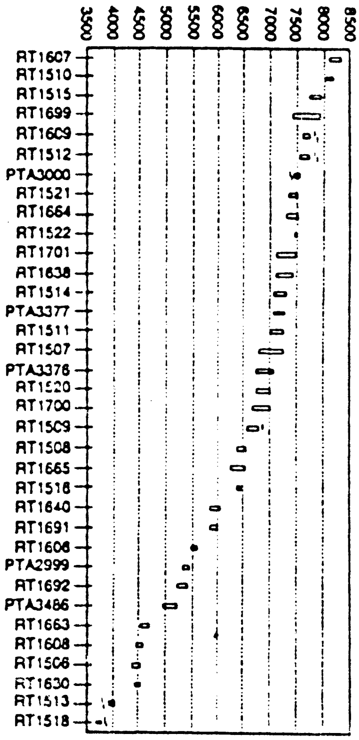

Table 8.1. Calibrated BC ^{14}C histogram of N. ʿIssaron site (after Carmi *et al.* 1994, Fig. 3).

8.2.2. CHALCOLITHIC AND EARLY BRONZE SITES

One hundred and eighty-six stone built sites, and 40 tent camps, were documented from the 5th–3rd millennia BC. Tent camp remains appear as rows of cleared circles 3–5 m in diameter, with medium and large size stones scattered around them (Fig. 8.7), originally used to secure the tent ropes. The camps, 50–200 m long, each contain 8–25 tent bases. Approximately 2/3 of the sites were in locations shielded from north or west winds, meaning they were suitable for winter occupation. About 1/3 were exposed to the north wind, and therefore best suited for the summer. Winter

Figure 8.6. Late Neolithic arrowheads from ʿIssaron site: #1 – 'Nizzanim' point; #2,3 – 'Haparsa' points; #4 – Transversal arrowhead, scale 1:1.5 (after Goring-Morris & Gopher, 1993, Fig. 4).

Table 8.2. ^{14}C dating of ʿUvda Valley sites

No.	Lab No.	Site	BP	Cal. BC	Prob. %	Ref
1	Pta 3137	N. Recuel	8720±70	7891–7612	100	1, 5
2	Pta 2848	N. Recuel	8670±50	7854–7575	100	1, 5
3	Pta 3202	N. Recuel	8650±90	7873–7542	100	1, 5
4	Pta 3000	N. ʿIssaron(C)*	8430±80	7539–7325	100	2, 5
5	Rt 1509	N. ʿIssaron(C)	7870±55	6765–6562	93	6, 7
6	Rt 1640	N. ʿIssaron(C)	7135±95	6046–5854	99	6, 7
7	Rt 670D	ʿUvda 9	7960±200	7048–6549	97	5
8	Pta 3646	ʿUvda 9	6960±70	5858–5714	96	5
9	Rt 628A	ʿUvda 6	6560±200	5611–5247	100	3, 5
10	Rt 628B	ʿUvda 6	6400±70	5422–5273	100	3, 5
11	Pta 3621	ʿUvda 6	6400±60	5420–5275	100	3, 5
12	Rt 1739	ʿUvda 6	6390±60	5412–5264	100	7
13	Pta 2999	N. ʿIssaron(B)1	6460±70	5437–5323	100	2, 5
14	Rt 1630	N. ʿIssaron(B)	5625±70	4515–4428	100	6, 7
15	Rt 724B	ʿUvda 7	6410±120	5442–5241	100	4, 5
16	Rt 648A	ʿUvda 151	5670±85	4654–4368	100	3, 5
17	Rt 724D	ʿUvda 4	5400±110	4348–4049	100	4, 5
18	Rt 640A	ʿUvda 16	4800±70	3654–3509	95	3, 5
19	Rt 724C	ʿUvda 7	4540±100	3369–3044	100	4, 5
20	Rt 899A	ʿUvda 9	4530±50	3344–3107	100	4, 5
21	Rt 899B	ʿUvda 9	4520±60	3340–3104	100	4, 5
22	Rt 864B	ʿUvda 9	4440±180	3354–2895	100	4, 5
23	Rt 1436	ʿUvda 9	4440±60	3235–2984	100	5, 7
24	Rt 864A	ʿUvda 9	4310±90	3086–2703	100	4, 5
25	Rt 714A	ʿUvda 9	4070±100	2862–2469	100	3, 5
26	Rt 1452	ʿUvda 124/IV	4370±50	3031–2916	100	5, 7
27	Rt 1451	ʿUvda 124/IV	4370±50	3031–2916	100	5, 7
28	Rt 1419	ʿUvda 124/IV	4370±100	3105–2883	88	5, 7
29	Rt 1449	ʿUvda 124/IV	4285±60	3021–2704	100	5, 7
30	Rt 1448	ʿUvda 124/IV	4120±60	2865–2582	100	5, 7
31	Rt 1450	ʿUvda 124/IV	4075±55	2856–2494	100	5, 7
32	Rt 640B	ʿUvda 16	4400±60	3091–2919	100	3, 5
33	Rt 640C	ʿUvda 16	4280±80	3029–2696	100	3, 5
34	Rt 648B	B. ʿUvda 96/III	4250±50	2914–2702	100	3, 5
35	Rt 1420	ʿUvda 166	4210±60	2888–2668	100	5, 7
36	Rt 714B	ʿUvda 166	3850±80	2452–2197	100	3, 5
37	Rt 1421	ʿUvda 166	3680±50	2134–1974	100	5, 7
38	Rt 724F	ʿUvda 21	4015±80	2852–2403	100	4, 5
39	Rt 899C	Shaharut IV	3700±55	2172–1979	100	4, 5
40	Rt 771B	Shaharut IV	3580±130	2126–1743	100	4, 5

References for ^{14}C dates: 1=Ronen, 1980; 2=Goring-Morris & Gopher, 1983; 3=Carmi, 1987; 4=Carmi & Segal, 1992; 5=Avner, Carmi & Segal, 1994; 6=Carmi, Segal, Goring-Morris & Gopher, 1994; 7=Segal & Carmi, 1996. Nos. 1,3: excavated by Ronen; Nos. 4–6,13,14: by Goring-Morris & Gopher; Nos. 7,8,20–25: by Amiran, Arnon, Ilan & Avner; Nos. 9–11,18,32,33: by Yogev; Nos. 15,19: by Sass & Goren; Nos. 17,38: by Eisenberg; others: excavated by Avner.

* The full ^{14}C series of the N. ʿIssaron site (after Goring-Morris & Gopher, see in Table 6).

camps contained larger amounts of pottery sherds and flint than their summer counterparts. The circular shape of most of the tent bases is evident in all ancient tent camps of various periods. Remains of large, rectangular tents, such as those known in Bedouin societies today, were found only in a small number of sites, always in association with recent artifacts, such as segments of ropes and black 'Ghaza' sherds.

Comparisons with recent Bedouin tent camps in the Negev and Sinai, indicate that tent dwellers are not necessarily nomads or semi-nomads (Al ʿAref, 1936:117–119; Marx, 1967:83–86; Eldar *et al.*, 1992; Avni, 1992b; Goren-Inbar, 1993). These Bedouin populations practiced only short distances migration, 2–3 times a year, in order to adjust to the seasons, seek fresh grazing, and free the camp of insects attracted by the herds. Similarly, it is reasonable to assume that the ancient tent camps in ʿUvda Valley could have also been used for approximately half the year and were an integral part of the settlement system in the valley (see below).

Built dwelling sites may be divided into three main types:

Small sites. A group of 19 sites, consisting of 1–3 rooms, without a built-in courtyard. At first glance, they seem to be temporary dwellings for small or nuclear families. However, two excavated sites of this type revealed uninterrupted living levels, up to 80 cm thick, as well as installations and artifacts which contradict this impression (Fig. 8.8). One contained a well-paved granary.

Figure 8.7. Eastern ʿUvda Valley site 100, remains of a tent camp.

Figure 8.8. Eastern ʿUvda Valley site 124/IV, consists of two rooms. The main room, from west, with installations (excavated by Avner).

Medium size sites. A group of 132 sites, usually covering an area of 200–900 m^2 (Figs. 8.9, 8.10), though some were larger, up to 1,600 m^2 (Fig. 8.11). All have the same basic pattern, a courtyard or several courtyards surrounded by clusters of rooms. The majority of courtyards and rooms are circular, while some are square or rectangular. One site (No. 9) breaks the pattern by having one row of five square rooms with a courtyard at each end (Fig. 8.12). At a later stage (EB IV), a dwelling unit consisting of a circular courtyard and circular rooms was built on top of the earlier southern courtyard. Most sites of this group are suited to one extended family, though the largest ones may have contained up to five (see below). Ten of these sites were excavated (*Hadashot Archaeologot* 54–55, 1980:35–49). Small cattle bones were discovered in all of them, while some also contained dung deposits. Evidently, herding was one element of the inhabitants' economy (see below).

Large sites. This group comprises only three sites, covering 3,300–4,500 m^2, characterized by a scatter of single or clusters of rooms, with almost no courtyards. It is difficult to discern the precise plan of these sites since they were all damaged by later Nabataean building activity. In only one, a fair proportion of the rooms could be identified, at least 23, and two separate courtyards.[6]

Figure 8.9. Eastern ʿUvda Valley, aerial photograph of site 18, consists of courtyards and rooms (excavated by A. Eitan).

Figure 8.10. Northern ʿUvda Valley, an aerial photograph of site 19, from east (excavated by Eisenberg).

Figure 8.11. Eastern ʿUvda Valley site 173, from north, consists of 4 extended family dwelling units.

Figure 8.12. Eastern ʿUvda Valley, an aerial photograph of site 9, from west (excavated by Amiran, Arnon and Avner).

Figure 8.13. Eastern ʿUvda Valley site 121, from north, remains of a corral on a low ground.

In addition to the built dwelling sites, there were 32 built animal pens. They are characterized by broad courtyards surrounded by stone walls, with the remains of a few adjacent rooms or huts. This group is divided into two subtypes. One is built on a low ground (Fig. 8.13), the other on a slope and under a cliff with a rock shelter (Fig. 8.14). The animal pens each cover an area of 80–750 m^2, most being located in the wadis east of the valley (see map, Fig. 8.2).

Social analysis of the various site types (not described here due to lack of space), and the uniformity of artifacts indicates that tent camps, built dwelling sites, corrals, agricultural installations (see below), and cult and burial sites effectively belong to one culture and one socio-economic system.

8.3. Finds in the 5th–3rd Millennia Sites

In this section only those finds directly connected to economy will be addressed, such as flint agricultural implements, botanical remains, *etc.*

Tabular Scrapers (Fan Scrapers, Fig. 8.15). Some 300 fan scrapers were collected during the survey and excavations. In excavated sites they comprise 4–16.7% of the flint assemblage (Rosen, 1983a:206–238). These implements are especially dominant in desert sites rather than in fertile zones, from the early 6th to late 3rd millennia BC (Avner *et al.*, 1994:281, with references). A common assumption is that they are directly related to grazing, and were used for skinning and leather processing (*e.g.* Lee, 1973:252), and possibly for shearing as well (Henry, 1995:372). However, micro-wear analysis has shown they were more widely used for cutting meat and even wood and bone (McConaughy, 1979:69, 301–304; Bueller, 1988:30; Rizkana & Seeher, 1988:29–31; Rowan & Levy, 1991). In general, it seems that the presence of tabular scrapers do indicate a herding economy.

Figure 8.14. Eastern ʿUvda Valley site 106 from south, remains of a corral with a rock shelter.

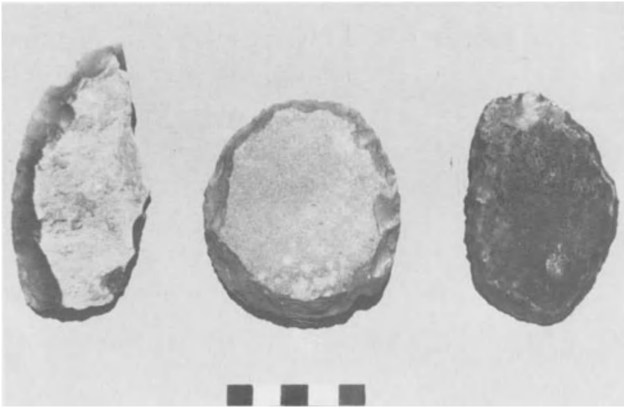

Figure 8.15. Tabular scrapers from various ʿUvda Valley sites (surface collection).

Figure 8.16. Flint adzes from various ʿUvda Valley sites (photographed by Radovan).

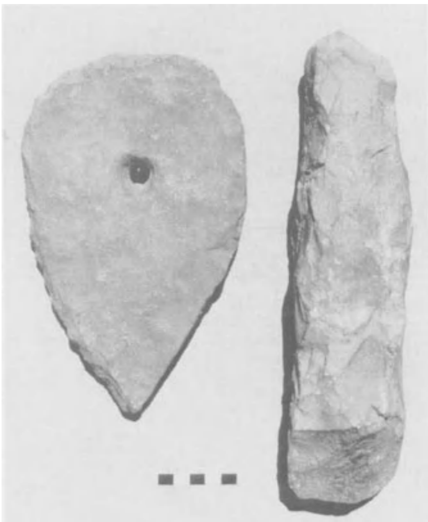

Figure 8.17. Limestone hoe from site 90 (left), and proposed limestone plough-tip (right) from site 96/III.

Adzes (Fig. 8.16). These tools were probably utilized for several purposes (Rosen, 1997:97, with references). One is tilling the soil, probably the primary use in ʿUvda Valley. Adzes are characteristic of Chalcolithic sites and are absent from Early Bronze assemblages. Possible explanations for their disappearance include replacement by copper tools at the end of the Chalcolithic period (McConaughy, 1979:217; Rosen, 1984b, 1997:93–97) and the emergence of the plough (see below). A complete adze handle was found in W. MurbaCat, Judean Desert (Benoit *et al.*, 1961:20), and similar tools made only of wood were still used in Africa until recently (Storkbaum, 1983:251). Adze are generally rare in the Negev south of the Beer-Sheva Valley; only 18 have been found to date in ʿUvda Valley.

Stone Hoes (Fig. 8.17). Stone hoes are another type of tilling tool. Only two were found during the survey, but their parallels are known from Chalcolithic sites of the Golan Heights (Epstein, 1978. Fig. 8.3A), and from Tuleilat Ghassul (Mallon *et al.*, 1934:70, Fig. 8.25, Pl.38:2; North, 1961 Pl. 11; Lee, 1973:273–275). Since these hoes are so different from adzes, it may be suggested that they were used for the preparation and maintenance of water channels and earth embankments (on the latter, see below).

Plough-Tip (Fig. 8.17). This implement was made of hard limestone, similar in shape and workmanship to the adze, but much larger. It is 28 cm long and weighs 2.6 kg (3–5 times larger and 13–32 times heavier than the average adze). It was found during excavation of a flint workshop adjacent to a threshing floor, with no evidence of wear. Due to its size and weight, it is difficult to describe it as another type of adze, however, based on parallels, we suggest it was used as a plough-tip. Incipient ploughs (ards), made of stone but shaped differently, are known from 3rd millennium sites in Syria, mainly from the El Hamah area (Steensberg, 1964). Closer parallels have been found in the Sahara Desert (Hugot, 1968; Rees, 1979; Milburn & Rees, 1984; Milburn, 1989) and in Mongolia (Licent & DeChardin, 1925), bearing the clear striation and sheen resulting from ploughing.

The dates of the plough tips from the Sahara and Mongolia remain uncertain. They were found on the surface in association with Neolithic artifacts, however, the Neolithic period in both areas ended in the late 3rd millennium BC On the other hand, the ʿUvda Valley tip was found in an Early Bronze I level, *i.e.* the 2nd half of the 4th millennium BC If the identification of the ʿUvda Valley implement is correct, then it must be considered the earliest archaeological relic of a plough thus far. This date is concurrent with the proposed emergence of the plough in human culture, based on accumulated circumstantial evidence (Sherratt, 1981:261–272).

Sickle Blades (Fig. 8.18). These are usually characterized by gloss created by friction of the blades with microscopic silica needles stabilizing the stems of cereal. Generally, several blades were connected in a row to a wood or bone handle, as exemplified by a Neolithic sickle from N. Hemar (Bar Yosef and Alon, 1988:16–19, Pl. 5). Several hundred sickle blades were found during the survey and excavations in ʿUvda Valley. Their proportion in the total flint assemblage is up to 10.7% (Rosen, 1983a:138–143, 199–238), surprisingly high in view of the environment.[7]

Three main types of sickle blades were distinguished in ʿUvda Valley: 1. Backed blades, with a steep retouched back and one cutting edge (Fig. 8.18:3,4). This type is characteristic of Chalcolithic sites in the Levant (Levy & Rosen, in Levy, 1987:288; Gilead, 1995:245–255) but continued into the Early Bronze (McConaughy, in Rast & Schaub, 1980:54; Rosen, 1997:65). 2. Canaanean sickle blades (Fig. 8.18:1,2) with a trapezoid or triangular section and two long cutting edges, are known mainly from the central and northern parts of the country, and rare in the south (Rosen, 1997:58–9). Only five have been found to date in ʿUvda Valley. Until recently these blades were considered to be restricted to the Early Bronze I–III periods (Rosen, 1983c, 1989:207, 1997:59–60). Nevertheless, their roots are much earlier.[8] 3. Crescent-shaped blades (Fig. 8.18:5–9). Only very few are known from Negev Highland sites (Haiman, 1986:58,93; Rosen, 1993;62–63), from Tuleilat Ghassul (Mallon *et al.*, 1934, Pl. 29:14; Koppel, 1940, Pl. 110:8) and the Hula Valley.[9] In ʿUvda Valley, however, more than 300 crescent-shaped blades were found, 116 from Site 9 alone, 19% of the flint tool assemblage (Rosen, 1983a:216). Therefore, these blades may be characteristic of the material culture of ʿUvda Valley. Only 20% of the blades had use-gloss, but under micro-wear analysis, others showed striation of non-cereal hard stems, probably legumes (Bueller, 1988:30). Indications for bone hafting were found on their back side (Bueller, *ibid.*). Most probably, several blades were hafted together, similar to the sickle from Nahal Hemar (Bar Yosef and Alon, 1988:16–19, Pl. V).

Grinding Stones (Fig. 8.19). More than 400 upper and lower grinding stones were collected during the survey and excavations, some were very worn. The majority were of exceptionally hard sandstone,[10] others were of limestone, flint or granite. The large numbers of grinding stones in ʿUvda Valley sites further emphasizes the importance of agriculture in the economy of the population.

Botanical Remains (Fig. 8.20). Botanical remains in archaeological sites are usually found charred, otherwise they hardly survive at all (*e.g.* Hopf, in Amiran *et al.*, 1978:64). The excavated sites in ʿUvda Valley showed no evidence of fire or other violent destruction. Therefore, the chance of detecting vegetation remains were low, and only limited attempts have been made to locate microscopic remains.[11] Nevertheless, two sites provided botanical remains which have been analyzed by M. Kislev. During excavation of the threshing floor (see below) an olive pit (Fig. 8.20:1) and a cultivated barley grain *(Hordeum sativum)* were discovered (Fig. 8.20:2). The excavation of an Early Bronze I tomb in a rock shelter (Avner, 1989a) revealed remains of several cultivated cereals: two-rowed barley (*H. distichum,* Fig. 8.20:3), six-rowed barley (*H. hexastichum,* Fig. 8.20:4), naked barley (*H. spontaneum,* Fig. 8.20:4) cultivated barley (*H. sativum,* Fig. 8.20:6) and domesticated wheat (*Triticum spelta,* Fig. 8.20:7).

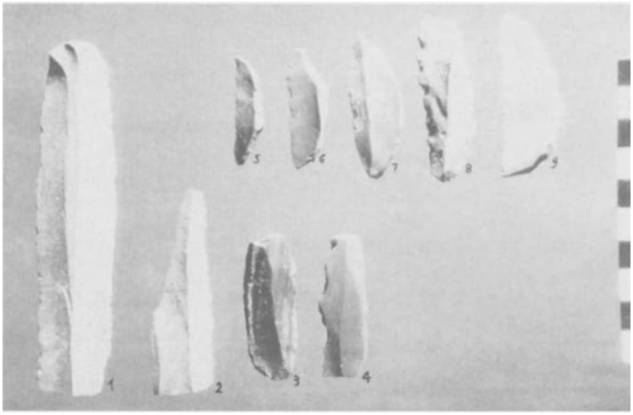

Figure 8.18. Various types of sickle blades from ʿUvda Valley sites (photographed by Radovan), #3,4 – backed-blade, #1,2 – Canaanean blades, #5–9 – crescent-shaped blades.

Figure 8.19. Grinding stones from site 9, as found in a stone compartment.

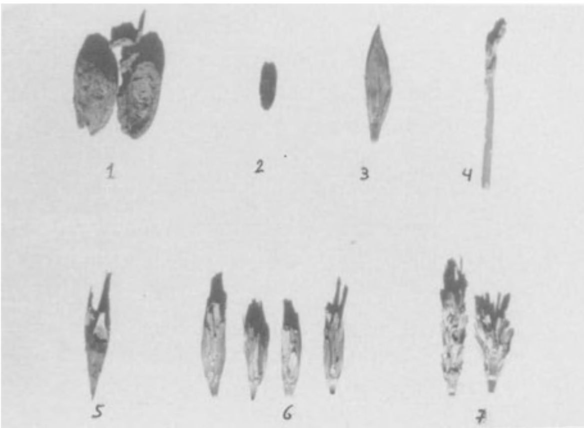

Figure 8.20. Botanical remains from various ʿUvda Valley sites: #1 – olive pits; #2–4 – domesticated wheat; #5–7 – domesticated barley.

Based on these limited finds, it seems that among the cereals, barley was dominant, while wheat was of only secondary importance. This situation was not unique to ʿUvda Valley; it was also found in Arad (Hopf, in Amiran *et al.*, 1978;66–67), in 3rd millennium Mesopotamia (Jacobsen, 1982:14–35), and in the Near East in general (Harlan, 1967:201). At the beginning of the 20th century, the Bedouin of the Beer-Sheva district still sowed barley on 90% of their cultivated area (Al ʿAref, 1936:157).

The find of the olive pit may be interesting in light of additional pits recently discovered during excavations of two 'desert kites' near Eilat. In one, they were found in association with Chalcolithic pottery sherds; and in the other, with Early Bronze finds (Holzer & Avner in preparation). Intensive utilization of olives for oil production had begun by the mid-5th millennium BC (Galili & Shavit, 1995; Kislev, 1995), while domestication probably occurred only during the 4th millennium BC (Neef, 1990). Olive horticulture and oil production greatly expanded during the Chalcolithic period (Baruch, 1986; Epstein, 1993). The olive pits from ʿUvda Valley and the Eilat area could have arrived through trade. However, during the Nabataean period, olive horticulture and oil production did take place in the Arabah Valley (Cohen, 1987), in a climatic regime similar to that of ʿUvda Valley and the southern Arabah. Therefore, it is not improbable that olives were grown in ʿUvda Valley, at least in past climatic conditions (see below). Moreover, several of the ʿUvda Valley sites from the second half of the 4th millennium and early 3rd millennium (Table 8.1, #18, 26) contained installations with anvils, hammer stones, mortars and basins, which may have been used for household oil production (Fig. 8.21).[12] For possible indications of other crop species, see below.

Granaries. Well-built installations paved with flagstones were found in two of the excavated sites (Fig. 8.22), while similar unpaved installations were exposed in most others, usually in corners of rooms. In light of parallels from other sites,[13] there is no difficulty in interpreting them as granaries, with the pavement made to prevent rodents from penetrating it from below. There were no remains of grain within the granaries. However, ceremonial, crop-related objects were found in several, among them sickle blades and grinding stones.[14]

The items described to this point, stone and flint tools in large numbers, plus the botanical remains and granaries, present an exceptional picture when compared to other contemporary sites in the Negev and Sinai. Together they attest to the agricultural economy of the ʿUvda Valley population, but these are not the only indications.

8.4. Threshing Floors

8.4.1. IDENTIFICATION OF THE INSTALLATIONS

During the survey 31 circular installations were found, 8–15 m in diameter. Most were dug into the rock surface to a depth of 30–80 cm (Fig. 8.23), while a few were

Figure 8.22. Site 16, a paved granary, from west (excavated by Yogev).

Figure 8.21. Site 124/IV, domestic installation, probably for olive oil production.

Figure 8.23. Site 96/III, aerial photograph of two threshing floors, dug into an older, large threshing area, from north.

circles of beaten earth. Pottery sherds and flint implements from the 4th–3rd millennia BC, as well as from later periods, were collected in and around them. We interpreted these installations as threshing floors (Avner, 1979, 1982a, 1982b), though without known archaeological parallels, this was difficult to support. Over time, however, we have collected ample conclusive evidence; presented here first are a few ethnographic arguments:

1. During a field trip with Haiwat Bedouin (December 29, 1979), who cultivated ʿUvda Valley until the mid-20th century (see below), they confidently identified these installations as threshing floors. They claimed that they were built by their fathers, and explained in detail the successive stages of threshing barley and wheat.

2. Near Borot Lotz in the Negev Highland, similar installations were observed (by Y. Porat) adjacent to agricultural terraces. During my own visit to the site, in April 1980, one of these threshing floors was found covered by a dense growth of barley, the result of seeds remaining from the previous Bedouin threshing season. Around the threshing floor I collected flint implements and pottery sherds from several periods, beginning with the 4th millennium BC Another threshing floor excavated nearby (Avner, 1981) revealed a similar picture.

3. A threshing floor still used by Bedouin was found in the western Negev Highland by M. Haiman (Fig. 8.24), confirming the function of these circular installations without need for further explanation. Flint and pottery later collected by Haiman indicated its ancient origin.

4. Small circular installations, about two meters in diameter were adjacent to all threshing floors (Figs. 8.23, 8.25). Similar installations were found near recent Bedouin threshing floors in Nahal Yaham (southern Negev) and in Be'er Sheva Valley. In the latter, cereal straw and grains were found indicating their use as granaries. Therefore, these circles further support the interpretation of the old threshing floors. Similar to those of the Bedouin, they were most probably used for storing clean grain during the threshing season, prior to being transferred to granaries in the dwelling sites.[15]

As described above, some threshing floors were beaten earth circles, sometimes scarcely distinguishable. Probably, many more of this type in the cultivated fields left no remains; and if so, their original number in the area was far greater. Even as things stand, the concentration of threshing floors in ʿUvda Valley is the largest known in Israel; twenty-six were built within one square kilometer on the eastern side of the valley.

8.4.2. EXCAVATIONS OF A THRESHING FLOOR IN ʿUVDA VALLEY

The aerial photograph (Fig. 8.23) of one site in the mouth of Nahal Yetro exposed intriguing details, mainly a light-colored strip stretching north-south, which included two rock-cut threshing floors. The survey revealed that the upper layer of weathered rock had been removed along the strip, while the circular threshing floors

Figure 8.24. N. Horsha, Negev Highland, an ancient threshing floor presently used by Bedouin (after Haiman 1986, cover).

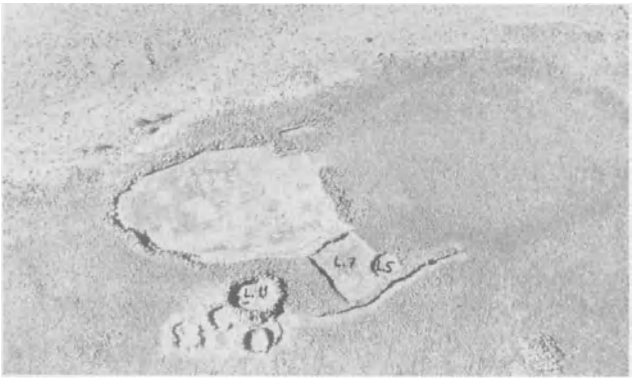

Figure 8.25. Site 96/III, an aerial photograph of the northern threshing floor, from east, after excavation.

Figure 8.26. Flint sledge blades (top) and flint points (bottom) from site 96/III (photographed by Radovan).

were apparently cut into this cleared area at a later time. This was confirmed during the excavation when more information was derived concerning the history and development of the threshing floor. Flint and pottery, as well as one radiometric date, indicate 4th and 3rd millennia BC dates, with continuing use through several periods up to the 20th century. Following is a description of the stages as discerned during excavation, from early to the late.

In the first stage, the upper layer of weathered rock was removed from a strip of 120 m long and 10–20 m wide. These rocks had been used for building a low wall, whose remains were found mainly to the east of the excavated threshing floor, as well as to the northwest (Fig. 8.25). This wall, termed '*Shall*' in the Bible (2 Sam 6:7), was built around the perimeter of the cleared rock surface, similar to the walls surrounding most threshing floors in the valley. Excavation of the area between the threshing floor and a section of wall to the east (Loc. 7) exposed the earliest cleared rock floor. Dark gray patination preserved on this floor indicates prolonged exposure and use, probably for several centuries.

During the second stage, part of this large floor (in Loc. 7) remained unused and was covered by a thin level of dirt (5–7 cm) mixed with ash, some pottery sherds and a large amount of flint debitage. This level also included the plow tip (see above) as well as two crescent shaped arrowheads dating it to the Early Bronze I, the second half of the 4th millennium BC (*cf.* Rosen, 1983b). In the third stage, a granary (Loc. 5) was built on top of the dirt level, testifying to the continued use of other parts of the early threshing floor.

In the fourth stage, the two circular threshing floors were cut into the rock floor of the large cleared area. The northern one measures 13 x 17 m and is 0.6 m deep. This depth was reached after three separate cutting phases, each removing another rock level in an attempt to renew the decaying rock floor. The *Shall* surrounding the threshing floor was built from some of the dug-out rock, while the rest was dumped on top of the living level to the east (Loc. 7). An inclined thin wall was built to support the dumped material and prevent it from falling back onto the threshing floor.

At this stage, a circular room (Loc. 11), was built, and served as a flint workshop. A flagstone bench was found in it, as well as a stone anvil, hundreds of flint debitage pieces and several dozen tools (see below). In the adjacent cluster of granaries, a pair of grinding stones was found, along with the remains of burnt straw, one burnt grain of barley and one olive pit. These items further support the interpretation of the smaller circular installations as granaries. They were all built on top of a cleared rock floor, functioning similar to the stone pavement in granaries within the dwelling sites. Use of the two rock-cut threshing floors, the flint workshop and part of the granaries, continued through the Early Bronze Age until the early 2nd millennium BC, and in part, in the Iron Age and Nabataean periods as well.

8.4.3. FLINT IMPLEMENTS IN THE THRESHING FLOORS

Among the finds from the excavated threshing floor, mainly inside the flint workshop, were 39 flint tools, somewhat similar in shape to tabular scrapers. An additional group of 29 similar tools was found well ordered in a pile (Fig. 8.26) in a granary adjacent to the southern threshing floor in the site (the upper one in Fig. 8.23). Others were found on the surface surrounding each surveyed threshing floor. This type of tool consists of several subtypes varying in size and shape. Based on ethnographic parallels, mainly from Turkey and Cyprus (Fig. 8.27),[16] these tools can be identified as sledge blades. Similar tools from 3rd millennium Mesopotamia were also interpreted as sledge blades based on the use-gloss found on some (Adams, 1975).

There are both differences and similarities between the tools from the ʿUvda threshing floors and those described in articles referred to in Note 16. The essential difference is the lack of gloss on the ʿUvda blades, for two possible reasons. First, since most were collected from the surface, long-term preservation of the gloss was disturbed. The second was brought to light by Bueller after micro-wear analysis. He found vegetation wear with no gloss, created by non-cereal plants, probably legumes. He also identified wear on the edges created by friction with limestone, most probably the rock floor (Bueller, 1988:32, Pl. 7,8). On this basis it may be suggested that the sledge, equipped with the larger blades, was used for threshing chickpeas *(Cicer aritinum)*. Chickpea was already domesticated in the Early Neolithic

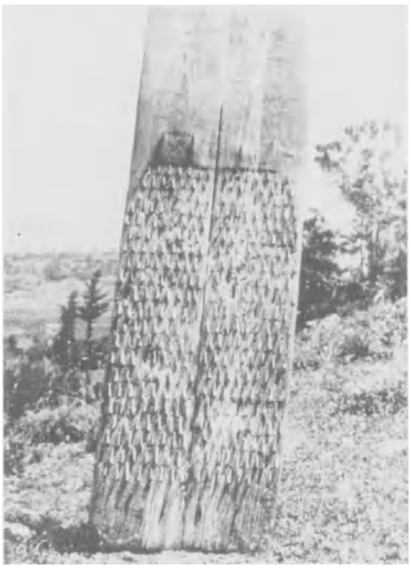

Figure 8.27. A sledge equipped with flint blades from Cyprus (after Crawford 1935, Pl. 1).

period (Zohary & Hopf, 1993:101–106) with threshing still being practiced in traditional agriculture until the middle of the 20th century (Avitsur, 1984:3). It is logical that chickpea threshing required larger sledge blades than those used for cereals, and the finds from ʿUvda Valley encompass a variety of sizes for both uses.

It should be noted that most sledges in recent traditional agriculture in Israel were equipped with basalt stone points (not blades). However, in areas lacking basalt, flint blades were used (Avitsur, 1965:20, 1966:64, Fig. 32). A complete sledge of this type from the Judean Hills is on exhibit at the Kfar Etzion Field School.[17] In addition, around several threshing floors in ʿUvda Valley, including the excavated one, flint points were found, some still bearing use-gloss (Fig. 8.26), similar in size and shape to the basalt stones used for sledges in the northern part of the country. Their low numbers indicate they were probably of secondary importance to the sledge blades.

8.4.4. HISTORY AND DEVELOPMENT OF THE THRESHING FLOOR

The finds from the excavated threshing floor and identification of the various stages enable us to suggest a reconstruction of the development and evolution of threshing floors in general:

When wild cereal collection began, some threshing became necessary to separate the grain from the glumes. To our present knowledge, intensive cereal collection began 19,000 years ago (Kislev *et al.*, 1992). The domestication process brought a continual selection of individual stems with strong rachis that minimized the loss of grain during harvesting. As a result, domesticated cereal demanded systematic sowing on one hand, and efficient threshing on the other hand (Harlan, 1967:199; Zohary, 1973:309f, 313f; Redman, 1978:110,122f; Kislev, 1984:63; Hillman & Davies, 1990:161–165). The simplest threshing method, beating by stick, was most probably employed in the incipient stage of agriculture. Beating was also mentioned in the Bible (Judges 6:13, Ruth 2:17), and actually persisted until the 20th century (Dalman, 1933, Abb. 25). Since the shape and dimensions of the first stage, large threshing floor in ʿUvda Valley is not compatible with the circular treading of beasts, it should be assumed that beating was the method practiced here as well. One may imagine several families laying and beating their heaps of crops along this threshing floor. The direct dating of this first stage is difficult, however, based on dates of the next stages, it should be prior to 3500 BC

The next stage was the emergence of the circular threshing floor. The shape is undoubtedly suited for threshing by beasts treading the crops while walking in circles, thereby separating the grains with their hooves. This method, known in ancient Egypt (Skira, 1954:76), was employed in the Near East until recently (Avizur, 1966:62,1984:14–16), and was the only method used by Bedouin in the Negev (Fig. 8.24). In all locations, threshing was achieved by a group of animals, usually five or six. Hence, we must deduce that the first appearance of the circular threshing floor directly corresponds to the first enhancement of oxen and asses for work, one innovation of the 4th millennium (Sherratt, 1981:263–266; Epstein, 1985; Ovadia, 1992). In ʿUvda Valley, the stage of cutting the circular threshing floor was dated to the Early Bronze I, while in Nahal Elot, the Negev Highland, I found one Chalcolithic adze in each of two threshing floors. It is therefore likely that the circular threshing floor had already appeared in the Chalcolithic period.

During the third stage a sledge pulled by animals was invented. Adams (1975) proposed that the sledge appeared in Uruk during the 3rd millennium BC Now, as a result of the ʿUvda Valley excavations, we may date the appearance of the sledge equipped with flint blades, to the Early Bronze I, the second half of the 4th millennium BC

An interesting philological detail connects ʿUvda Valley, in Arabic *Wadi ʿUqfi*, with the threshing floor. *ʿUqfi* means a sledge wooden pole made of wood (Avitsur, 1966:76), originating from the root 'walk about'. However, since Bedouin in the Negev and Sinai never used sledges, they could not have been the source of the geographical name. An alternative source could be Nabataean, with numerous remains in the valley and their Aramaic language (Naveh, 1982:9–11, 135–159). The root 'walk about' does appear in Aramaic documents (*e.g.* Jean & Hoftizeizer, 1965:220). Therefore, it is possible that the source of the name of the valley is ancient, and was passed on to the Arabs as were many other geographical names. In any case, it is intriguing to find an geographical- agricultural name in such a harsh desert environment.

8.5. ʿUvda Valley Agriculture in Light of the Environment

Following the description of the agricultural evidence of the 5th–3rd millennia BC, it becomes necessary to examine the environmental conditions which allowed human existence in this desertic area.

8.5.1. CULTIVATED SOIL AND WATER

The soil along the eastern side of ʿUvda Valley is unique in its physical characteristics, which bears botanical implications. Lime-sand, the unique component, makes the soil light, well ventilated, easily tilled, and highly water absorbent. These characteristics are quantifiable. At a depth of 0.5 m it consists of 50–70% lime-sand, 20–40% silt and 10–14% clay. The soil is slightly alkaline (pH 7.8–8.35), with a low level of salinity. The water absorption capacity is very high, up to 39% of its volume.[18] The clay percentage increases with depth, a situation which minimizes water loss through seepage. This fact, along with the high water content, enabled excellent watering of the soil in an efficient root-depth for cereal, bushes, and even trees. These qualities are well demonstrated by fairly thick plant growth on the eastern side of the valley; and, in contrast to the southern Negev in general, outside the wadi channels (Fig. 8.28).

In a situation of low precipitation (28 mm annual average), only floods can supply the amount of water necessary for cereals. The drainage area of ʿUvda Valley is 400 km^2, mainly to the south, 500–884 m above sea level. Most of the surface is barren rock which absorbs a comparatively small amount of the rain water. In addition, most desert rains fall in a concentrated way (Shanan *et al.*, 1967; Finkel & Finkel, 1979; Sharon, 1979) so even a small amount is often enough to create floods. Because all wadis merge on the eastern side of the valley, this area enjoys the best flood irrigation (Fig. 8.29). Due to the very low gradient of the valley (see above) flood water flows slowly and is well absorbed by the soil.[19] Following the flood, a thick growth of wild cereal appears (Fig. 8.30)

Nevertheless, the citizens of the valley did not rely on natural conditions alone. Observation of surface and aerial photographs reveals low earth embankments perpendicular to the water channels, sometimes with one layer of rocks on top (Fig. 8.31). These embankments may have contributed greatly to the quality of the cultivated land. They retarded the flow of water, further increasing the amount of water absorbed by the soil, prevented soil and seed erosion, increased the sedimentation of new soil enriched with organic material with each flood, and widened the irrigated strip. The resulting advantage enabled the citizens to plough and sow the land before the first flood, and not 2–3 weeks later, as practiced by the Bedouins (see below). In addition, the enrichment of the soil with organic material brought by the flood could enable cultivation year after year, with no need to lay the fields fallow.[20]

Although it is not totally clear when these embankments were first built, evidence suggests that they already existed by the 4th millennium BC (to be described elsewhere). Supporting this possibility was the discovery of an agricultural field in Nahal Paran, 40 km north of ʿUvda Valley. This site consists of similar embankments built over an area of seven hectars, with one course of rocks at the top, forming rectangular 'limans' (Fig. 8.32). Twenty flint adzes were collected in the site, attesting to its Chalcolithic date (Avner, 1997). The site provides unequivocal evidence for the existence of agricultural flood water engineering and soil improvement during the Chalcolithic period. The Bedouin population of the Negev and Sinai, on the other hand, did not build any embankments (see below).

The cultivated strip in eastern ʿUvda Valley is 12 km long and averages 0.5 km wide, equal to 600 hectares (Fig. 8.2). Another plot in the south center of the valley provides an additional 150 hectares, cultivated fields in the wadis to the east render another 250 hectares, and Nahal Hayun which drains the valley to the north, adds approximately 200 hectares with embankment remains. Altogether, these cultivated fields covered at least 1200 hectares. This brings to light the special aspect of ʿUvda Valley. While in the Negev Highland cultivated plots are divided into relatively small

Figure 8.28. *Artemisia monosperma,* Mediterranean plant, growing well in the lime-sand of ʿUvda Valley.

Figure 8.29. Eastern side of ʿUvda Valley after a flood, from north.

Figure 8.30. Eastern ʿUvda Valley, a natural growth of wild cereals, following a flood.

wadis, ʿUvda Valley provides large, uninterrupted fields with high quality soil, efficiently irrigated by floods.

Besides embankments and the broad cultivated fields, the presence of small semicircular stone terraces should be mentioned. These are built on the lower part of the slopes, consisting of one row of medium size fieldstones, up to four courses high and three meters in diameter. They are irrigated by rain water through natural channels and probably made for growing bushes or small fruit trees. Most plausible are vines, which were domesticated during the Chalcolithic period (Zohary & Hopf, 1993:148).

8.5.2. DRINKING WATER

The presence or absence of available drinking water for man and animal is critical to the existence of any settlement, especially in the desert. Five natural water sources are found around ʿUvda Valley, within a half day walk from the heart of the settlement, but only one, the Yotvata Oasis, supplies an unlimited amount of water. Nevertheless, during the survey we encountered three different methods for rain water collection and the utilization of underground water:

A. In 36 sites a series of dams were found in small wadis (Fig. 8.33). Usually, only limited remains of 3–6 dams are visible, but in one site 17 dams were found in a single wadi channel. All dams are situated near 5th–3rd millennia BC dwelling sites. Since these dams are generally in a poor state of preservation, one may assume that many others were totally eroded or covered. Another variant of this method was the conveyance of flood water into caves, as has been detected in two sites to date.

B. Along the cliff edges east of the valley is a series of closed depressions developed from vertical faults in the rock, and filled with marl and clay soil. They naturally collect and retain the rain water. Inside or near most of these depressions dwelling sites of the 5th–3rd millennia BC were found, enabling utilization of the accumulated water (Fig. 8.34). In some depressions, Bedouin dug cisterns to collect and store rainwater. They used to clean them before every winter, then covered them with plants after filling, to prevent evaporation. It is highly likely that similar cisterns were also excavated in the past in these depressions.

C. In the same area, east of the valley, four wells have been found to date, two of them were never completed. One was fairly well preserved, enabling study of how it functioned. The well's neck is 0.8 x 1.4 m, cut into a hard and cracked layer of limestone, which has a low gradient from east to west (Figs. 8.35, 8.36). At a depth of 1.5 m the rock changes to a soft chalk, where the well widens to a bell shape, presently filled with debris. The cracks in the hard rock enable rain water to seep down, while the chalk layer stops the seeping water, causing it to flow slowly and drip into the well. Flint and pottery collected from around the well dates it to the 5th–3rd millennia BC. However, the well still remains moist today.

The well in ʿUvda Valley is one of the earliest dated in the Near East. The oldest known are two wells in Haçilar, Anatolia (Mellaart, 1970:35) and in the submerged Neolithic village near Atlit, northern Israel (Galili & Nir, 1993). Both sites have

Figure 8.31. Eastern ʿUvda Valley, remains of embankments, an aerial photograph from northwest before construction of the air base.

Figure 8.32. N. Paran, remains of Chalcolithic embankments, aerial photograph from west.

Figure 8.33. Site 124/II, remains of water dams.

Figure 8.34. Eastern ʿUvda Valley, a well in site 135.

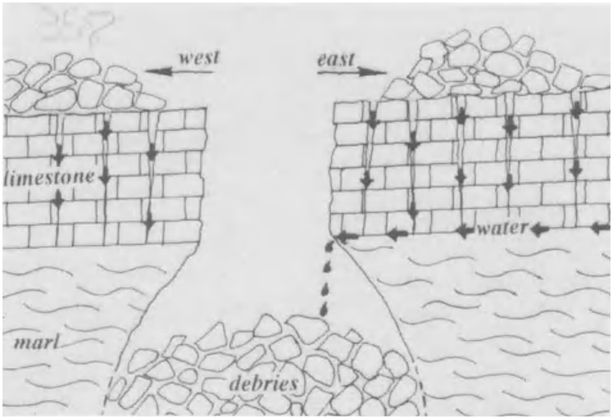

Figure 8.35. Sketch of the well in site 135.

Figure 8.36. Eastern ʿUvda Valley site 78A, a dwelling unit built in a depression which collects rain water.

been dated to the 6th millennium BC. A well in the site of Abu Hof, central Israel, was dated to the Chalcolithic period (Alon, 1988), as was the well in Wadi Sirhan, northern Saudi Arabia (Zarins, 1979:76). This date is also suggested for the wells in ʿUvda Valley, based on collected flint and pottery.

Another well was discovered by A. Shapira on Ma'aleh Shaharut (Avner, 1989a). It is covered by clay soil (of Orah formation), and the fill is moist even during the summer. A few flint pieces were collected near the well, in addition to recent Ghaza ware. In 1989, a flow of water appeared 20 m below the well, with a supply of 100 liters per hour, while another spring appeared 500 m to the south. Both are still active today and have created a narrow grove of tamarisk trees.[21] Ancient remains located below the springs indicate activity during the 5th–3rd millennia BC.

Among the three artificial means for water collection described above, the series of dams are the most important quantitatively. An average series of dams can contain at least 100 cubic meters. If annual water consumption per capita in the past was 1 m^3 (Rosnan, in Amiran *et al.*, 1978:14) this amount of water could support 100 people. If we consider water loss, due to seepage and evaporation, and use for the herds, it could be roughly calculated that a series of dams would support one extended family, *ca.* 25 people with their herds. If so, the series of 36 dams identified during the survey could supply water for 900 people. As stated, however, the original number of dam sites was higher. In general, we may assume that utilization of all water sources, natural and artificial, could support a considerable population, even through the summer. The estimated density of the population will be addressed below.

8.5.3. BEDOUIN AGRICULTURE IN ʿUVDA VALLEY

During the survey, in addition to the ancient remains, we also found many indications of Bedouin agriculture from past generations. Several caves contained stored agricultural equipment, such as iron sickles and hoes, iron plow tips, wooden plows and millstones (Fig. 8.37). We also found plowing marks, and hundreds of granaries (Fig. 8.38).

Short descriptions of Bedouin cultivating the valley were provided by several surveyors. Musil, who visited the area in 1902, mentioned the valley while describing the people of ʿAqaba, and a similar description was given by Braslavi (1952:471).

> In Wadi ʿUqfi, the soil is good for cultivation and the crop is plentiful when the rain is sufficient to create floods. In this valley the citizens (of ʿAqaba) are renting plots from the Haiwat, sowing wheat and barley and living in tents by the fields in the seasons of sowing and harvesting. After the harvest they return to Aqaba with the threshed grain (Musil, 1926:85).

In December 29, 1979, I visited the valley with four Bedouin elders of the Haiwat tribe who had cultivated the land in the past, in order to learn their agricultural

Figure 8.37. Eastern ʿUvda Valley site 60, a Bedouin storage cave with agricultural tools: A wooden plough, an iron plough-tip, two iron sickles, a millstone, *etc.*

Figure 8.38. Northern ʿUvda Valley, aerial photograph of Bedouin granaries.

practices. The visit was undertaken with the assistance of anthropologist F. Stewart, who was studying the legal system of the tribe. Following is the essential information:

A. Approximately two weeks after a flood, while the soil was moist but not too wet, they sowed barley and wheat by hand. Then, in order to cover the seeds, they plowed the soil with light wooden ploughs, pulled by donkeys or camels (*cf.* Marx, 1967:20–22). No attempt was made to build embankments or control the flood water.

B. Between the years 1939–1948, the Bedouins cultivated the valley four times, when the amount of rain was sufficient to create floods. However, almost every year of sowing provided a good crop, enabling grain storage for 2–3 years, with enough grain reserved for sowing and even sale of part of the yield in ʿAqaba and Al ʿArish. In addition, the tribesmen rented land to the citizens of ʿAqaba in return for 1/4 or 1/2 of the yield. The Bedouin claim that their yield reached 800 kg per hectare (translated from their own measurements of area and weight) and the number of people subsisting on the yield was more than 4,000. Although there is some doubt in the accuracy of these figures, there are several indications (see below) that they are not that far from reality. Threshing took place both on temporary threshing floors in the fields and on the ancient ones, which they claimed were made by their own fathers.

C. All the granaries excavated in the ground presently visible in the valley (Fig. 8.38) belonged to the Haiwat. Each had the capacity of 10 *quntars* of clean grain, which equates to three tons. The number of granaries in the valley known to date is over 400, and could therefore contain a total amount of 1200 tons of grain. However, according to the Bedouin, only a part of the yield was stored in the granaries, while the rest was immediately sold or distributed in the tent camps throughout their tribal territory. If a Bedouin family of five consumed approximately 500 kg annually (see below), then the amount of grain stored after a successful season could have supported at least 2,400 families. But, since this yield was to be used over a 2–3 year period, as well as for sowing, we reach an estimation of 800 families or 4,000 people. This figure is four times higher than what was known about the tribes in the first half of the 20th century (Marx, 1967:12). Yet, it seems that ʿUvda Valley was indeed an important food source for the Bedouin. According to the testimony of members of Kibbutz Yotvata, the Haiwat still continued to penetrate the Sinai border after the 1948 war until as late as 1957, in an attempt to sow the land of the valley and cut the yield.

The Bedouin success in gaining good yields in ʿUvda Valley, with a relative inferior agro-technology, and in present day climatic conditions, gives additional support to the scenario of the ancient agricultural settlement described above. The question of the past climate will be addressed later.

8.6. Ancient Population and Demography of ʿUvda Valley

Several questions arise from the documented remains in the valley concerning the nature of the sites and settlement in general. Were the inhabitants nomadic, seminomadic or permanent? Were the sites seasonal? And is it possible to estimate the population density?

8.6.1. NOMADIC OR PERMANENT SOCIETY?

Ethnographic and sociological studies show a clear distinction between full nomads and partial nomads. Full nomads worldwide developed from agricultural, permanent settlements, not from hunters and gatherers. They subsisted on a 'specialized economy' of herding only, while for all other living needs, excluding weaving, they relied on various craftsman. Also, they were dependent on riding and carting animals that enabled a migration cycle of hundreds of kilometers twice a year (*e.g.* Rowton, 1973, 1974; Khazanov, 1984). Partial nomads, on the other hand, maintained a 'combined economy' based on grazing, agriculture, crafts and trade. The proportion between the economical branches may have varied from place to place, according to local environmental conditions. Partial nomads practiced only limited annual migration cycles, over tens of kilometers only. Sometimes they moved their camp only a few hundred meters in order to accommodate the changing seasons (Al ʿAref, 1936:117–119; Marx, 1967:84,86). In the Near East, camels could have served the nomads but they were not domesticated before the 12th century BC (Resto, 1991; Kohler-Rollefson, 1993, *contra*-Ripinsky, 1983). Moreover, some scholars assert that full nomadism never existed in the deserts surrounding the Levant, and even Bedouin of recent centuries were not true nomads (*e.g.* Marx, 1992). In light of these arguments, the remains of ʿUvda Valley could be attributed to partial nomads or permanent dwellers only.

8.6.2. SEASONAL OR PERMANENT SITES?

Many sites similar to those in ʿUvda Valley are known in the Negev, Sinai, Jordan and other desert areas of the Near East. They are most commonly attributed to nomadic or semi nomadic groups and are considered to be 'short lived' (*e.g.* Beit Arieh, 1982:155, 1986:51), 'seasonal', 'inhabited for only a few years' (*e.g.* Haiman, 1986:16, 1989:185). In ʿUvda Valley, however, a large portion of the sites are apparently permanent settlements, for the following reasons:

A. In similar sites outside ʿUvda Valley, only a small number of surface finds are usually collected (*e.g.* Avni, 1992a:15,25), and sometimes even an excavation does not render finds (*e.g.* Cohen, 1986:98). ʿUvda Valley sites, on the other hand, yielded ample surface finds and excavated artifacts. They included pottery sherds, flint implements, grinding stones, ostrich egg shells, sea shells, copper nodules, various beads and so forth.

B. The architecture of many ʿUvda sites is fairly good. Many are built of large, roughly cut stones, sometimes over 1 m long. In several places, walls up to six courses have been preserved, 1.4 m high. In most sites, several strata were found on top of one another, a phenomenon only rarely found in similar sites of other desert areas. The living levels show a slow and gradual accumulation, while gaps in settlement are not discernible (probably excluding site 17, Beit Arieh, 1989:195). The architecture and accumulation of living levels contrasts with the common situation in the Negev Highland sites (*cf.* Haiman, 1994, 1995).

C. Radiometric dates obtained from dwelling and cult sites in ʿUvda Valley, as well as in a burial site in Eilat (Avner *et al.*, 1994, Avner in print), indicate a lifespan much longer than expected, several hundred years and even more. Two examples will be addressed. In site 9, the lower level (known only by probes), was dated by finds and a single radiometric date (Table 8.2, #8) to the 6th and 5th millennia BC The middle level, with six dates (#20–25) indicates a span of approximately 500 years, between *ca.* 3230 and 2700 BC The upper level belongs to the

Early Bronze IV, late 3rd and early 2nd millennium BC. Altogether, the site was generally inhabited for four millennia. The site in N. ʿIssaron was dated by 35 radiometric dates, indicating a time span from *ca.* 8200 to 3500 BC (Table 8.1). Although the sites are not considered to have been occupied every day or even every year, these results are still surprising and of great cultural and settlement significance.

D. The ground plan of the sites indicates suitability mainly to winter conditions. For example almost all doorways face east, towards the morning sun, and are protected from the westerly winter winds. Nevertheless, the rich agricultural finds imply the presence of the population during other seasons. As described above, the embankments in the fields enabled plowing and sowing before the first rain. Therefore, the plowing season would have begun in August or September. During May and June the crops were cut, and threshing took place during July and August.[22] This way a year-round cycle of agricultural activity was completed. It is certainly possible that parts of families, or their employees, attended the herds some distance from the dwellings. Also, it is possible that families, or parts of them, went to Timna Valley, less then a day's walk away, to produce copper during the winter. Even then, the definition of the settlement as permanent is still valid.

8.6.3. ESTIMATION OF POPULATION DENSITY

As previously described, ʿUvda Valley is characterized by an exceptionally high density of sites. To date, 233 sites of the 5th–3rd millennia BC have been documented, many consisting of more than one element. Altogether, 154 dwelling units have been counted, 40 tent camps, 32 animal pens, 36 water installations, 31 threshing floors and 42 cult and burial sites, over an area of 50 km^2.

These figures raise the next question. What was the population size at its climax (the 3rd millennium BC)? In an attempt to arrive at an estimation, precautions will be taken to avoid exaggerated figures: 1. We shall ignore the tent camps. 2. We shall ignore the 32 animal pens with their adjacent rooms. 3. We shall assume that only half of the 3rd millennium sites were simultaneously occupied, although all excavated sites revealed habitation levels of that period. As stated, 154 dwelling units were surveyed. However, if the survey had been completed, the number would be over 200. Therefore, we will assume 100 sites were concurrently inhabited. Since the possible margin of error in demographic calculations is fairly large, I scrutinized several methods of calculation practiced in the research:

A. *Family units.* This simple method assumes that each dwelling unit inhabited one extended family of 25 people in average (*cf.* Gopher, 1981:125), although many sites, due to their area and ground plan, were suited to larger numbers. According to such calculations, the population of 100 sites would be 2,500 people.

A more detailed analysis of the ground-plans reveals the following data. The overall area of 100 sites (including open spaces) would be 69,023 m^2. Of this, the built area (rooms and courtyards) was 30,465 m^2; the roofed area—5,975 m^2 (for a total of 710 rooms) and the courtyards—24,492 m^2. This data will be used for the following calculations:

B. *Persons per total inhabited area* (including open spaces). Demographic, ethnographic and archaeological research in the Near East offered coefficients of between 20 and 50 persons per 1,000 m^2 of settlement.[23] The average of these various coefficients is 30 persons. Accordingly, the population density in ʿUdva Valley would be 2,070 people.

A study based on a census of Palestine in 1938 demonstrated, among other things, that settlement density in cities and villages increased when they were smaller. In settlements covering one hectare or less, the population coefficient was 50 persons per 1,000 m^2 (Biger & Grossman, 1993, Table 1). The main reason for the density difference between large and small settlements is that every place required open spaces. The smaller the settlement, the better the use of the surrounding open spaces; and therefore, the demand for open spaces within it is lower (Hassan, 1981:69, Fig. 6.2; Biger & Grossman, 1993:23). Since all settlements in ʿUvda Valley are much smaller than one hectare, the coefficient of 50 persons per 1,000 m^2 should be adopted. Accordingly, the population of the valley would be estimated at 3046 people.

C. *Persons per built area* (rooms and courtyards). Density coefficients proposed for built areas are 5 or 10 m^2 per person (Marfoe, 1980:319; Hassan, 1981:74, Fig. 6.7, respectively). Accordingly, the population size of the valley would be 3046 or 6092.

D. *Persons per roofed area* (rooms only). Similar figures, 5–10 m^2 per person, were proposed for density coefficients based on room area only (Naroll, 1962; Kramer, 1978, 1979; Hassan, 1981;74). These make the ʿUvda Valley population 597–1195, low figures when compared to the above.

E. *Persons per numbers of rooms.* Assuming the number of rooms identified during the survey are realistic, that 2/3 of the rooms in each site served for dwelling, and that each dwelling room served a nuclear family of 5 persons (*cf.* Hassan, 1981:74–75), the population of the valley would have been 2366.

F. *Cultivated area.* Another calculation may be of the cultivated grain area and estimation of yields. As described above, yields of the Haiwat Bedouin were stated to be up to 800 kg per hectare. For our calculation, a lower amount of 500 kg will be adopted,[24] although, in my opinion, the ancient agriculture was of a higher level then that of the Bedouin (see above). If the cultivated area of the ʿUvda settlements was 1,200 hectares, the yield could have reached 600 tons. Even if 25% of the yield was contaminated during storage, 11% reserved for sowing, and 16% used as animal fodder (McAdams, 1981:86), the amount remaining for human consumption would have been 300 tons. An average annual grain consumption of a 7 person present day traditional family is approximately 700 kg,[25] and a rounded-out average of 100 kg per person per year can be adopted for a population of men, women and children. Based on this data, 300 tons of grain supported 3,000 people in ʿUvda Valley.

Averaging the results of the various calculations renders 2,989, and for convenience 3,000. This is the estimated population for eastern ʿUvda Valley during the settlement climax, 3rd millennium BC, in an area of 80 km², which includes the cultivated fields. Considering the harsh desert environment, and our knowledge of desert settlement in general, this population figure is unexpected, indicating a very high density of 37.5 persons per 1 km². It should be noted again that these results were achieved by combining several methods of calculation, and cautious measures not usually applied in demographic studies.

It is interesting to compare the site's density and population of ʿUvda Valley to that of the Negev Highland during the 3rd millennium BC, an area of 1400 km². Counting the dwelling units in published surveys of the area (Cohen, 1981, 1985; Haiman, 1986, 1991, 1993; Lender, 1990; Avni, 1992a; Rosen, 1994) a total of 279 dwelling units is reached, in an area of 800 km², or an average of 0.35 sites per 1 km². Similar calculations by Haiman (1992b:121) rendered 550 sites throughout the Negev Highland, or 0.39 sites per 1 km². Comparatively, the dwelling unit density of ʿUvda Valley was 8.3 times higher.

As to the population size of the Negev Highland during this period, Cohen (1986:229–230) estimated 1,650–2,350 persons. Haiman (1992a:101) suggested the potential number of adults in the Negev Highland sites at 800, and the total population as not more than 1,000. Elsewhere (1992b:121) he suggested 270 adults, or 540 people. Rosen and Finkelstein (1992:50) calculated that over an area of 200 km² in the Sede Boker area, a population of 300 shepherds and farmers could subsist. If this density is applied to the Negev Highland in general, the result will be 2,100 inhabitants. Ignoring the lowest figure, the average would be 1,925 people, and a density of 1.4 persons per 1 km². In ʿUvda Valley, on the other hand, density reached 37.5 persons per km², almost 26.8 times higher.

8.6.4. ESTIMATED HERD SIZE

In an attempt to estimate the herd size in the Negev Highland, Haiman (1992b:121) calculated the area of corrals occupied simultaneously, and multiplied it with a density coefficient of one goat or sheep per 2 m² (after Becker, 1948:219). His total was 8,000 small cattle (sheep and goats) in an area of 1400 km², or 5.7 head per 1 km². A similar density of 5 head per 1 km² was reached by Seligman *et al.* (1962:21–22, 31–33), based on analyses of the vegetational carrying capacity of the Negev Highland. Rosen and Finkelstein (1992:55), preferred a different data base, one goat or sheep per 1 m² (after Epstein, 1985:74), which renders 16,000 head in the same area of corrals. The average between the two, 12,000 small cattle, will be adopted here, or 8.6 head per 1 km².

Following these studies, I will attempt to calculate the herd size and density for the ʿUvda Valley area, although reliability of the results is questionable. The area of the courtyards in 100 dwelling sites was 24,500 m². Since dung deposits were found in only some of the excavated courtyards, we assume that only half served as corrals; and therefore, the estimated corral area in these sites would be 12,250 m^2. To these we should add the animal pens outside the dwelling sites, covering 20,750 m^2, thus the total calculated area of corrals is 33,000 m^2. If only a third of the corrals were used simultaneously (due to grazing cycles), the area occupied by herds at any one time would be *ca.* 11,000 m^2. According to both density coefficients the number of small cattle occupying the corrals was 5,500 or 11,000; the average of the two is 8,250. Assuming that the grazing area of ʿUvda Valley extended from the western margin of the valley to the Arabah Valley in the east, 300 km^2, the density would then be 27.5 head per 1 km^2, 3.2 times higher than in the Negev Highland. Now, if the relation between man and animal is examined in both areas, it would be 1:6.2 in the Negev Highland and 1:2.7 in ʿUvda Valley.

This comparison demonstrates two points. One is that the grazing carrying capacity of ʿUvda Valley was probably 3 times higher than that of the Negev Highland. This could be explained partially by the unique combination of local environmental conditions (see above) and partially by the large supply of stubble and hay from the 1,200 hectares of cultivated fields. Secondly, the ratio of small cattle per person in the Negev Highland was more than double that in ʿUvda Valley. The impression is that the economy of the Negev Highland population was closer to that of true nomads, who specialize in herding and maintain a relation of *ca. 10* small cattle per person (*e.g.* Khazanov, 1984:30). In ʿUvda Valley, however, the relative importance of the herds was lower or, in other words, agriculture was of higher importance. This interpretation is well supported by the ample agricultural finds and installations in ʿUvda Valley, and their rarity in the Negev Highland sites.

If indeed the carrying capacity of the ʿUvda Valley area, for both man and animal, was higher than that of the Negev Highlands, it was in contrast to the present situation. Thus, unavoidably, the question arises as to whether past climate conditions were similar or different from today?

8.7. Paleoclimate

The first attempts to describe a sequence of climate changes in the Near East were made by Huntington (1911, 1924). In his first book, he established a sequence based on contemporary historical and archaeological information, from the 4th millennium BC to the early 20th century. He related times of better climate to periods of political and cultural fluorescence, and times of drought to 'dark ages' and cultural crises. In his second book, he included descriptions of climate changes of the time, and their influences on the people and cultures. Huntington's theory was rejected by most scholars (*e.g.* Reifenberg, 1955:22–23; Albright, 1949:250f; Glueck, 1968:7–12, 209–210, 1970:33–44, 184), and references to his works were few. Also today, several scholars assume that past climate of the Negev was similar to the present time (*e.g.* Finkelstein & Perevolotsky, 1990:80).

Studies on paleoclimate during the last decades, from various fields of research, reveal some discrepancies in results. Therefore, in order to obtain as reliable a picture as possible, a compilation of studies is needed. The paleoclimatic data from the Negev itself is limited. However, with the aid of studies from adjacent desert areas the picture becomes clearer. In view of the large amount of information, only a selection of the studies will be briefly considered here. Most studies presented in the Sede Boker workshop will not be referenced in order to avoid repetition. In most paleoclimate researches radioactive dates are considered as before present (BP) and without calibration. Here, however, dates are all BC and calibrated (following CALIB 3.03, Stuiver and Reimer, 1993).

8.7.1. DATA FROM THE NEGEV AND SINAI, AND THE ADJACENT SEAS

Pollen sections from the Negev Highland sites, although somewhat problematic, indicate wetter climate vegetation than today in the Neolithic and Chalcolithic periods. In the Neolithic site D 1, near Sede Boker, 8% of arboreal pollen was observed, while in two Chalcolithic sites in the same vicinity (D 60 & D 62) 3% arboreal pollen was counted. The species of trees consisted of olive, oak, pine, almond and juniper; all totally absent from present day vegetation (Horowitz in Marks, 1976:66; Horowitz, 1979:248). The fauna in the Neolithic sites included cattle bones *(Bos primigenius)* and a red deer *(Dama mesopotamica)*. These imply the existence of groves along the wadis, and an average annual precipitation of approximately 200 mm, double that of today (Marks, 1976:351–353). In other Neolithic and Chalcolithic sites in the Negev Highland, charcoal samples were identified as European olive and Cypress, both absent from the present natural vegetation (Liphschitz, 1986:85–86). Based on weathering patterns caused by endolithic lichens on stones in ancient Negev sites from various periods, it has been suggested that by the 8th–5th millennia BC species of lichens associated with rainy areas lived in the Negev. From the Chalcolithic period and later, these lichens are no longer found. According to this study, by the 4th millennium BC the climate had stabilized and was similar to the present (Danin, 1985).

Botanical studies from the southern Negev, involving the identification of plant species from 16 charcoal samples from 5th–3rd millennia BC sites in ʿUvda Valley, point to a desertic vegetation, basically similar to the present, with acacia, tamarisk, desert broom and Persian haloxilon. However, a Persian pistacia from site 17 (excavated by Beit Arieh) indicated wetter conditions than today.[26] In PPNB sites in southern Sinai, the pollen of oak and olive was identified (Horowitz in Bar Yosef, 1981b:226). Oak is totally absent in Sinai today; olive only grows under cultivation whereas it was not yet domesticated during the PPNB.

Most 'desert kites' (gazelle traps) in southern Sinai, dated to the 5th–3rd millennia BC, were found in places with extremely poor present vegetation. In the past, however, they must have been built in areas rich enough in vegetation to attract the gazelles (Perevolotzky & Baharav, 1987). In a desert kite excavated in W. Jenah, eastern Sinai, dated to the second half of the 4th millennium BC, wild ass bones were found (Goren & Tchernov, oral communication) while in a burial site near Ain Hudra, a bone implement made of a cow's pelvis was found (Bar Yosef *et al.*, 1977:78). Both animals demand steppe vegetation, richer than is found in the area today.

Geomorphological research in the Negev and Sinai, carried out in archaeological sites and ancient alluvial terraces, identified a few periods of sedimentation, implying a larger amount of rainfall. In the Nahal ᶜIssaron site, ᶜUvda Valley, sedimentation was post PPNB occupation, from the end of the 7th millennium BC and later. In the Siqmim site, northwestern Negev, sedimentation was identified during the Chalcolithic period (Goldberg, 1986; Goldberg & Bar Yosef, 1982, 1990; A. Rosen, 1986). In Nahal Resisim, Negev Highland, sedimentation was dated to Early Bronze I, the second half of the 4th millennium BC (Goldberg & Bar Yosef, 1990:74). The same studies, however, dated to the Neolithic and Chalcolithic periods the accumulations of sand dunes in the Mediterranean coast valley and in the northern Negev, as well as erosion in some wadis. Both phenomena indicate aridity.

A recent study of the Holocene history of the Dead Sea level (Frumkin *et al.*, 1991; Frumkin *et al.*, 1994) revealed that in the 8th millennium BC the Dead Sea water level was 120 m higher than today, but during the 6th millennium BC the level fell. From the 5th millennium to the first half of the 3rd millennium BC the surface rose again to a level 120 m higher than today.[27]

Measurements of oxygen isotopes in land snails from the northern Negev showed a low ratio of ^{18}O between 9000–2000 BC, the result of better climatic vegetation for the snails subsistence. The wettest climate was recorded between 5400 and 4950 BC, when the northern Negev probably received double the amount of present precipitation. According to this research a rise in ^{18}O ratio to the present value occurred only *ca.* 1900 BC (Goodfriend *et al.*, 1986; Goodfriend, 1988, 1991).

8.7.2. DATA FROM OTHER DESERTS OF THE NEAR EAST

Early and late Neolithic sites in the Azraq Basin in eastern Jordan have yielded botanical remains that indicate tree groves along the wadis, including fig, almond and pistacia. Today, no trees grow in this area (Garrard *et al.*, 1994:104–105).

The Rub al Khali, today a virtually sterile 'sand ocean' in the center of the Arabian peninsula, contains remains of ancient fossil lakes. Layers of organic material and sweet water shells from various depths, indicating wet periods, have been dated radiometrically to the 8th–6th millennia BC (McClure, 1976). A Neolithic population lived around the lakes until the end of the 3rd millennium BC and subsisted on herding and agriculture. Finds from these sites included bones of wild ox, antelopes and ostrich egg shells, which testify that savannah vegetation characterized the area (Zarins *et al.*, 1980:19–21, 1981:20; Edens, 1982, with refs.)

Indications of a wetter climate have been observed in other parts of the Arabian Peninsula. Alluvial sedimentation in terraces of Wadi Dawasir, central Arabia, swamp sedimentations in Juba Oasis and Wadi Lahi, and travertin sedimentation in Ein Qanas, al Hesa Oasis, have all been dated to the 8th–5th millennia BC (Garrard *et al.*, 1981; Oates, 1982). In the Tihama Plain, southwestern Saudi Arabia, playa sediments are dated to the 5th–3rd millennia BC, surrounded by Neolithic sites (Zarins & Al-Badr, 1986). On the eastern side of the Peninsula a vast settlement of the Ubeidien culture spread from Mesopotamia, throughout a period of 1000 years, from the mid-5th millennium to the mid-4th millennium BC (Oates, 1982; Zarins *et al.*, 1982). A wetter climate must be the primary explanation for these settlements. A study combining data from several fields of research, as well a series of 14C dates from various sites, also indicates a wetter climate during the 8th–4th millennia BC (Sanlaville, 1992).

Detailed research on the changes in the reproduction rates of small seashells *(Pteropodea)* in the Red Sea sediments has indicated changes in the temperature and salinity of sea water, implying climate changes: a comparatively dry period between 9600 and 8900 BC, a wet period between 8900 and 6600 BC, a comparatively wet period between 6600–5200 BC, a gradual desiccation (but still remaining wetter than today) between 5200 and 3000 BC, and a dry period with maximum desiccation around 2750 BC (*ca.* 500 years earlier than in most studies[28]). Then, a gradual improvement took place towards the present climate (Almogi-Labin *et al.*, 1991).

In the Persian Gulf, clay and argonitic sediments, indicating a wetter climate, were dated around 7000 BC. Microfauna within these sediments was characteristic of a lower salinity than today, as a result of an increased water flow from the Tigris and Euphrates (Luz, 1982). In Abu Dabi and Qatar there are indications of successive overflows of the gulf water, up to 2.5 m higher than today, between 5900 and 2400 BC This was also explained by an increased flow from the rivers (Al Asfour, 1978; Wilkinson, 1978; Luz, 1982). A series of floods and clay sedimentations caused by the rivers were recognized in the ancient towns of the lower Euphrates valley during the 4th–3rd millennia BC (Mallowan, 1964; Raikes, 1966).

In upper Egypt, in presently arid and unpopulated areas, several lakes existed from the late 8th millennium BC to the late 6th millennium BC, though in the second half of the 6th millennium BC the water level receded somewhat (Roberts, 1982). Renewed research in lake remnants of Gilf Kebir, supported by 43 radiometric dates from sediments, showed that the lake reached its maximum during the 5th millennium BC. It was surrounded by comparatively rich semi-desertic fauna and flora, as well as Neolithic sites. In the early 4th millennium BC the sand dune that had dammed the lake was breached by the water and the lake ceased to exist.

Nevertheless, the Neolithic settlement continued until 2900 BC (Schon, 1989:220) or 2,500 BC (Kropelin, 1987). In the Nile Delta two short phases of some decrease in the amount of sedimentation were identified around 6300 and 5400 BC.Between the two, however, a red paleosol had been created, indicating an increase in precipitation in lower Egypt (Butzer, 1975).

Across the Sahara, many Neolithic sites existed from the 8th millennium BC until the end of the 3rd millennium BC, with a settlement climax during the 4th millennium BC. Pollen analyses from Hojar Heights, in the central Sahara, a very arid area today, have shown a mixed Mediterranean and tropical vegetation between 6000–2300 BC (Camps, 1975; Hays, 1975, 1992; Aliman, 1982). In the fossil lakes of the Sahara a wet phase has been dated at between 7500–5900 BC (Street & Grove, 1979; Nicholson & Flohn, 1980). Pollen analysis of the fossil lakes in Sudan pointed to a northward shift of the monsoon belt by 500 km from its present location between 10000 and 3000 BC (Kutzbach & Street Perrott, 1985). In the Libyan desert wet periods were determined by ^{18}O rates and ^{14}C dates of underground water bodies, around 7400 BC and between 6500 and 3200 BC. A short, comparatively dry phase, was dated *ca.* 5500 BC. Two peaks of wetness were identified around 7400 and 3800 BC, while since *ca.* 2000 BC the climate remained as dry as today (Edmunds & Wright, 1979). Many rock drawings dated to the 'pre-pastoral Neolithic' (8th–7th millennia BC), present tropical fauna (Lohte, 1958, 1961; Hays, 1975; Muzzolini, 1994, 1995, Ch. 1). The general picture from the Sahara studies indicate that the wet period lasted from the 8th millennium BC to the second half of the 3rd millennium BC. A comparatively dry period occurred in part of the 6th millennium BC, though still wetter than nowadays.

Support for the Sahara studies was found in a research on oxygen isotope rates, ^{14}C dates of fauna, and dates of sapropelitic sediments, from 47 cores drilled in the bottom of the Mediterranean, (Fontugne *et al.*, 1994). This research points to an increased flow of sweet water and/or temperature decrease in the Mediterranean water between 8000–2500 BC, while the wettest period was between 8000–6800 BC. The research proposes that one or more rivers existed in Libya and contributed sweet water to the Mediterranean. It also points to a moderate rise of ^{18}O rates in shells during the 4th–3rd millennia BC, *i.e.* a gradual reduction in the flow of sweet water and/or increase in water temperature over this time span.

As stated above, the study results quoted here from the Negev and Sinai, as well as from other desert areas of the Near East, do not present a full consensus. This is partly due to difficulties in precise and uniform dating. Nevertheless, a dominant line does emerge. From the 8th to the end of the 4th millennium BC the prevailing climate in the Near Eastern deserts was somewhat wetter than today, with higher precipitation, but still mostly desertic. Several studies suggest an annual average temperature and humidity higher than today, with a lower rate of evaporation. Between 6000–4300 BC there were comparatively dry periods, but these were still wetter than today.

It should be remembered that in a desert environment, even minor changes in rain and evaporation rates would have a significant influence on the carrying capacity. Several studies have specifically proposed a northward shift of the monsoon belt as a possible factor in the climate change during the Neolithic and Chalcolithic periods (Gat & Magaritz, 1980; Roberts, 1982; Kutzbach & Street-Perrott, 1985; Goldberg & Rosen, 1987). Another theory suggests that in the Arabian Peninsula the south eastern monsoon shifted northward through Nefud and northern Hejaz, some 600 km east of Eilat area (Zarins, 1992 with refs.). This means that the Negev and Sinai may have occasionally been influenced by both the Indian and African monsoons. Still another theory suggests local summer rains across the Near East during the first half of the Holocene, until the early 6th millennium BC (El Moslimani, 1994). In either case, a single summer rain could have tremendously improved the carrying capacity and living conditions in the desert.

During the 3rd millennium BC most indications of a better climate disappeared, and it seems that a gradual desiccation occurred. The Dead Sea level decreased sharply, reaching its minimum in the early 2nd millennium BC (Frumkin *et al.*, 1991; Frumkin *et al.*, 1994). In Tel-Aviv excavations, an accumulation of sand was observed on top of Early Bronze remains (Ritter-Kaplan, 1984) also indicating desiccation. In the Nile Valley, the sedimentation rate significantly decreased (Butzer, 1975); and on the Hojar Heights, the tropical and Mediterranean elements disappeared from the pollen spectra (Hays, 1975; Camps, 1975; Aliman, 1982). In the northern Near East, lake levels decreased (Roberts, 1982) and in north Africa lakes began to dry around 2500 BC (Street & Grove, 1979; Gasse, 1980). In the Euphrates Valley and the Persian Gulf, no indications were found for the increased flow of the rivers (Al Asfour, 1978; Luz, 1982). The accumulation of data points to a severe climatic crisis towards the end of the 3rd millennium BC, probably one of the factors for drastic political and socio-economical changes effecting large parts of the Near East (Crown, 1971; Bell, 1971; Yakar, 1976). Yet it is interesting to note that the desert population (at least in the Negev and Sinai) reached its climax during the 3rd millennium BC, simultaneously with the gradual desiccation.

Shortly thereafter, the climate probably improved. On the Israel Coastal Plain and in the Jordan Valley, sweet water marshes caused clay sedimentation dated to approximately 2000 BC (Neev, 1980). In upper Egypt, a new but short stage of flooding of the dried lakes occurred (Roberts, 1982), while the first Dark Age of Egypt lasted about 15 years, at the end of Dynasty VI (Bell, 1971). In Qatar, a new rise of Gulf level was identified, dated around 2500 BC (Al Asfour, 1978). These data imply that the climatic crisis at the end of the 3rd millennium was short-lived, and was followed by a stabilized climate regime similar to the present one.

Relating now to the archaeological remains in ʿUvda Valley, in light of the Paleoclimate and of recent Bedouin agriculture, we may conclude that a broad agricultural settlement could have indeed existed here in the past.

8.8. Summary

The common image of human history in the desert was, until recently, very ephemeral. Desert sites were considered to be 'short lived', a 'passing phenomenon', or 'inhabited for only a few years' (see above). The desert history was described in terms of 'ups and downs', of short periods of settlement with long gaps in-between.[29] Moreover, periods of settlement have usually been described as the result of outside initiatives of a political power penetrating the desert from the sown.[30] Rarity of remains from periods of gaps was explained by assumptions that desert nomads left little or no archaeological remains.[31] Similar contentions still persist even in recent studies.[32]

This portrayal of the Negev and Sinai in general has already received serious criticism (Avner *et al.*, 1994). Now, at least for ʿUvda Valley, the evidence presented reflects a totally different picture. Many sites were permanent settlements, some continued relatively uninterrupted during hundreds and even thousands of years. The settlements experienced a constant and complex cultural and economical evolution, from hunting and gathering to desert agriculture and grazing. As a result, the three early Neolithic habitation sites (8th–7th millennia BC) turned to 154 habitation sites of the Late Neolithic, Chalcolithic and Early Bronze (6th–3rd millennia BC) and the population increased more than 50 times. The growth and duration of settlement demonstrates both the efficient and sustainable use of the environment by the population.

The existence of an agricultural society in such an extreme desertic area raises unavoidable questions about the paleoclimate. The growing accumulation of studies indicates that during the time span discussed here the environment was desertic, but somewhat milder than today, probably with higher precipitation and lower fluctuation. If indeed the monsoonal trajectories penetrated the area, even only once a year, then it would have provided a possible explanation for the fact that the population density in ʿUvda Valley was more then 20 times higher than in the Negev Highland, despite the present clear climatic advantages of the latter.

Most surprising is that the population of the valley reached its climax during the 3rd millennium BC, simultaneously with the onset of gradual desiccation. This phenomenon was probably related to the overall social and historical development of the Near East during the Early Bronze. However, it could not have occurred without the firm cultural base of the population, and the desert 'law of living' previously developed. We may assume that these skills enabled the population to formulate solutions for the increasing aridity. Nevertheless, the climatic crisis of the late 3rd millennium BC did demand radical adjustments in the way of life. The material culture of the Early Bronze IV, which continued in the Early Bronze I–III, was characterized by fewer agricultural tools and more indications of grazing and mobility. This meant a lower standard of living but enabled the population to endure the crisis. The limited agricultural evidence found in ʿUvda Valley, and most of the Negev and Sinai in general, during the following Middle and Late Bronze Age periods remains an issue demanding further research.

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Notes

1. Many thanks to the officers of the U.S. Army Corp. of Engineers: Gen. Morris, Gen. Noah, Gen. Lewis, Col. Miller and Col. Kelly for their consideration and support; and to the officers of the Israel Air Force: Brig. Gen. Bartov, Col. Tal, Major Zamir and many others.
2. All dates in this paper are calibrated based on Calib. 3.0, Stuiver & Reimer 1993.
3. The excavators of N. ʿIssaron pointed out the high percentages of *Capra* as a possible indication of an attempt to domesticate these animals, but this is not easily accepted. In the Neolithic site of Wadi Tbeik, southern Sinai, no sign of domestication was found (Tchernov & Bar Yosef 1982) and in ʿUjrat el Mehed, southern Sinai, a high percentage of Ibex bones, one year old or older, were only interpreted as an incipient attempt to domesticate Ibexes, which could not be continued (Dayan *et al,* 1986). In any case, the high percentage of arrowheads in N. ʿIssaron obviously indicates the importance of hunting in the economy of the population.
4. In the first publication (Goring-Morris & Gopher, 1983) the excavators did not mention the open space, but did describe an open courtyard surrounded by rooms in a later one (Goring-Morris & Gopher, 1987:18). In personal communication they preferred to term it as an open space, since built walls were not preserved, probably due to reuse in Stratum A, dated to the Chalcolithic and Early Bronze.
5. In the eastern desert of Jordan, in Wadi Jilat and Wadi Ruweshid, the emergence of herding was dated to the early 6th millennium BC (McCartney, 1992; Martin *et al.*, in Garrard *et al.,* 1994). In the Duwela site, the same area, the earliest known built courtyard was dated by 14C to the early 6th millennium BC (Betts, 1988). In the desert fringe of Jordan, in Ein Ghazal (Rollefson, 1992) and in Beidha and Basta (Baker, in Nissen *et al.,* 1991:29–32), the first occurrence of domesticated sheep and goats was dated as early as the late 7th millennium BC (PPNC).
6. Sites of the third type are somewhat similar to several large sites in the Negev Highland, considered as central settlements: Ein Ziq, Masha'bei Sadeh, Be'er Resisim and N. Nizzana (Cohen, 1986:70–82,86–96,253; Cohen & Dever, 1981; Haiman, 1996), and W. Fuqia in central Sinai (Rothenberg, 1979;117–119). The main difference between the large sites in ʿUvda Valley and those in the Negev and Sinai is their size, the latter are much larger, up to 2 ha. Comparison of sites from the different areas may present chronological difficulties. The site in Sinai was dated by Rothenberg to the Chalcolithic period, while those of the Negev Highland were dated by

Cohen & Dever to the Early Bronze IV. Early Bronze II sherds were found only on the bottom of the Negev Highland sites. Possibly, this type of site actually existed for a long period of time. According to Haiman (1996) the central settlements served mainly for copper trade.

7. This percentage is very high in comparison with those of contemporary sites in the Negev Highland, that enjoyed a better climate: 0% in Kvish Harif (Rosen, 1984a), 0% in N. Horsha (Rosen, 1991:170-173), 0.5% in Ramat Matred (Haiman, 1994:30), 3.7% in N. Mitnan (Rosen, 1993:63). Even in the Beer-Sheva Valley sickle blade percentages are lower than in ʿUvda Valley, 2.3% in Shikmim (Levy & Rosen, 1987:288, 291), 2.5% in Arad (Schik, pers. comm.). A very high percentage of sickle blades, 24%, was found in the Chalcolithic site of N. Grar which was considered to be exceptional (Gilead, 1995:146,272,278). The sickle blade percentage in ʿUvda Valley is similar to that in Bab Adh Dhra', 9.6%, which enjoyed good water supplies from W. Kerak (McConaughy in Rast & Schaub 1980:53–55) and is even similar to those in fertile areas in the north, such as Ein Shadud in the Israel Valley—9.8% (Rosen in Braun, 1985:155,166). Indeed, the percentage of sickle blades in the ʿUvda Valley sites is probably higher than published, due to identification of non-cereal vegetation wear found on part of the sickle blades (Bueller, 1988:30).

8. Tools similar or identical to the Canaanean blades were found in a series of Calcolithic sites: Giv'atim (Sussman & Ben Arieh, 1966, Fig. 6), Azor (Perrot, 1961, Fig. 43), Nahal Grar (Gilead, 1995:257), Gilat (Rowan & Levy, 1994), Serabit el Khadem (Beit Arieh, 1980, Fig. 8), Tel Magass (Khalil, 1992:143) and Maadi in Egypt (Rizkana & Seeher, 1988 II, Pl. 76:1–7). In Gilat they were termed 'Proto-Canaanean'. Canaanean blades were even found in two Pottery Neolithic sites: Ashkelon (Perrot & Gopher, 1996:156) and Jillat 13, eastern Jordan (Garrard *et al.*, 1994:86-87).

9. I thank Amnon Asaf, curator of Macayan Baruch Museum, for showing me these blades.

10. The grinding stones were examined by T. Weissbrod, Israel Geological Survey, Jerusalem, who concluded that their raw material was exceptionally dense and hard. In his opinion, in order to locate such sandstone, a very intimate knowledge of the area was required. It is possible this kind of knowledge was acquired during copper mining in Timna and other Arabah Valley sites, where mines penetrated the sandstone.

11. Pollen analysis did not show results, while results of phytolits analysis are forthcoming.

12. I am grateful to David Eitam, curator of the Oil Museum, Haifa, for his comments on these installations after examination in April 1992. In his opinion, it is highly possible they were used for oil production, although close parallels are not known.

13. *E.g.* Tuleilat Ghassul (Koppel, 1940,Pl. 13), Arad (Amiran *et al.*, 1978:28,32,33,37), Kh. et-Tel. (Callaway, 1980:189) and Tel Dalit (Gophna, 1997, Fig. 21). On the functioning and efficiency of stone-paved and lined granaries in the early Iron Age, see Currid & Navon, 1989).

14. *Cf.* a pair of grinding stones in the granary at Arad (Amiran *et al.*, 1978:33, Pl. 155) and fertility female figurines in granaries in Çatal Hüyük in Anitolia (Mellaart, 1967:183).

15. Additional threshing floors were published during the last years from the Negev Highland (Haiman, 1986:53, 156, 213; 1991:37, 47, 80, 84, 97, 100, 101, 109, 118, 120, 125, 133; 1993:28, 46, 63, 65, 87, 91, 97, 106; Lender, 1990:105, 114, 115, 172, 191, 197, 202; Avni, 1992a:49, 57, 94, 126, 127). However, also here, according to my own observation, the numbers of threshing floors are higher than published. In several cases, threshing floors were described as 'circular structures' or 'corrals' (*e.g.* Lender, 1990:165; Avni, 1992a:41,50). In the western Negev Highland, in the northwestern Negev and the Beer-Sheva Valley, I examined *about* 40 threshing floors. All showed indications of both recent Bedouin and ancient use. It was found that all examined 'Bedouin' threshing floors were actually old, constructed several millennia ago.

16. Hornell, 1930; Crawford, 1935; Bordez, 1969; Whallon, 1978; Cheetham, 1982; Pearlman, 1984, 1987, Fox and Pearlman, 1987. In 1990 Prof. A. Ronen, Haifa Univ., documented on video a craftsman in Nicosia preparing this type of sledge blades for sale. I am grateful to Ronen for showing me the film. In Cyprus in 1993, I was told by Pearlman that the craftsman had died.

17. I am grateful to David Amit and Hanan Eshel who called my attention to and showed me this sledge.

18. I am grateful to Igor Mindel, Jewish Agency, who provided this information, Mindel tested the soil during 1983–1986.

19. The total drainage area of N. Hayun, including ʿUvda Valley, is 1,116 km^2, and the annual average flood water volume in the northern end reaches 1,000,150 m^3. This is the second largest amount of flood water in the Negev, after N. Paran, with 2,005,000 m^3 (Finkel & Finkel, 1979:134).

20. Repeated cultivation of fields without fallowing or overuse of the soil is described by Marx (1988:90) in connection with Bedouin agriculture in the Beer-Sheva area, in spite that in this case it was not flood agriculture and did not enjoyed annual refertilization.

21. The springs were first noticed by N. Minkowski, from the Ma'aleh Shaharut settlement. Shortly thereafter, the flow measurement was taken by A. Greenberg from the Agricultural Research Center in Yotvata.

22. *Cf.* Dalman, 1933:216–217; Marx, 1967:83,94. In my own observations in the Negev and Sinai during the last 20 years, threshing season lasts until the end of August.

23. Frankfort, 1950; McAdams and Nissen, 1972; Rosenan, in Amiran *et al.,* 1978; Marfoe, 1980; Shiloh, 1980; Van Beek, 1982; Broshi and Gophna, 1984,1986; Gophna and Portugali, 1988.

24. The traditional agriculture in the Beer-Sheva area reached yields of 200–1,000 kg wheat per hectare and slightly higher yields of barley (Ben-David, 1988:46–47). In this area, the agriculture was based on direct rain irrigation, which is less productive than flood irrigation. In the traditional agriculture in the Petra area, which generally enjoys goods amounts of rain, the wheat yields are similar to the Beer-Sheva area, while barley yields are between 200–2,000 kg per hectare (Russell, 1995:697).

25. Dar, 1982:327—500–700 kg for 7-person Arab family in Samaria; Ben-David, 1982:180—800 kg per 6-person Bedouin family, when grains were subsidized.

    In several studies higher values of grain consumption were adopted, 180–230 kg per person (*e.g.* Broshi, 1986:42, with refs.). In my opinion, these figures are too high, for the following reasons: 1. These figures usually consider adult consumption, while more realistic calculations should take the average between adults and children. 2. A baked bread contains approximately 50% water. 200 kg of grain equates to 400 kg of bread, *i.e.* 571 present day loaves per person per year. This means 1.5 loaves per day, or 2,600 cal., almost the total calorie consumption for a working person per day. Even in a society that subsists on a limited food basket, there is no justification for such high bread consumption, certainly not for a society which subsists on a combined economy, such as existed in ʿUvda Valley.

26. The identification of the species from ʿUvda Valley were made by U. Baruch and E. Verker (unpublished) and by Liphschitz (1986:87).

27. In this point there is some discrepancy between the two publications. The graphs (Figs. 7 & 8) in both show a decrease and desiccation beginning at 3000 BC. The text in the first article (p. 198) relates the beginning of decrease to approximately 2750 BC, while in the second article the beginning of decrease is dated to 2500 BC, all in calibrated dates. Therefore, it would be correct to say that the desiccation, which caused the decrease in the Dead Sea level, began by the first half of the 3rd millennium BC without more precise dating in this stage of the research.

28. This difference may be due to the fact that dating marine material demands a marine calibration, in addition to the terrestrial one. Marine calibration of open oceans reduce between 0–1% of the common calibrated result, depending on the amount of ancient carbon dissolved in the water (*e.g.* Stuiver and Reimer, 1993; Biagi, 1994; Dye, 1994). Foraminifera demands an even higher percentage reduction, according to its size (Thomson *et al.*, 1995).
29. Glueck, 1935, 1961, 1968, 1970; Reifenberg, 1955; Rothenberg and Cohen, 1968; Evenari *et al.*, 1971; Baron, 1981.
30. Rothenberg, 1970:21 *etc*; Amiran *et al.*, 1973; Beit Arieh, 1974, 1981, 1983; Baron, 1981; Cohen, 1981, 1985, 1986; Rosen, 1987b, 1988, 1992; Haiman, 1988, 1989, 1992a; Finkelstein, 1988.
31. Glueck, 1935:183; 1968:11-12, 127; 1970:11-12, 65; Rothenberg, 1970:22; Cohen, 1986:433; Finkelstein, 1984:198; Finkelstein and Perevolotsky, 1990:67-68, 77.
32. *E.g.* Cohen, 1985, 1986, 1988; Haiman, 1986, 1991, 1992a, 1993; Rosen, 1987b, 1994; Lender, 1990; Avni, 1992a.

Abbreviations

ASOR	American Schools of Oriental Research
AASOR	Annual of the American Schools of Oriental Research
ADAJ	Annual of the Department of Antiquities of Jordan
AJA	American Journal of Archaeology
'Atiqot	Atiqot: Journal of the Israel Antiquities Authority
ATLAL	Journal of Saudi Arabian Archaeology
BA	Biblical Archaeologist
BAR	Biblical Archaeology Review
BASOR	Bulletin of the American Schools of Archaeological Research
CA	Current Anthropology
ESI	Excavations and Surveys in Israel
IEJ	Israel Exploration Journal
IMJ	Israel Museum Journal
JNES	Journal of Near Eastern Studies
JWP	Journal of World Prehistory
PEQ	Palestine Exploration Quarterly
TA	Tel-Aviv
ZDPV	Zeitschrift des Deutschen Palastina-Vereins

References

Adams, R. (1975) An ancient Uruk threshing sledge or harrow. *Sumer* 31, 17–20.

Al ʿAref, A. (1936) *The Beduin Tribes in the Be'er Sheva District* (Hebrew translation by M. Capeleuk). Bustenai, Tel-Aviv.

Al Asfour, T. (1978) The marine terraces of the Bay of Kuwait. *In* W.C. Brice (ed.), *The Environmental History of the Near and Middle East Since the Last Ice Age*, Academic Press, London, pp. 245–254.

Albright, W.F. (1949) *The Archaeology of Palestine*. Pelican Books, Middlesex.

Aliman, M. H. (1982) Le Sahara: Grande zone desertic Nord-Africaine. *In* T.L. Smiley (ed.), *The Geological Story of the World's Desert, Striae 17*, Societas Upsaliensis pro Geologia Quaternaria, Uppsala, pp. 35–51.

Almogi-Labin, A., Hemleben, C., Meischner D., and Erenkeuser, H. (1991) Paleoenvironmental events during the last 13,000 years in the central Red Sea as recorded by Pteropoda. *Paleoceanography* **6**, 83–98.

Alon, D. (1988) The extent and history of the Chalcolithic settlement in the southern Shefela. *In* D. Bauman and E. Stearn (eds.), *Men and Environment in the Southern Shefela, Israel,* Masada, Giv'ataim, pp. 84–89 (in Hebrew).

Amiran. R., Arnon, C. and Avner, U. (1980) Biq'at ʿUvda site 124. *I.E.J.* **29**, 256.

Amiran, R., Beit-Arieh, I. and Glass, J. (1973) The interrelationship between Arad and sites in southern Sinai in the early Bronze Age II. *IEJ* 23, 193–197.

Amiran, R., Paran, U., Shiloh, Y., Brown, R., Tsafrir Y., and Ben Tor, A. (1978) *Early Arad I: The Chalcolithic Settlement and Early Bronze City, I: First-Fifth Seasons of Excavations, 1962–1966.* Israel Exploration Society, Jerusalem.

Avitsur, S. (1965) *The Native Ard of Eretz Israel, Its History and Development.* Adam v' Amalo, Tel-Aviv (Hebrew).

——. (1966) *Implements for Harvesting and Similar Purposes Used in the Traditional Agriculture of Eretz Israel.* Adam v' Amalo, Tel-Aviv (Hebrew).

——. (1984) *The Threshing Floor.* Adam v' Amalo, Tel-Aviv (Hebrew).

Avner, U. (1979) A survey in the ʿUvda Valley. *Hadashot Archeologiot* 69–71, 1–2 (Hebrew).

Avner, U. (1981) Threshing floors near Borot Lotz. *Hadashot Archeologiot* **76**, 38 (Hebrew).

——. (1982a) Biq'at ʿUvda Survey. *ESI* 1, 79–81.

——. (1982b) Survey of Biq'at' ʿUvda Eilat Road. *ESI* 1, 81–82.

——. (1982c) Excavation of a desert kite near Kibbutz Samar. *ESI* 1, 103–104.

——. (1982d) Excavation of a threshing floor in Biq'at ʿUvda. *ESI* 1, 80–81.

——. (1983) Excavation of cult sites in Biq'at ʿUvda. *ESI* 2, 14–15.

——. (1984) Ancient cult sites in the Negev and Sinai Deserts. *TA* **11**, 115–131.

——. (1986) Excavation of tombs in a rock shelter near Ma'ale Shaharut. *ESI* 5, 62–63.

——. (1987) Desert kites. *In* B. Shalmon (ed.), *The Gazelles in Israel,* The Society for the Protection of Nature, Tel-Aviv, pp. 20–27 (Hebrew).

——. (1989a) Ma'ale Shaharut, survey. *ESI* **7/8**, 114–116.

——. (1989b) Mispe Sayyarim area, survey. *ESI* **7/8**, 134–135.

——. (1990a) Excavation of Tumuli in Eilat. *ESI* 9, 76–78.

——. (1990b) Ancient agricultural settlement and religion in the ʿUvda Valley in southern Israel. *BA* 53, 125–141.

——. (1991) A note on a late Neolithic burial site in Eilat. *AJA* **95**, 496–497.

——. (1993) Masseboth sites in the Negev and Sinai and their significance. *In* J. Aviram (ed.), *Biblical Archaeology Today, 1990. Proceedings of the Second International Congress on Biblical Archaeology,* Israel Exploration Society, Jerusalem, pp. 166–181.

——. (1997) Nahal Paran Survey. *ESI* **16**, 132–133.

——. (in print) *Late Neolithic-Early Chalcolithic Burial Site in Eilat, by the Red Sea.* 'Atiqot Monographies Series, Jerusalem.

Avner, U., Carmi, I. and Segal, D. (1994) Neolithic to Bronze Age settlement of the Negev and Sinai in light of radiocabon dating: A view from the southern Negev. *In* O. Bar-Yosef and R. Kra (eds.), pp. 265–300.

Avni, G. (1992a) *Map of Har Saggi Northeast (225).* Israel Antiquities Authority, Jerusalem.

——. (1992b) Survey of deserted Bedouin campsites in the Negev highlands and its implications for archaeological research. *In* O. Bar Yosef and A. Khazanov (eds.), pp. 241–254.

Baron, A. (1981) Adaptive strategies in the archaeology of the Negev. *BASOR* 242, 51–81.

Baruch, U. (1986) The Late Holocene vegetational history of Lake Kinneret (Sea of Galilee), Israel. *Paleorient* 12/2, 37–48.

Bar-Yosef, O. (1981a) The 'Pre-Pottery Neolithic' period in the southern Levant. *In* J. Cauvin and P. Sanlanville (eds.), *Prehistoire du Levant,* Editions du Centre National de la Recherch Scientifique, Paris, pp. 555–569.

——. (1981b) Neolithic sites in Sinai. *In* W. Frey and H.P. Uerpmann (eds.), *Contribution to the Environmental History of Southwest Asia,* Beihette zum Tubinger Atlas des Vorderen Orientes, Reihe A, No. 8, pp. 217–235.

——. (1984) Seasonality among Neolithic hunter-gatherers in southern Sinai. *In* J. Clutton-Brock and C. Grigson (eds.), *Animals and Archaeology: Early Herders and Their Flocks,* BAR International Series 202, Oxford, pp. 145–160.

Bar-Yosef, O. and Alon, D. (1988) Nahal Hemar cave: The excavations. *'Atiqot* 18, 1–30.

Bar-Yosef, O., Belfer, A., Goren, A., and Smith, P. (1977) The Nawamis near 'Ein Huderah. *IEJ* 27, 65–88.

Bar Yosef. O. and Khazanov, A.,eds. (1992) *Pastoralism in the Levant.* Prehistory Press, Madison.

Bar Yosef, O. and Kra, R.S., eds. (1994) *Late Quaternary Chronology and Paleoclimates of the Eastern Mediterranean.* University of Arizona, Tucson,

Becker, D. (1948) *Raising Cattle.* HaKibbutz HaMeuchad, Ein Harod (Hebrew).

Beit-Arieh, I. (1974) An early Bronze Age II site at Nabi Salah in southern Sinai. *TA 1*, 144–156.

——. (1980) A Chalcolithic site near Serabit El-Khadim. *TA* 7, 45–64.

——. (1981) A pattern of settlement in southern Sinai and southern Canaan in the Third Millennium BC. *BASOR* 243, 31–55.

——. (1982) An early Bronze Age II site near the Feiran oasis in Southern Sinai. *TA* 9, 146–156.

——. (1983) Central southern Sinai in the early Bronze Age II and its relations with Palestine. *Levant* 16, 39–48.

——. (1986) Two cultures in southern Sinai in the third millennium BC. *BASOR* 263, 27–54.

——. (1989) The early Bronze Age pattern of settlement in the Sinai. *In* P. de Miroschedji (ed.), *L'Urbanisation de la Palestine a l'Age du Bronze Ancien. Actes du Colloque d'Emmaus October, 1986,* BAR International Series 527, Oxford, pp. 189–197.

Bell, Barbara. (1971) The Dark Ages in ancient history I. The first Dark Age in Egypt. *AJA* 75, 1–25.

Ben-David, J. (1982) Stages in the Sedentarization of the Negev Bedouin: A Transition from Farmer Semi Nomadic to Settled Population. Unpublished Ph.D. dissertation, Hebrew University, Jerusalem (Hebrew).

——. (1988) Agricultural Settlements for the Beduin Population—Policy Proposal. Jerusalem Institute for Research of Israel, Jerusalem (Hebrew).

Benoit, P., Milik, J.T. and de Vaux, R. (1961) *Les Grottes de Muraba'at, Discoveries in the Judean Desert ii.* Clarendon Press, Oxford.

Betts, A. (1988) 1986 Excavations at Dhuweila, Eastern Jordan: A Preliminary Report. *Levant* 20, 7–21.

Biagi P. (1994) A radiocarbon chronology for the aceramic shell-middens of coastal Oman. *Arabian Archaeology and Epigraphy* 5, 17–31.

Biger G. and Grossman, D. (1993) Village and town population in Palestine during the 1930s–1940s and their relevance to ethnoarchaeology. *In* A. Biran and J. Aviram (eds.), pp. 19–30.

Bordaz, J. (1969) Flint flaking in Turkey. *Natural History* 78, 73–79.

Braslavi, J. (1952) *Knowing the Country,* Vol. IV, *Eilat and the Red Sea.* HaKibbutz HaMeuchad, Tel-Aviv (Hebrew).

Broshi, M. (1986) The diet of Palestine in the Roman period—Introductory notes. *IMJ* 5, 41–56.

Broshi, M. and Gophna, R. (1984) The settlements and population of Palestine during the early Bronze Age II–III. *BASOR* 253, 41–53.

——. (1986) Middle Bronze Age II Palestine: Its settlements and population. *BASOR* 261, 73–90.

Bueller, J. (1988) Handling, hafting and ochre stains. *In* S. Beyries (ed.), *Industries Lithiques, Traceologie et Technologie*, Vol. I, Bar International Series 411 (i), Oxford, pp. 5–32.

Butzer, K. W. (1975) Patterns of environmental change in the Near East during late Plaistocene and early Holocene time. *In* F. Wendorf and A. Marks (eds.), *Problems in Prehistory: North Africa and the Levant*, Southern Methodist University Press, Dallas, pp. 389–410.

Callaway, J.A. (1980) *The Early Bronze Age Citadel and Lower City at Ai (Et Tell)*. ASOR, Cambridge.

Camps, C. (1975) The prehistoric cultures of North Africa radiocarbon chronology. *In* F. Wendorf and A. Marks (eds.), *Problems in Prehistory: North Africa and the Levant*, Southern Methodist University Press, Dallas, pp. 181–204.

Carmi, I. (1987) Rehovot radiocarbon measurements III. *Radiocarbon* **29**, 100–114.

Carmi, I., Segal, D., Goring-Morris, A.N., and Gopher, A. (1994) Dating the prehistoric site Nahal Issaron in the southern Negev, Israel. *Radiocarbon* **36**, 391–398.

Cheetham, L. (1982) Threshing and einnowing—An ethnographic study. *Antiquity* **56**, 127–130.

Cohen, R. (1981) *Archaeological Survey of Israel: Map of Sede Boqer—East (168)*. Archaeological Survey of Israel, Jerusalem (Hebrew).

——. (1985) *Archaeological Survey of Israel: Map of Sede Boqer—West (167). Archaeological Survey of Israel*. Jerusalem (Hebrew).

——. (1986) The Settlement of the Central Negev in the Light of Archaeology and Literary Sources During the 4th–1st Millennia BCE. Unpublished Ph.D. dissertation, Hebrew University, Jerusalem (in Hebrew with English summary).

——. (1987) Excavations at Moa, 1981–1985. *Qadmoniot* 26–31.

——. (1988) Settlement in the Negev highlands from the fourth millennium BCE to the fourth century BCE. *Qadmoniot* **22**, 62–81 (Hebrew).

Cohen, R. and Dever, W.G. (1981) Preliminary report of the third and final season of the 'Central Negev Highlands Project'. *BASOR* **243**, 57–77.

Crawford, O.G.S. (1935) A primitive threshing machine. *Antiquity* **9**, 335–339.

Crown, Alan, D. (1971) Climatic change, ecology and migration. *Australian Journal of Biblical Archaeology* **1**, 3–22.

Currid, J.D. and Navon, A. (1989) Iron Age pits and the Lahav (Tell Halif) Grain Storage Project. *BASOR* **273**, 67–78.

Dalman, G. (1933) *Arbeit und Sitte in Palastina, Band III*. Bertelsmann, Gutersloh.

Danin, A. (1985) Palaeoclimates in Israel: Evidence from Weathering Patterns of Stones in and Near Archaeological Sites. *BASOR* **259**, 33–43.

Dar, S. (1982) *The Settlement Extent in Western Sameria During the Roman and Byzantine Periods*. Society for the Protection of Nature, Tel-Aviv (Hebrew).

Dayan, T., Tchernov, E., Bar-Yosef, O. and Yom-Tov, Y. (1986) Animal exploitation in Ujrat El-Mehed, a Neolithic site in southern Sinai. *Paleorient* **12**, 105–116.

Dye, T. (1994) Apparent ages of Marin shells: Implications for archaeological dating in Hawaii. *Radiocarbon* **36**, 51–57.

Edens, C. (1982) Towards a defintion of the Western ar-Rub' al-Khali 'Neolithic'. *ATLAL* **6**, 109–124.

Edmunds, W.M. and Wright, E.P. (1979) Groundwater recharge and Palaeoclimate in the Sirte and Kufra Basins, Libya. *Journal of Hydrology* **40**, 215–241.

Eldar, I., Nir, Y. and Nahlieli, D. (1992) The Bedouin and their campsites in the Dimona region of the Negev: Comparative model for the study of ancient desert settlements. *In* O. Bar Yosef and A. Khazanov (eds.), pp. 205-217.

El-Moslimany, A.P. (1994) Evidence of early Holocene summer precipitation in the Continental Middle East. *In* O. Bar Yosef and R.S. Kra (eds.), pp. 121–130.

Epstein, C. (1978) A new aspect of Chalcoliothic culture. *BASOR* **229**, 27–45.

——. (1993) Oil production in the Golan Heights during the Chalcolithic Period. *TA* **20**, 133–146.

Epstein, H. (1985) The Awassi Sheep. F.A.O. *Animal Production and Health Paper* 57. Food and Agricultural Organization of the United Nations, Rome.

Evenari, M., Shanan, L., and Tadmor, N. (1971) *The Negev.* Harvard University Press, Cambridge.

Finkel, H. and Finkel, M. (1979) Water runoff in the Arabah. *In* A. Shmueli and J. Gradus (eds.), *The Land of the Negev,* Tel-Aviv, pp. 125–139 (Hebrew).

Finkelstein, I. (1984) The Iron Age 'Fortresses' of the Negev highlands: sendentarization of the nomads. *TA* 11, 189–209.

——. (1988) Arabian trade and socio-political conditions in the Negev in the twelfth- eleventh centuries BCE. *JNES* 47, 241–252.

Finkelstein, I. and Perevolotsky, A. (1990) Processes of sedentarization and nomadization in the history of Sinai and the Negev. *BASOR* 279, 67–88.

Fontugne, M., Arnold, M., Labeyrie, L., Paterne, M., Calvert S.E., and Duplessy, J.-C. (1994) Paleoenvironment, Sapropel chronology and Nile River discharge during the last 20,000 years as indicated by deep-sea sediment records in the Eastern Mediterranean. *In* O. Bar Yosef and R.S. Kra (eds.), pp. 75–88.

Fox, W.A. and Pearlman, D.A. (1987) Threshing sledge production in the Paphos district, Cyprus. *Sima* 77, 227–231.

Frank, F. (1934) Aus der 'Araba. *ZDPV* 57, 191–280.

Frankfort, H. (1950) Town planning in ancient Mesopotamia. *Town Planning Review* 21, 98–115.

Frumkin, A., Carmi, I., Zak, I. and Magaritz, M. (1994) Middle Holocene environmental change determined from the Salt Caves of Mount Sedom, Israel. *In* O. Bar Yosef and R.S. Kra (eds.), pp. 315–332.

Frumkin, A., Magaritz, M., Carmi, I., and Zak, I. (1991) The Holocene climatic record of the Salt Caves of Mount Sedom, Israel. *Holocene* 1, 191–200.

Galili, E. and Nir, Y. (1993) The submerged pre-pottery Neolithic water well of Atlit-Yam, northern Israel, and its palaeoenvironmental implications. *Holocene* 3, 265–270.

Galili, E. and Sharvit, J. (1994–5) Evidence of olive oil production from the submerged site at Kfar Samir, Israel. *Mitekufat Haeven, Journal of the Israel Prehistoric Society* 26, 122–133.

Garrard, A., Baird, D., Colledge, S., Martin, L., and Wright, K. (1994) Prehistoric environment and settlement in the Azraq Basin: an interim report on the 1987 and 1988 excavation seasons. *Levant* 26, 73–109.

Garrard, A.N., Harvey, C.P.D. and Switsur, V.R. (1981) Environment and settlement during the Upper Pleistocene and Holocene at Jubba in the Great Nefud, Northern Arabia. *ATLAL* 5, 137–148.

Gasse, F. (1980) Late Quaternary changes in lake-levels and Diatom assemblages on the southeastern margin of the Sahara. *Palaeoecology of Africa* 12, 333–350.

Gat, J.R. and Magaritz, M. (1980) Climatic variations in the eastern Mediterranean sea area. *Naturwissenschaften* 67, 80–87.

Gilead, I. (1995) *Grar, A Chalcolithic Site in the Northern Negev.* Ben-Gurion University Press, Be'er-Sheva.

Ginat, H. and Zilberman, E. (1992) Structural and morphological development of the ʿUvda valley. *Israel Journal of Earth Science* 40, 209–218.

Glueck, N. (1935) *Explorations in Eastern Palestine II.* AASOR 15, New Haven.

——. (1961) The archaeological history of the Negev. *Hebrew Union College Annual* 32, 11–18.

——. (1968) *Rivers in the Desert.* Grove Press, New York (revised edition).

——. (1970) *The Other Side of the Jordan.* American School of Oriental Research, Cambridge (revised edition).

Goldberg, P. (1986) Late Quaternary environmental history of the southern Levant. *Geoarchaeology* 1, 225–244.

Goldberg, P. and Bar-Yosef O. (1982) Environmental and archaeological evidence for climatic change in the southern Levant. *In* J.L. Bintliff and W. Van Zeist (eds.), *Palaeoclimates, Palaeo-*

environments and Human Communities in the Eastern Mediterranean Region in Later Prehistory, BAR International Series 133, Oxford, pp. 399–418.

——. (1990) The effect of man on geomorphological processes based upon evidence from the Levant and adjacent areas. *In* S. Bottema, G. Entjes-Nieborg and W. Van Zeist (eds.), *Man's Role in the Shaping of the Eastern Mediterranean Landscape,* A.A. Balkema, Rotterdam, pp. 71–86.

Goldberg, P. and Rosen, A.M. (1987) Early Holocene Palaeoenvironments of Israel. *In* T.E.Levy (ed.), *Shiqmim, Studies Concerning Chalcolithic Societies in the Northern Negev Desert, Israel (1982–1984),* BAR International Series 356 (i), Oxford, pp. 23–33.

Goodfriend, G.A. (1988) Mid-Holocene rainfall in the Negev desert from ^{13}C of land snail shell organic matter. *Nature* 333, 757–760.

——. (1991) Holocene trends in ^{18}O in land snail shells from the Negev desert and their implications for changes in rainfall source areas. *Quaternary Research* 35, 417–426.

Goodfriend, G.A., Magaritz, M. and Carmi I. (1986) A high stand of the Dead Sea at the end of the neolithic period: Paleoclimatic and archeological implications. *Climatic Change* 9, 349–356.

Gopher, A. (1981) The Stratrgraphy and Flint Industry in Wadi Tbeik, a PPNB Site in Southern Sinai. Unpublished MA thesis, Hebrew University, Jerusalem (Hebrew).

——. (1994) *Arrowheads of the Neolithic Levant, a Seriation Analysis.* Eisenbrauns, Winona Lake.

Gopher, A., Goring-Morris, A.N. and Rosen, S.A. (1995) 'Ein Qadis I: A pre-pottery neolithic B occupation in eastern Sinai. *'Atiqot* 27, 15–33.

Gophna, R. (1997) *Excavations at Tel Dalit.* Tel-Aviv University, Tel-Aviv.

Gophna, R. and Portugali, Y. (1988) Settlement and demographic processes in Israel's coastal plain from the Chalcolithic to the Middle Bronze Age. *BASOR* 269, 11–28.

Goren-Inbar, N. (1993) Ethnoarchaeology: The southern Sinai Bedouin as a case study. *In* J. Aviram (ed.), pp. 417–419.

Goring-Morris, AN. (1993) From foraging to herding in the Negev and Sinai: The early to late Neolithic transition. *Paleorient* 19, 65–89.

Goring-Morris, A.N. and Gopher, A. (1983) Nahal Issaron: A neolithic settlement in the southern Negev. *IEJ* 33, 149–162.

——. (1987) Nahal Issaron: a neolithic site in the southern Negev. *Qadmoniot* 20, 18–21(Hebrew)

Haiman, M. (1986) *Archaeological Survey of Israel: Map of Har Hamran—Southwest (198).* Israel Antiquities Authority, Jerusalem.

Haiman, M. (1988) The Iron Age Sites of the Negev Highland. Unpublished M.A. thesis, Hebrew University, Jerusalem.

——. (1989) Preliminary Report of the Western Negev Highlands Emergency Survey. *IEJ* 39, 173–191.

——. (1991) *Archaeological Survey of Israel: Map of Mizpe Ramon Southwest (200).* Israel Antiquities Authority, Jerusalem.

——. (1992a) Sedentarism and pastoralism in the Negev highlands in the early Bronze Age: Results of the Western Negev Highlands Emergency Survey. *In* O. Bar Yosef and A. Khazanov (eds.), pp. 93–104.

——. (1992b) Early Settlement Patterns of the Negev Highlands. Unpublished Ph.D. dissertation. Hebrew University, Jerusalem (in Hebrew).

——. (1993) *Map of Har Hamran Southeast (199).* Israel Antiquities Authority, Jerusalem.

——. (1994) The early Bronze Age site of Ramat Matred 3 in the Negev highlands. *'Atiqot* 25, 23–32.

——. (1996) Early Bronze Age IV settlement pattern of the Negev and Sinai deserts: View from small marginal temporary sites. *BASOR* 303, 1–31.

Harlan, J.R. (1967) A wild wheat harvest in Turkey. *Archaeology* 20, 197–201.

Hassan, F.A. (1981) *Demographic Archaeology.* Academic Press, New York.

Hays, T. R. (1975) Neolithic settlement of the Sahara as it relates to the Nil Valley. *In* F. Wendorf and A. Marks (eds.), *Problems in Prehistory: North Africa and the Levant,* Southern Methodist University Press, Dallas, pp. 193–201.

Hays, T.R. (1992) Neolithic chronology in the Sahara and Sudan. *In* Ehrich (ed.), *Chronology in Old World Archaeology (3rd. edition)*. Vol. I, 309–316, Vol. II, 268–275, University of Chicago Press, Chicago.

Henry, D.O. (1995) *Prehistoric Cultural Ecology and Evolution*. Plenum Press, New York.

Hillman, G.C. and M.S. Davies. (1990) Measured domestication rates in wild wheats and barley under primitive cultivation, and their archaeological implications. *JWP* **4**, 157–222.

Holzer, A. and Avner, U. (In print) Excavations of desert kites in the Southern Arabah. *Hadashot Archeologiot* and *IES*.

Hopf, M. (1978) Plant remains, strata V-I. *In* R. Amiran, *et al.*, (eds.), *Early Arad I*, Israel Exploration Society, Jerusalem, Pp. 64–82.

Hornel J. (1930) The Cypriot treshing sledge. *Man* **30**, 135–139.

Horowitz, A. (1976) Late quaternary paleoenvironments of prehistoric settlements in the Avdat/Aqev Area. *In* A. Marks (ed.), *Prehistory and Paleoenvironment in the Central Negev, Israel I*, Southern Methodist University Press, Dallas., pp. 57–72.

——. (1979) *The Quaternary of Israel*. Academic Press, New York.

Hugot, H.J. (1968) The origins of agriculture: Sahara. *CA* **9**, 483–488.

Huntington, E. (1911) *Palestine and its Transformation*. Houghton Mifflin Co., Boston.

——. (1924) *Civilization and Climate*. Yale University Press, New Haven.

Jacobsen, T. (1982) Salinity and Irrigation Agriculture in Antiquity, Diyala Basin Archaeological Projects: Report on Essential Results, 1957–58. *Bibliotheca Mesopotamia 14*. Undena Publication, Malibu.

Jean, C.F. and Hoftizeizer, J. (1965) *Dichtionary des Inschriptions Semitique de l'Ouest*. E.J. Brill, Leiden.

Khalil, L. (1992) some technological features from a Chalcolithic site at Magass-Aqaba. *In* S. Tell (ed.), *Studies in the History and Archaeology of Jordan IV*, Department of Antiquities, Amman, pp. 143–148.

Khazanov, A.M. (1984) *Nomads and the Outside World*. Cambridge University Press, Cambridge.

Kislev, M.E. (1984) Emergence of wheat agriculture. *Paleorient* **10**, 61–70.

——. (1995) Wild olive stones at submerged Chalcolithic Kfar Samir, Haifa, Israel. *Mitekufat Haeven, Journal of the Israel Prehistoric Society* **26**, 134–145 (Hebrew).

Kislev, M.E., Nadel, D. and Carmi, I. (1992) Epipalaeolithic (19,000 BP) cereal and fruit diet at Ohalo II, Sea of Galilee, Israel. *Review of Palaeobotany and Palynology 73*, 161–166.

Kohler-Rollefson, I. (1993) Camels and camel pastoralism in Arabia. *BA* **56**, 180–188.

Koppel, S.J.R. (1940) *Teleilat Ghassul II*. L'Institut Biblique Pontificial, Rome.

Kramer, C. (1978) Estimating prehistoric population: An athnographical approach. *L' Archeologie de l' Iraq du debut de l' Epoque Neolithique a 33 avent notre Ere-Perspective et limites l'Interpretation anthropologique des documents*. Colloque International C.N.R.S. No. **580**, Paris.

——. (1979) An archaeological view of a contemporary Kurdish village: Domestic architecture, household size, and wealth. *In* C. Kramer (ed.), *Ethnoarachaeology: Implications of Ethnography for Archaeology*, Columbia University Press, New York, pp.139–163.

Kropelin, S. (1987) Palaeoclimatic evidence from early to mid-holocene playas in the Gilf Kebir (Southwest Egypt). *In* J. A. Coetzee (ed.), *Palaeoecology of Africa and the Surrounding Islands*, A.A. Balkema, Rotterdam, pp.189–208.

Kutzbach, J.E., and Street-Perrott, F.A. (1985) Milankovitch forcing of fluctuations in the level of Tropical Lakes from 18 to 0 kyr BP. *Nature* **317**, 130–134.

Lee, J.R. (1973) Chalcolithic Ghassul: New Aspects and Master Typology. Unpublished Ph.D. dissertation. Hebrew University, Jerusalem.

Lender, Y. (1990) *Archaeological Survey of Israel: Map of Har Nafha (196)*. Israel Antiquities Authority, Jerusalem.

Levy, T.E., and Rosen, S.A. (1987) The chipped stone industry at Shiqmim: Typological considerations. *In* T. E. Levy (ed.), *Shiqmim I, Studies Concerning Chalcolithic Societies in the Northern Negev Desert, Israel (1982–1984)*, BAR International Series **356** (i), Oxford, pp. 281–294.

Lhote, H. (1958) *Peinturs Prehistoriques Du Sahara.* Musee des Arts Decoratifs, Paris.

——. (1961) The rock art of the Maghreb and Sahara. *In* H.G. Bandi (ed.), *Art of the Stone Age*, Crown Publishers, New York, pp.99–152.

Licent, E. and de Chardin, P.T. (1925) Note sur deux instruments agricoles du Neolithique de Chine. *L'Anthropologie* 35, 63–74.

Liphschitz, N. (1986) The vegetational landscape and the macroclimate of Israel during prehistoric and protohistoric periods. *Mitekufat Haeven* 19, 80–90.

Luz, B. (1982) Paleoclimatic interpretation of the last 20,000 years. Record of deep sea cores around the Middle East. *In* J. L. Bintliff and W. Van Zeist (eds.), *Palaeoclimates, Palaeoenvironments and Human Communities in the Eastern Mediterranean Region in Later Prehistory*, BAR International Series 133, Oxford, pp.41–61.

Mallon, A., Koeppel, R. and Neuville, R. (1934) *Teleilat Ghassul I.* Scripta Pontificil Instituti Biblici. Rome.

Mallowan, M.E.L. (1964) Noah's flood reconsidered. *IRAQ* 26, 62–82.

Marfoe, L. (1980) Book review of early Arad I (R. Amiran). *JNES* 31, 315–322.

Marks, A. (1976) *Prehistory and Paleoenvironment in the Central Negev, Israel I.* Southern Methodist University Press, Dallas

Marx, E. (1967) *Bedouin of the Negev.* Manchester University, Manchester.

——. (1988) The ecological base of the Beduin society. *In* J. Eini and E. Orion (eds.), *The Beduins, Notes and Articles.* Ben-Gurion University, Be'er Sheva, pp. 73–95. (Hebrew).

——. (1992) Are there pastoral nomads in the Middle East? *In* O. Bar Yosef and A. Khazanov (eds.), pp. 255–260.

McC. Adams, R. (1981) *Heartland of Cities.* University of Chicago, Chicago.

McC. Adams, R., and Nissen, H.J. (1972) *The Uruk Countryside.* University of Chicago, Chicago.

McCartney, C.J. (1992) Preliminary report of the 1989 excavations at site 27 of the Burqu'-Ruweishid project. *Levant* 24, 33–54.

McClure, H.A. (1976) Radiocarbon chronology of late quaternary lakes in the Arabian desert. *Nature* 263, 755–756.

McConaughy, M.A. (1979) Formal and Functional Analyses of the Chipped Stone Tools from Bab Edh-Dhra, Jordan. Unpublished Ph.D. dissertation, University of Pittsburgh, Pittsburgh.

Mellaart J. (1967) *Çatal Huyuk.* Thames and Hudson, London.

——. (1970) *Haçilar I.* British Institut of Archaeology at Ankara. Edinburg.

Milburn M. (1989) Enigmatic stone objects of the Saharan neolithic and post neolithic. *In* L. Krzyzaniak and M. Kobusiewich (eds), *Late Prehistory of the Nile Basin and the Sahara,* Poznan Archaeological Musseum, Poznan, pp. 531–547.

Milburn M. and Rees, S.E. (1984) Possible neolithic ard-tip from southern Sahara. *Antiquity* 58, 45–48.

Musil, A. (1907) *Arabia Petraea, Vol. II.* Keiserliche Akademie der Wissenschaften, Vienna.

——. (1926) *The Northern Hegaz.* American Geographical Society, New York.

Muzzolini, A. (1994) Chronologie reisonee des diverses écoles d'art rupestre du Sahara Central. *Memorie della Societa' Italiani di Scienzi Naturalie del Museo Civico di Storia Natural di Milano* XXVI, Fosc. II, 387–397.

——. (1995) *Les Images Rupestres du Sahara.* Muzzolini, Toulouse.

Naroll R. (1962) Floor area and settlement population. *American Antiquity* 27, 587–589.

Naveh, J. (1982) *Early History of the Alphabet.* Magnes, Jerusalem

Neef, R. (1990) Introduction, development and environmental implications of olive culture: the evidence from Jordan. *In* Bottema, Entjes-Neieborg and Van Zeist (eds.), *Man's Role in the Shaping of the Eastern Mediterranean Landscape,* A.A. Balkema, Rotterdam, pp. 295–306.

Neev, D. (1980) A correlation between four holocene tectonic and pluvial phases in Israel. *In* Z. Reiss, M. Shilo, D. Yaalon, B. Luz, and J. Gat (eds.), *Approaches and Methods in Paleoclimatic Re-*

search with Emphasis on Aridic Areas, Bat Sheva Seminar 37, (1980), Bat Sheva de Rothchild Foundation, Jerusalem, pp. 55–56.

Nicholson, S.E. and H. Flohn. (1980) African environmental and climatic changes and the general atmospheric circulation in the late pleistocene and holocene. *Climatic Change* 2, 313–348.

Nissen, H.J., Muheisen, M. and Gebel, H.G. (1991) Report on the excavations at Basta (1988). *ADAJ* 35, 13–40.

North, S.J.R. (1961) Ghassul 1960, Excavation Report. Pontifical Biblical Institute, Rome.

Oates, J. (1982) Archaeological evidence for settlement patterns in Mesoptamia and Eastern Arabia. *In* J. L. Bintliff and W. Van Zeist (eds.), *Palaeoclimates, Palaeoenvironments and Human Communities in the Eastern Mediterranean Region in Later Prehistory,* BAR International Series 133, Oxford, pp. 359–398.

Ovadia, E. (1992) The domestication of the ass and pack transport by animals: a case of technological change. *In* O. Bar Yosef and A. Khazanov (eds.), pp. 19–28.

Pearlman, D.A. (1984) Threshing Sledges in the East Mediterranean: Ethnoarachaeology with Chert Knappers and Dhoukanes in Cyprus. Unpublished M.A. thesis. University of Minnesota, Minneapolis.

Pearlman, D.A. (1987) On the Dhoukani trail. *Cyprus Life* 1, 57–58.

Perevolotsky, A., and Baharav, D. (1987) The abundance of desert kites in eastern Sinai—An ecological analysis. *In* G. Gvirtzman, A. Shmueli, Y. Gradus, I. Beit-Arieh and M. Har-El (eds.), *Sinai,* Tel-Aviv University, Tel-Aviv, pp. 595–604 (in Hebrew).

Perrot, J. (1961. Une tombe a ossuaires du quatrieme millennium a Azor. *'Atiqot* 3, 1–83.

Perrot, J. and Gopher, A. (1996) A Late Neolithic site near Ashkelon. *IEJ* **46**, 145–166.

Raikes, R.L. (1966). The physical evidence for Noah's flood. *IRAQ* **28**, 52–63.

Rast, W.E. and Schaub, R.T. (1980) Preliminary report of the 1979 expedition to the Dead Sea plain, Jordan. *BASOR* **240**, 21–61.

Redman, G.L. (1978) *The Rise of Civilization.* Freeman, San Francisco.

Rees, S.E. (1979) *Agricultural Implements in Prehistoric and Roman Britain.* Bar International Series 69, Oxford.

Reifenberg, A. (1955) *The Desert and the Sown.* Jewish Agency, Jerusalem.

Retso, J. (1991) The domestication of the camel and the establishment of the Frankincense Road from South Arabia. *In* T. Kronholm (ed.), *Orientalia Suecana XL,* Almqvist & Wiksell, Stockholm, pp. 187–219.

Ripinsky, M. (1983) Camel ancestry and domestication in Egypt and the Sahara. *Archaeology* **36**, 21–27.

Ritter-Kaplan, H. (1984) The impact of drought on the third millennium BCE. Cultures on the basis of excavations in the Tel-Aviv Exhibition Grounds. *Eretz Israel* 17, 333–338 (in Hebrew, English summary).

Rizkana, I. and Seeher, J. (1988) *Maadi II.* Philipp Von Zabern, Mainze am Rein.

Roberts, N. (1982) Lake levels as an indicator of near eastern paleoclimate. *In* J.L. Bintliff and W. Van Zeist (eds.), *Palaeoclimates, Palaeoenvironments and Human Communities in the Eastern Mediterranean Region in Later Prehistory,* BAR International Series **133**, Oxford, pp. 235–271.

Rollefson, G.O. (1992) Neolithic settlement patterns in northern Jordan and Palestine. *In* S. Tell (ed.), *Studies in the History and Archaeology of Jordan IV,* Department of Antiquities, Amman, pp. 123–127.

Ronen, A. (1980) Nahal Re'uel site. *Hadashot Archeologiot* **74/75**, 47–48.

——. (1986) Quaternary alluvial stratigraphy of the Shephela and its paleoclimatic implications. *Geological Survey of Israel, Report 25.*

Rosen, S. (1982) Flint sickle blades of the late protohistoric and early historic periods in Israel. *TA* **9**, 139–145.

——. (1983a) Lithics in the Bronze and Iron Ages in Israel. Unpublished Ph.D. dissertation, University of Chicago, Chicago.
——. (1983b) The microlithic lunate: an old-new tool type from the Negev, Israel. *Paleorient* 9, 81–83.
——. (1983c) The Canaanean blade and the early Bronze Age. *IEJ* 33, 15–29.
——. (1984a) Kvish Harif: Preliminary investigation at a late neolithic site in the central Negev. *Paleorient* **10**, 111–121.
——. (1984b) The adoption of metallurgy in the Levant: A lithic perspective. *CA* 25, 504–505.
——. (1985) The En Shadud Lithics. *In* E. Braun (ed.), *En Shadud, Salvage Excavations at a Farming Community in the Jezreel Valley, Israel,* BAR International Series **249**, Oxford.
——. (1987a) The potentials of lithic analysis in the chalcolithic of the northern Negev. *In* T.E. Levy (ed.), *Shiqmim I, Studies Concerning Chalcolithic Societies in the Northern Negev Desert, Israel (1982–1984)*, BAR International Series **356** (i), Oxford, pp. 295–309.
——. (1987b) Demographic trends in the Negev highlands: Preliminary results from the Emergency Survey. *BASOR* **266**, 45–58.
——. (1988) Notes on the origins of pastoral nomadism: A case study from the Negev and Sinai. *CA* **29**, 498–506.
——. (1989) The analysis of early Bronze Age chipped stone industries: a summary statement. *In* P. de Miroschedji (ed.), *L'Urbanisation de la Palestine a l'Age du Bronze Ancien. Actes du Colloque d'Emmaus October (1986),* BAR International Series **527**, Oxford, pp.199–222.
——. (1991) The lithic assemblage from the EB site at Har Horsha. *'Atiqot* **20**, 169–176.
——. (1992) Nomads in archaeology: A response to Finkelstein and Perevolotsky. *BASOR* **287**, 75–85.
——. (1993) The lithic assemblages from Nahal Mitnan. *'Atiqot* **22**, 62–69.
——. (1994) *Map of Makhtesh Ramon (203).* Israel Antiquities Authority, Jerusalem.
——. (1997) *Lithics After the Stone Age.* Altamira, Walnut Creek.
Rosen, B. and Finkelstein, I. (1992) Subsistence patterns, carrying capacity and settlement oscillations in the Negev Highlands. *PEQ* **124**, 42–58.
Rothenberg, B. (1967a) *Negev. Archaeology in the Negev and the Arabah.* Masada, Ramat Gan (Hebrew).
——. (1967b) Archaeological sites in the southern Arabah and Eilat mountain. *In* T. Ron (ed.), *Survey of Eilot District, Vol. II,* Eilot District Municipality, pp. 303–307.
——. (1970) An archaeological survey of south Sinai, First Season 1967/1968. *PEQ* **102**, 4–29.
——. (1979) *Sinai.* Kummerly und Frey, Bern.
Rothenberg, B., and Cohen, E. (1968) An archaeological survey of the Eloth district and the southernmost Negev. *Museum Ha'aretz* **10**, 25–35.
Rowan Y. and Levy, T.E. (1991) Use wear analysis of a chalcolithic scraper assemblage from Shiqmim. *Mitekufat Haeven, Journal of the Israel Prehistoric Society* **24**, 112–134.
——. (1994) Proto-Canaanean blades of the chalcolithic period. *Levant* **26**, 167–174.
Rowton, M.B. (1973) Autonomy and nomadism in Western Asia. *Orientalia* **42**, 247–258.
——. (1974) Enclosed nomadism. *Journal of the Economic and Social History of the Orient* **17**, 1–30.
Russell, K.W. (1995) Traditional Bedouin agriculture at Petra: ethnoarchaeological insights into the evolution of food production. *In* G. Bisheh (ed.), *Studies in the History and Archaeology of Jordan V,* Department of Antiquities, Amman, pp. 693–705.
Sanlaville, P. (1992) Changements climatiques dans la Peninsule Arabique durant le Pleistocene Superieur et l'Holocene. *Paleorient* **18**, 5, 26.
Schon, W. (1989) New results from two playa-sites in the Gilf Kebir (Egypt). *In* L. Krzyzaniak and M. Kobusiewicz (eds.), *Late Prehistory of the Nile Basin and the Sahara*, Poznan Archaeological Museum, Poznan, pp. 215–222.
Segal D. and Carmi, I. (1996) Rehovot radiocarbon date list V. *'Atiqot* **29**, 79–106.
Seligman, N., Tadmor, N. and Raz, T. (1962) *Range Survey of the Central Negev.* National and University Institute of Agriculture, Ramat Gan (in Hebrew).

Sharon, D. (1979) Local rainfall in the Arabah. *In* A. Shmueli and J. Gradus (eds.), *The Land of the Negev,* Tel-Aviv, pp. 119–124. (in Hebrew).

Shanan L., Evenari, M. and Tadmor, N. (1967) Rainfall pattern in the Negev desert. *IEJ* 17, 163–184.

Sherratt, A. (1981) Plough and pastoralism: aspects of the secondary products revolution. *In* I. Hodder, G. Isaac and N. Hammond (eds.), *Pattern of the Past: Studies in Honour of David Clarke,* Cambridge University Press, Cambridge, pp. 261–305.

Shiloh, Y. (1980) The population of Iron Age Palestine in the light of a sample analysis of urban plans, areas, and population density. *BASOR* 239, 25–35.

Skira, A. (1954) *Egyptian Painting.* Skira, Geneva.

Steensberg, A. (1964) A Bronze Age ard-type from Hama in Syria intended for rope traction. *Beritus* 15, 111–139

Stenhill, G. (1978) The Felah farm and autarchic eco-system. *Agro-Ecosystem* 4, 433–438.

Storkbaum, W. (1983) Ghana. *In* G. Westerman (ed.), *Encyclopedia for the World's Nations,* Vol. 1, (Hebrew edition) Braunschweig.

Street, F.A. and Grove, A.T. (1979) Global maps of lake level fluctuations since 30,000 years B.P. *Quaternary Research* 12, 83–118.

Stuiver, M and Reimer, P.J. (1993) Extended ^{14}C data base and revised calib 3.0 ^{14}C age calibration program. *Radiocarbon* 35, 215–230.

Sussman, V. and Ben-Arieh, S. (1966) Ancient burials in Giv'atayim. *'Atiqot* 3, 27–39 (in Hebrew, English summary).

Tchernov, E. and Bar-Yosef, O. (1982) Animal exploitation in the pre-pottery neolithic B period at Wadi Tbeik, southern Sinai. *Paleorient* 8, 17–37.

Thomson, J., Cook, G.T., Anderson, R., Mackenzie, A.B., Harkness, D.D., and McCave, I.N. (1995) Radiocarbon age offsets in different-sized carbonate components of deep-sea sediments. *Radiocarbon* 37, 91–101.

Van Beek G.W. (1982) A population estimate for Marib, a contemporary tell in north Yemen. *BASOR* 248, 61–67

Whallon, R.J. (1978) Threshing sledge flints. A distinctive pattern of wear. *Paleorient* 4, 319–324.

Wilkinson, T.J. (1978) Erosion and sedimentation along the Euphrate valley in north Syria. *In* W.C. Brice (ed.), *The Environmental History of the Near and Middle East Since the Last Ice Age.* Academic Press, London, pp. 215–226.

Yakar, Jak. (1976) Anatolia and the 'Great movement of Indo-Europeans', *ca.* 2300 BCE—Another look. *TA* 3, 151–160.

Zarins, J. (1979) Rejejel, a unique Arabian site from the 4th millennium BC. *ATLAL* 3, 73–77.

——. (1992) Pastoral nomadism in Arabia: Ethnoarchaeology and the archaeological record—A case study. *In* O. Bar Yosef and A. Khazanov (eds), pp. 219–240.

Zarins, J. and Al-Badr, H. (1986) Archaeological investigation in the southern Tihama Plain II (including Sihi, 217–107 and Sharja, 217–172) 1405/1985. *ATLAL* 10, 36–57.

Zarins, J., Murad, A. and Al-Yish, Kh. (1981) The comprehensive archaeological survey program: The second preliminary report on the southwestern province. *ATLAL* 5, 9–42.

Zarins, J., Rahbini, A. and Kamal, M. (1982) Preliminary report on the archaeological survey of the Riyadh area. *ATLAL* 6, 25–38.

Zarins, J., Whalen, N., Ibrahim, M., Mursi, A., and Khan, M. (1980) Comprehensive archaeological survey program: Preliminary report on the central and southwestern provinces survey (1979). *ATLAL* 4, 9–36.

Zohary, D. (1973) The origin of cultivated cereals and pulses in the Near East. *Chromosomes Today* 4, 307–320.

Zohary, D. and Hopf, M. (1993) *Domestication of Plants in the Old World.* Clarendon Press, Oxford.

Chapter 9

Middle to Late Holocene (6,500 Yr. Period) Paleoclimate in the Eastern Mediterranean Region from Stable Isotopic Composition of Speleothems from Soreq Cave, Israel

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9.1. Introduction

Stalactites and stalagmites in the Soreq Cave, a well known tourist attraction in Israel, provide a very detailed climatic record for the Eastern Mediterranean region during the last 25,000 years (Bar-Matthews *et al.*, 1997a) and 58,000 years (Bar-Matthews *et al.*, 1997b). These studies indicate that the isotopic compositions of speleothems that are older than about 6,500 BP are significantly different from those of present-day speleothems; and that about 6,500 years ago the isotopic composition of the speleothems became similar to that of today. Thus, the climatic conditions that prevailed in the Eastern Mediterranean area before ~6,500 BP were very different from now. Only from that time have the conditions become similar to those of the present day. Soreq Cave is one of a series of karstic caves situated within the steeply westward dipping flank of the Judean Hill anticline. Its geological and hydrological setting and its environment are summarized by Asaf (1975), Even *et al.* (1986) and Bar-Matthews *et al.* (1991). The cave is located approximately 40 km inland from the Israeli Mediterranean coast, and is 400 m above sea level. Presently, the climate in the Soreq Cave area is typical of the semi-arid Mediterranean type,

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with an average annual air temperature of ~20.5°C, a cave water temperature varying between 18.0 and 20.5°C (and a mean annual rainfall of ~500 mm). The area is located in a narrow transition zone between humid and arid climates; and slight climatic variations, such as changes in the annual rainfall, would have affected the desert boundary and the inhabitants in the past.

In this paper we present a continuous $\delta^{18}O$ and $\delta^{13}C$ profile of speleothems that grew over the last 6,500 years, a time period that is of special interest due to the evidence of human settlement in the Eastern Mediterranean area. Even though there is a general similarity between the isotopic composition of present-day speleothems and those that grew during the last 6,500 years, the isotopic variations that are observed during this period indicate that fluctuations did occur in the paleorainfall which were large enough to cause populations to migrate (*e.g.* Nicholson and Flohn, 1980; Amiran, 1991; Issar, 1990, 1992; Issar *et al.*, 1991; Weiss *et al.*, 1993).

9.2. Methods

We studied stalagmites and stalactites that are composed of low magnesium calcite, and which were deposited in isotopic equilibrium according to the criteria established by Hendy (1971). The petrography of the speleothems was examined to ensure that they have not been altered by secondary processes as described in Bar-Matthews *et al.* (1997a). The speleothems were sectioned perpendicular to their length, in order to expose the concentric growth layers. Samples were obtained by drilling 0.2–0.5 mg calcite powders every 0.5–1.0 mm throughout the sectioned speleothem. Sampling in such detail was required in order to obtain an isotopic profile of sufficient resolution. The drilled powder was analyzed for $\delta^{18}O$ and $\delta^{13}C$ using a VG Isocarb system attached to a SIRA-II mass-spectrometer, as described in detail by Bar-Matthews *et al.* (1997a). All $\delta^{18}O$ and $\delta^{13}C$ values are reported relative to the PeeDee Belemnite (PDB) standard.

The age determination was performed using the ^{230}Th-^{234}U method as described by Bar-Matthews *et al.* (1997a). For the dating, growth layers weighing 2–10 g were separated, and thus the calculated age was considered to represent the midpoint of each growth layer. The growth rate was calculated by dividing the distance between dated layers by their age difference. The isotopic profiles were drawn assuming constant growth rate between adjacent dated growth layers.

9.3. Results

The $\delta^{18}O$ and $\delta^{13}C$ values of speleothems that form under present-day conditions and those that were formed during the last 6,500 years are compared in Figure 9.1

with those of speleothems that grew from 20,000–6,500 BP. This diagram emphasizes (a) the similarity between present-day speleothems and those that formed during the last 6,500 years, and (b) the significant difference in isotopic composition between those that were formed during the last 6,500 years and those that were formed earlier. Speleothems that were formed during the early Holocene (~9,500–6,500 BP) have the lowest $\delta^{18}O$ and the highest $\delta^{13}C$ values of all, and speleothems that were formed during the last glacial (17,000–58,000 BP) have generally both higher $\delta^{18}O$ and higher $\delta^{13}C$ values (Bar-Matthews *et al.*, 1997 a,b).

The $\delta^{18}O$ and $\delta^{13}C$ values of speleothems that were formed during the last 6,500 years are shown in detail in Figures 9.2 and 9.3. Based on the isotopic composition, mainly on $\delta^{18}O$ values, this time span can be divided into four main stages. Their significance will be discussed below.

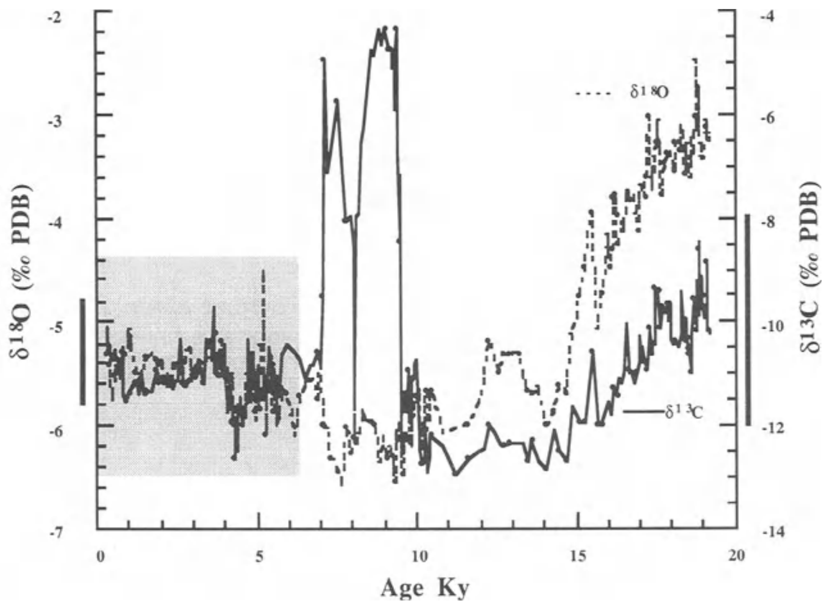

Figure 9.1. $\delta^{18}O$ and $\delta^{13}C$ variations of the speleothems back to 20,000 BP (modified from Bar-Matthews *et al.,* 1997a). The vertical bars represent the present-day range. The shaded rectangle represent the isotopic range of the speleothems that were formed during the last 65,00 years.

During stage 1, extending from 6,500–5,400 BP, the $\delta^{18}O$ values of the speleothems are the lowest for the whole period ranging from -6.1 to -5.5 ‰, with an average of -5.7‰. The $\delta^{13}C$ values vary from -11.2 to -10.5 ‰, with an average of -10.7‰.

Stage 2 which lasted from 5,400–3,050 BP is characterized by acute $\delta^{18}O$ and $\delta^{13}C$ variations. Four $\delta^{18}O$ maxima occur at 5,150 BP (-5.3‰), 5,000 BP (-4.5‰), 4,600 BP (-5.2‰) and at ~3,750 BP (-5.2 ‰); and three $\delta^{18}O$ minima occur at 5,050 BP (-6.1‰), 4,800 BP (-6.0‰) and 4,200 BP (-6.1‰). The most significant low $\delta^{18}O$ value, which occurs at 4,200 BP, is contemporaneous with the lowest $\delta^{13}C$ value of -12.7‰ (Fig. 9.3). These values are the very lowest during the last 6,500 years. The isotopic fluctuations observed during this stage generally occur over relatively short time spans, of about 100–200 years, but the most prominent low (with its turning point at 4,200 BP) lasted about 500 years, from 4,600–4,100 BP Towards the end of stage 2 there is a general trend in $\delta^{18}O$ and $\delta^{13}C$ towards maximal values.

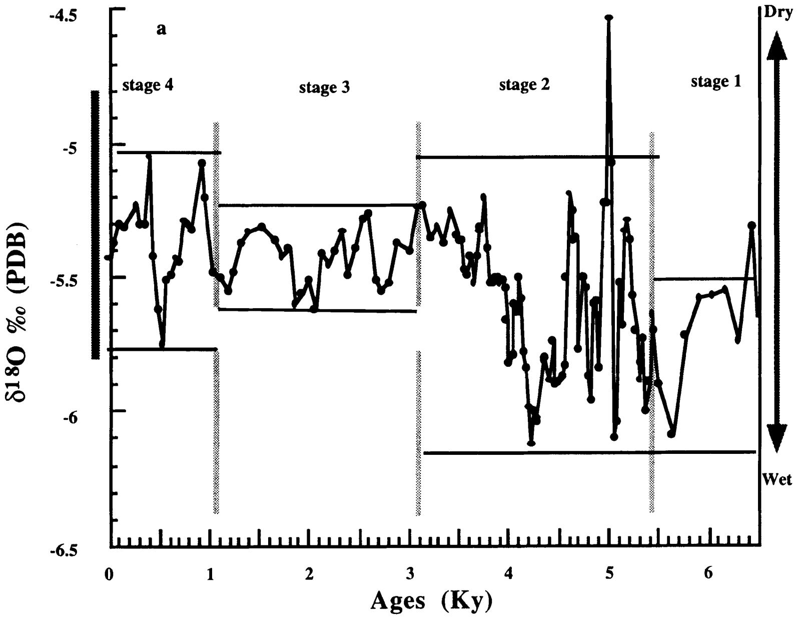

Figure 9.2. A close-up of the $\delta^{18}O$ variations of speleothems back to 6,500 BP Based on $\delta^{18}O$ values, four stages are observed and theses are divided by shaded vertical bars at 5,400, 3,050 and 1,050 BP. The range of the isotopic values within each stage are marked by black horizontal lines. The solid vertical bar on the left axis represent the present day isotopic range, and the solid vertical bar on the right axis represent the trend towards dry and wet climate.

Stage 3 extends from 3,050–1,050 BP and is characterized by quite the most steady $\delta^{18}O$ and $\delta^{13}C$ values during the last 6,500 years, with variations of -5.6 to -5.3‰ and -11.3 to -10.3‰ respectively.

The last stage, stage 4, extends from 1,050 BP to present. This period is also characterized by high $\delta^{18}O$ values. Thus $\delta^{18}O$ maxima occur at 900 (-5.1‰) and 400 BP (-5.1‰) while a $\delta^{18}O$ minimum occurs at ~500 BP (-5.75‰). The $\delta^{13}C$ values show a clear increase that starts about 700 BP at a value of -11.5‰ and continues to present-day value of -10.3‰.

9.4. Paleoclimate Reconstruction: Evidence from the Soreq Cave

The cave water in the Soreq Cave originates from the rainfall. Bar-Matthews *et al.* (1996, 1997a) have shown that the $\delta^{18}O$ values of the speleothems reflect the temperature of deposition and the $\delta^{18}O$ values of the cave water. It is thus possible to interpret the variations in the isotopic composition of fossil speleothems as a tool to evaluate changes in the paleotemperature and paleorainfall (Bar-Matthews *et al.*, 1997 a, b). This is based on the relationships observed between the amount of

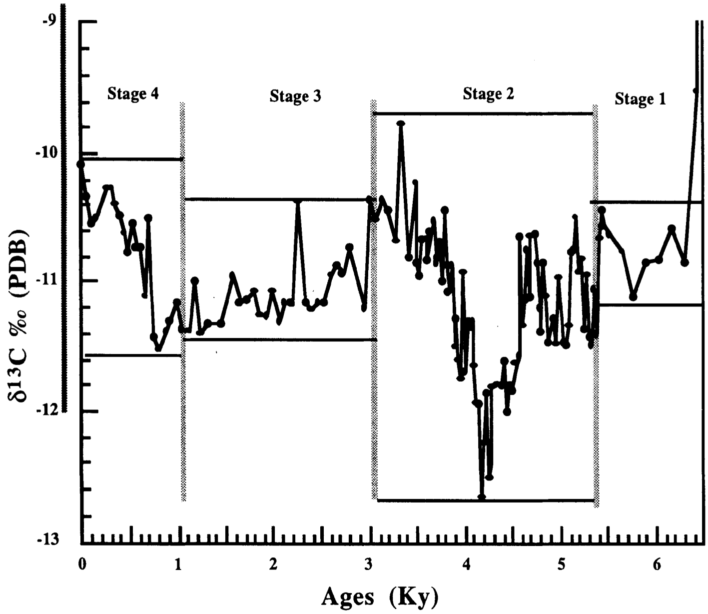

Figure 9.3. A close-up of the $\delta^{13}C$ variations of speleothems back to 6.500 BP. The four stages are divided as for the $\delta^{18}O$ (Fig. 9.2). The solid vertical bar on the left axis represent the present day isotopic range.

rainfall (and its isotopic composition) and the isotopic composition of the cave water recorded in the speleothems (Bar-Matthews *et al.*, 1996). Given the broad similarity shown by Bar-Matthews *et al.* (1997a,b) between the isotopic composition of present day speleothems and those formed in the course of the last 6,500 years, we assume that the relationships between the $\delta^{18}O$ values of speleothems, cave water and rainwater over the last 6,500 years are similar to the present day.

The $\delta^{13}C$ of cave water is the product of closed system reactions between soil-CO_2 derived from C3 type vegetation and dolomite host-rock with some CO_2 degassing. Thus the variations in $\delta^{13}C$ values of the speleothems are indicative of changes in vegetation, water-soil-rock interaction and/or CO_2 degassing (Bar-Matthews *et al.*, 1996).

The differences between the stages are emphasized in Figures 9.2 and 9.3, showing clearly that during stages 1 and 2 the $\delta^{18}O$ values of the speleothems were lower than in the later stages. In the Eastern Mediterranean area, periods with low $\delta^{18}O$ and $\delta^{13}C$ values are wetter than are periods with high $\delta^{18}O$ and $\delta^{13}C$ values (Bar-Matthews *et al.*, 1997 a). Thus this pattern is indicative of relatively wetter conditions during the first half of the interval. In addition, Stage 2 show the largest $\delta^{18}O$ and $\delta^{13}C$ fluctuations, which are indicative of 2,350 years of frequent climatic variations. The lowest $\delta^{18}O$ and $\delta^{13}C$ values occurring during this period between 4,600 to 4,100 BP (Figs. 9.2 and 9.3) most probably reflect the wettest period throughout the last 6,500 years. Meanwhile $\delta^{13}C$ values ranging from ~-12 to -10‰ reflect closed system reactions between marine carbonate host-rock and soil with C3 type vegetation. The higher values may indicate some additional degassing processes (Bar-Matthews *et al.*, 1996). The clear minimum in $\delta^{13}C$ values between 4,600 to 4,100 BP most probably reflects the larger contribution of soil CO_2 due to the larger input of infiltrating water, resulting in more efficient leaching of soil CO_2 and less degassing. From 3,050 years BP towards the present-day, the $\delta^{18}O$ values are higher and more stable (Fig. 9.2), indicating a trend towards drier and more steady climatic conditions. During the last 700 years, $\delta^{13}C$ values show a gradual increase (Fig. 9.3). This increase may be a result of deforestation and grazing during the Mamluke and Turkish periods (Sadot, personal communication).

In order to estimate the variations in the isotopic composition of the paleo cave water, we assume that the water temperature from which the speleothems were deposited was similar to that of present-day, *i.e.* 18 to 20.5°C (Bar-Matthews *et al.*, 1996). For this temperature range, we calculate the $\delta^{18}O$ value of the cave water using the calcite-water fractionation equation (O'Neil *et al.*, 1969). With the assumption that $\delta^{18}O$ of cave water in the past was ~1‰ higher than the coeval rainfall, which is the same difference that exist presently between rain and cave water (Bar-Matthews *et al.*, 1997a, Ayalon *et al.*, 1998), we then calculate the $\delta^{18}O$ of paleorain water. Using the relationships determined over the last seven years between the annual amount of rainfall and its $\delta^{18}O$ (shown in Fig. 9.4), we estimate the variations in paleorainfall during the last 6,500 years (Fig. 9.5).

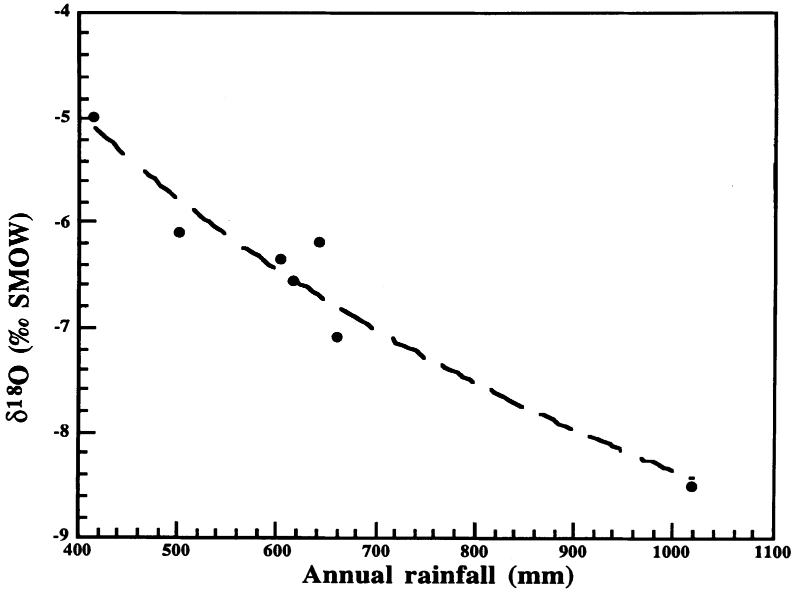

Figure 9.4. A plot showing the relationships between the average annual $d^{18}O$ of rain water and the annual rainfall, as determined since 1989, at the Soreq Cave area.

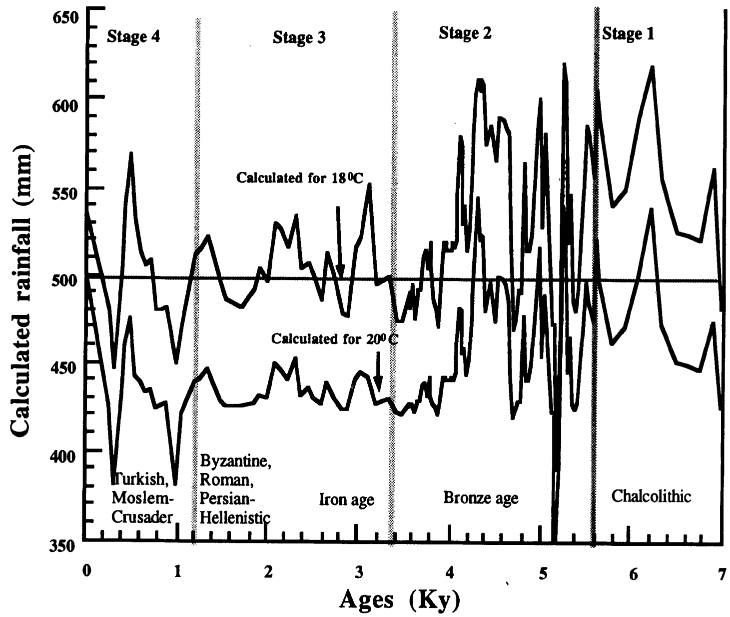

Figure 9.5. A plot showing the estimated rainfall (mm) vs. age calculated for 18°C and 20°C as discussed in the text. The vertical bars divide the period into the 4 isotopic stages (Fig 9.2). The main archeological periods are shown for references.

These paleorainfall estimates are expressed as ranges because the calculated $\delta^{18}O$ values of cave and rain waters differ if the calculation is made at 18 or 20°C, which is the range of cave water temperature from which deposition occurs (Bar-Matthews *et al.*, 1996). The estimated variations in paleorainfall corresponding to the lowest $\delta^{18}O$ values (Fig. 9.2) are ~620 and ~540 mm for 18 and 20°C respectively, while they are 440 and 380 mm for periods with the highest $\delta^{18}O$ values. The short term variations (averages of 5-year intervals) of the measured annual rainfall amount during this century (Fig. 9.6) are very similar to the estimated long term variations (50–100 year interval) in the paleorainfall we find during the last 6,500 years.

One of the key questions in studies is the scale of the climatic changes that will cause human settlements to be established, migrations to occur and cultures to develop.

Stage 1 (6,500–5,400 BP) which covers the Chalcolithic culture, follows the wettest period observed throughout the Holocene (Bar-Matthews *et al.*, 1997a and references cited therein). Although the transition to stage 1 was associated with a sharp decrease in the amount of rainfall, the paleorainfall is higher than during the later parts of the studied period. The estimated paleorainfall range for this stage varies from ~520–620 mm for 18°C (Fig. 9.5). Supporting evidence for relatively wet climatic conditions during this stage was derived also from other sources. North

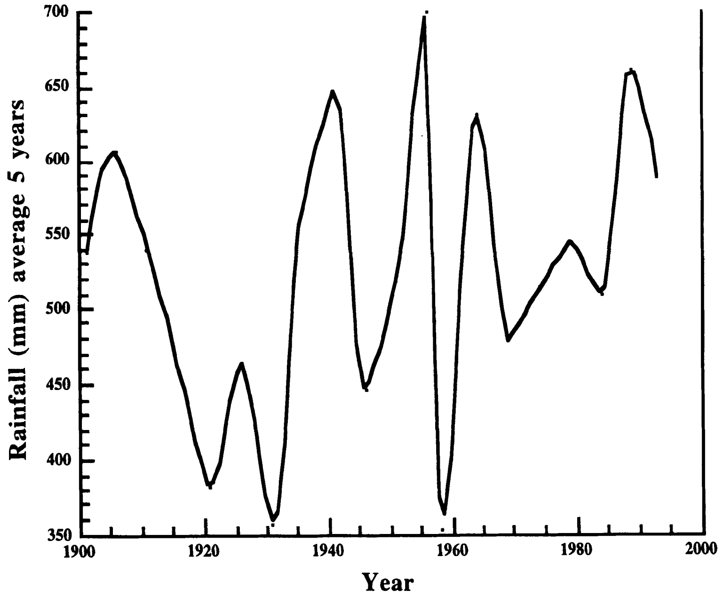

Figure 9.6. A plot showing the rainfall variation during the last century. Each data point represents the average over a five-year period.

African lakes had high stands (Street and Grove, 1979). Rainfall in the Northern Negev was moderately high as shown by the $\delta^{18}O$ values of land snails though these still indicate that the rainfall was not as high as in the early Holocene (Goodfriend, 1991). Settlements existed along the plain near river beds and along the foothills (Issar *et al.*, 1991). The city of Beer-Sheva received its water supply from shallow wells (Perrott, 1968). Goodfriend *et al.* (1986) and Goodfriend, (1987, 1991) also reported that at the beginning of this stage the Dead Sea appears to have been at a high stand, about 120 m above modern levels. However, this is rather questionable. Others suggest that the Dead Sea level dropped drastically to -400 m during this stage (Frumkin *et al.*, 1991). Interestingly, the Chalcolithic period ends very close to the end of this stage (~5,400 BP), and is followed by the Bronze Age. So is the change between the two cultures climatically controlled?

The most climatically disturbed conditions in the last 6,500 years are evidenced in the isotopic composition of stage 2 lasting from 5,600–3,050 BP During this stage the Bronze culture developed. The very sharp oscillations in $\delta^{18}O$ values indicate that the paleorainfall varied considerably between 425–610 (for 18°C, Fig. 9.5) over short time periods of 100–200 years. But although there are four short dry spells, most of the period is relatively humid. The very sharp oscillation between 5,200–5,000 BP corresponds to a period with considerable change in the population throughout the plain area (Issar *et al.*, 1991).

The wettest period recorded during this stage lasted ~500 years, between 4,600–4,100 BP, and occurs also with the lowest $\delta^{13}C$. We suggest that the average annual precipitation in the studied area reached 610 mm. A 'wet' event is discernible also from the lake levels in the Sahara desert (Nicholson and Flohn, 1980), as well as from organic matter in the Dead Sea area (Magaritz *et al.*, 1991). The city of Arad flourished (Amiran, 1991) while oak trees were found near Mount Sedom. The Dead Sea level rose by ~100 m, almost to the level typical of the early Holocene (Frumkin *et al.*, 1991; Neev and Emery, 1995). Issar (1990) suggests that at this time there was a drop in the temperature.

This event was followed by a very dry spell which, according to the $\delta^{18}O$ values of the speleothems, started ~4,150 years ago, and reached its maximum ~400 years later. We suggest that the average annual precipitation in the studied area dropped to a minimum of 470 mm. Other studies infer that this 'dry' event caused the total desertification of the Sahara, the drying out of lakes in North Africa; the salinization of the soil in Mesopotamia; the collapse of the Akkadian Empire, the northward migration of the Negev Desert, the desertion of the city of Arad; and a sharp drop in the Dead Sea level. There is also evidence for salt buildup during period of lowered sea level in the Red Sea (*e.g.* Street and Grove, 1979; Muzzolini, 1985; Street-Perrott and Perrott, 1990; Almogi *et al.*, 1991; Frumkin *et al.*, 1991; Issar *et al.*, 1991; Magaritz *et al.*, 1991; Issar, 1992; Weiss *et al.*, 1993; Hemleben *et al.*, 1996). Issar (1990) describes the changes and migration of population during the Bronze Age as a function of rapid climatic variations.

During stage 3 which lasted from 3,050–1,050 BP, the $^{18}O/^{16}O$ and $^{13}C/^{12}C$ ratios of the speleothems are higher but less variable than in the previous two stages, indicating a transition to drier and more stable conditions. The estimated annual precipitation is 470–530 mm (at 18°C). Many cultures developed in the region during this time period: Iron Age, Persian-Hellenist, Roman, and Byzantine. Many other studies indicate also that ~3,000 years ago arid conditions expanded throughout the entire region. Street and Grove (1979) claim that African oases began to shrink. Muzzolini (1985) and Issar (1992) reported that an increase in soil salinization in Mesopotamia. The worst aridity throughout the Holocene in the Negev is adjudged to have set in between 3,000–2,000 BP However Gat and Magaritz, (1980), Magaritz and Heller (1980), and Magaritz *et al.* (1991) suggest that the present climate regime started about 3,000 BP.

During the last stage, lasting from 1,050 to present, three major short-lived events can be recognized based on the $\delta^{18}O$ values: two maxima at ~900 BP and at ~400 BP (in both cases $\delta^{18}O$ values are ~-5.0‰), and a minimum at 500 BP (~-5.8‰). Because the lowest age that we have measured until now is ~1,000 BP, it is not yet clear whether our extrapolations within this stage are accurate. Thus, although these fluctuations are confirmed their timing is still questionable. What is important to emphasize, however, is the general trend of higher $\delta^{13}C$ values towards present day, starting ~700 years ago. The steady increase in $\delta^{13}C$ values could reflect extension of grazing, during the Turkish period and later, which involved the removal of the natural forest and the growth of C4 type grass.

The estimated paleorainfall varied over short periods of time. The minimum values are ~450 mm and the maximum ~560 mm (at 18°C). It might be that the relatively sharp isotopic fluctuations between 400 and 500 BP are associated with the Little Ice Age.

A comparison between the estimated paleorainfall and the measured annual rainfall during the last century shows that the two are of similar magnitude in both average rainfall and amplitude of fluctuation. In detail, however, the average estimated rainfall during stages 3 and 4 are very similar to the average of the this century (~500 mm), whereas during stages 1 and 2, conditions were somewhat more humid, with an estimated average of ~560 mm.

Acknowledgements

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References

Almogi-Labin, A., Henleben, C., Meischner, D., and Erlenkeuser, H. (1991) Paleoenvironmental events during the last 13,000 years in the central Red Sea as recorded by pteropoda. *Paleoceanography* 6, 83–98.

Amiran, H.K. (1991) The climate of the ancient Near East. The early third millennium BC in the northern Negev of Israel. *Erdkunde* 45 (3), 153–162.

Asaf, M. (1975) Karstic Phenomena in the Soreq Cave. Unpublished M.Sc. dissertation. Hebrew University of Jerusalem.

Ayalon, A., Bar-Matthews, M. and Sass, E. (1998) Rainfall-recharge relationships within a karstic terrain in the Eastern Mediterranean semi-arid region, Israel: $\delta^{18}O$ and δD characteristics. *Hydrology* (in press).

Bar-Matthews, M., Matthews, A. and Ayalon, A. (1991) Environmental controls of speleothems mineralogy in a karstic dolomitic terrain (Soreq Cave, Israel). *Journal of Geology* 99, 189–207.

Bar-Matthews, M., Ayalon, A., Matthews, A., Sass, E., and Halicz, L. (1996) Carbon and oxygen isotope study of the active water-carbonate system in a karstic Mediterranean cave: implications for paleoclimate research in semiarid regions. *Geochim. Cosmochim. Acta* 60, 337–347.

Bar-Matthews, M., Ayalon, A. and Kaufman, A. (1997a) Late Quaternary paleoclimate in the eastern Mediterranean region from stable isotope analysis of speleothems at Soreq Cave, Israel. *Quaternary Research* 47, 155–168.

——. (1997b) Paleoclimate evolution in the eastern Mediterranean region during the last 58,000 yr as derived from stable isotopes of speleothems (Soreq Cave, Israel). *In* P. Murphy (ed.), *Proc. of the Inter. Symp. on Isotopes Techniques in the Study of Past and Current Environmental Changes in the Hydrosphere and the Atmosphere.* Inter. Atomic Energy Agency, Vienna (in press).

Frumkin, A., Magaritz, M., Carmi, I., and Zak, I. (1991) The Holocene climatic record of the salt caves of Mount Sedom, Israel. *The Holocene* 1, 191–200.

Gat, J.R. and Magaritz, M. (1980) Climatic variations in the Eastern Mediterranean Sea area. *Naturwissenschaften* 67, 80–87.

Goodfriend, G.A. (1987) Chronostratigraphic studies of sediments in the Negev Desert, using amino acid epimerization analysis of land snail shells. *Quaternary Research* 28, 374–392.

——. (1991) Holocene trends in ^{18}O in land snails shells from the Negev Desert and their implications for changes in rainfall source areas. *Quaternary Research* 35, 417–426.

Goodfriend, G.A., Magaritz, M. and Carmi, I. (1986) A high stand of the Dead Sea at the end of the Neolithic period: Paleoclimatic and archaeological implications. *Climatic Change* 9, 349–356.

Hemleben, C., Meischner, D., Zahn, R., Almogi-Labin, A., Erlenkeuser, H, and Hiller, B. (1996) Three hundred eighty thousand year long stable isotope and faunal records from the Red Sea: Influence of global sea level change on hydrography. *Paleoceanography* 11, 147–156.

Hendy, C.H. (1971) The isotopic geochemistry of speleothems I. The calculation of the effects of different modes of formation on the isotopic composition of speleothems and their applicability as paleoclimatic indicators. *Geochimica et Cosmochimica Acta* 35, 801–824.

Issar, A.S. (1990) *Water Shall Flow from the Rock.* Springer Verlag, Heidelberg.

——. (1992) The Impact of Climatic Variations on Water Management Systems and Related Socio-economic Systems. UNESCO/IHP Project IIc.

Issar, A.S., Govrin, Y., Geyh, M.E., Wakshal, E., and Wolf, M. (1991) Climate changes during the upper Holocene in Israel. *Israel Journal of Earth Sciences* 40, 219–223.

Magaritz, M. and Heller, J. (1980) A desert migration indicator—Oxygen isotopic composition of land snail shells. *Palaeogeography, Palaeoclimatology, Palaeoecology* 32, 153–162.

Magaritz, M., Rahner, S., Yechieli, Y., and Krishnamurthy, R.V. (1991) $^{13}C/^{12}C$ ratio in organic matter from the Dead Sea area: paleoclimatic interpretation. *Naturwissenschaften* 78, 453–455.

Muzzolini, A. (1985) Les climats au Sahara et sur ses bordures, du Pleistocene final a l'aride actual. *Empuries* **47**, 8–27.

Neev, D. and Emery, K.O. (1995) *The Destruction of Sodom, Gomorrah, and Jericho.* Oxford University Press.

Nicholson, S.E. and Flohn, H. (1980) African environmental and climatic changes and the general atmospheric circulation in Late Pleistocene and Holocene. *Climatic Change* **2**, 313–348.

O'Neil J.R., Clayton R.N., and Mayeda T.K.(1969) Oxygen isotope fractionation of divalent metal carbonates. *Journal of Chemical Physics* **30**, 5547–5558.

Perrott, J. (1968) *La prehistoire Palestinienne,* un supplement au *dictionnaire de la Bible.* VIII: 416–438.

Street, F.A., and Grove, A.T.(1979) Global maps of lake-level fluctuations since 30,000 BP. *Quaternary Research* **12**, 83–118.

Street-Perrott, F.A., and Perrott, R.A. (1990) Abrupt climatic fluctuations in the tropics: The influence of Atlantic Ocean circulations. *Nature* **34**, 607–612.

Weiss, H., Corty, M.A., Wetterstorm, W., Guichard, F., Senior, L., Meadow, R., and Curnow, A. (1993) The genesis and collapse of third millennium north Mesopotamian civilization. *Science* **261**, 995–1,004.

Chapter 10

Early to Mid-Holocene Environmental Changes and Their Impact on Human Communities in Southeastern Anatolia

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10.1. Introduction

Southeastern Anatolia is a region which is influenced by both Mediterranean and continental climatic factors. It encompasses both mountain and steppic zones, and has had a history of settlement throughout the Holocene which is characterized by fluctuations between simple village farming communities and complexly organized cities and towns. Archaeologically it is a region where there have been major fluctuations in population, some perhaps related to environmental factors, others related to social and historical processes.

Apart from a few pollen and paleolimnological studies in adjacent areas, little work has been done on the Holocene climatic record in this region. Yet it is clear from geomorphological observation that there have been major landscape changes in the past 10,000 years. Some of these are related to human land use with the introduction of extensive agricultural and pastoral activity within this time period. However, much of this landscape change is also the result of fluctuations in climates that have occurred throughout the Holocene.

This paper will focus primarily on the northern portion of the Urfa Plain (also known as the Harran Plain) (Fig. 10.1) and a portion of the Euphrates Valley in Southeastern Turkey. It will explore the geomorphological evidence for both climatic change, and human impact on the landscape, as well as some of the archaeological evidence for the effects of changing environments on the human societies which occupied this culturally and environmentally marginal zone.

A. S. Issar and N. Brown (eds.), Water, Environment and Society in Times of Climatic Change, 215–240.

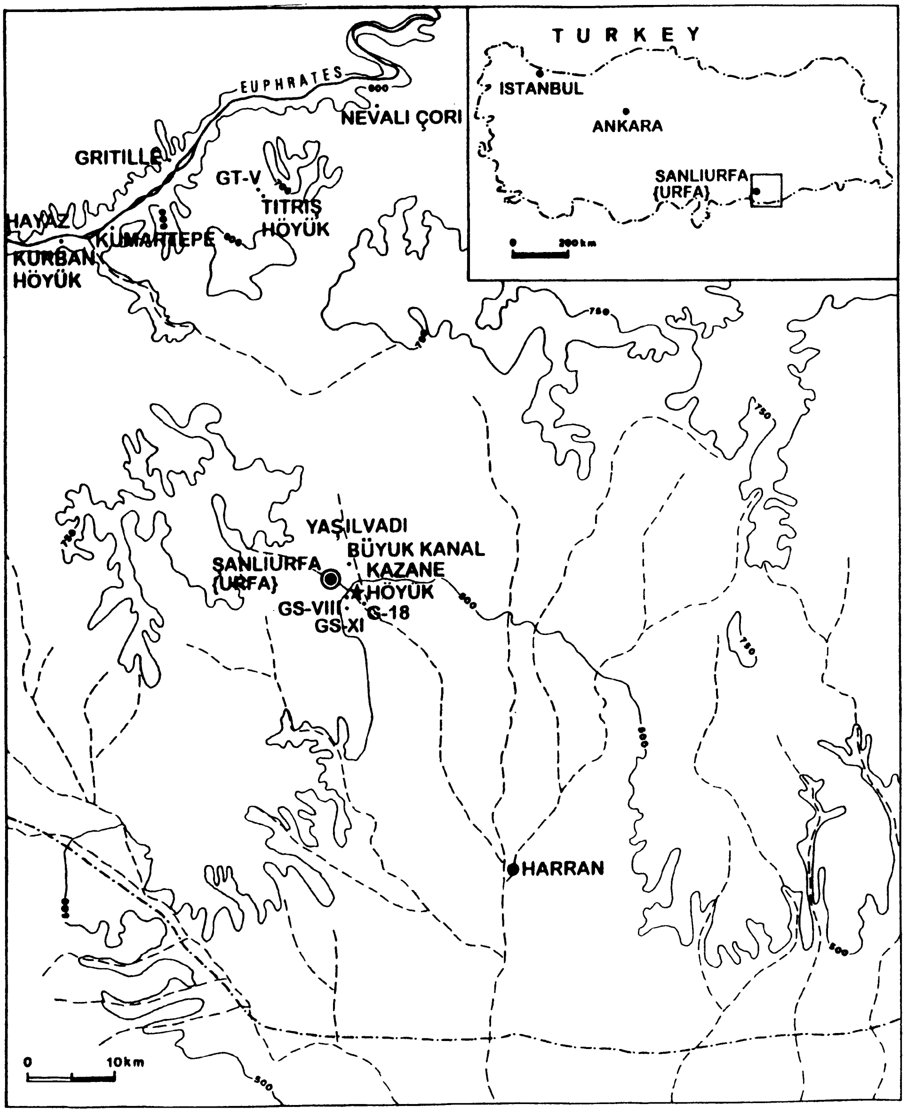

Figure 10.1. Map of the Northern Urfa Plain and Turkish Euphrates Valley, including locations of geological sections and sites mentioned in text.

10.2. Modern Environmental Setting

The chosen area of study lies at the foothills of the Eastern Taurus Mountain range between the northern Syrian plain to the south and the dissected plateau of Anatolia to the north. It is within Urfa Province, the capital of which is the modern city of Şanlıurfa (Fig. 10.1). Although under both Mediterranean and continental climatic influence, the region maintains the familiar Eastern Mediterranean pattern of warm dry summers (mean July temperature of 32°C) and cool moist winters (mean January temperature of 5°C). The effect of the continental factor is to broaden the temperature ranges, and occasionally allow heavier spring rains (Dewdney, 1971). The rainfall average for this area is approximately 450 mm per year.

The landscape is defined by low hills of Middle Eocene nummulitic limestone overlain by Oligocene-Lower Miocene chalky limestones and marls, dissected by a series of small streams draining into the Turkish Euphrates catchment. Outcrops of Pleistocene basalts cap the hills around the city of Şanlıurfa. These formations contribute limestone, flint and basalt cobbles as major components of the stream gravels in the region (Tolun, 1975). The original soils have been almost completely eroded from the hillslopes, but small remnants show that the hilly areas were mantled by red-brown terra rossas, while the valley soils are for the most part dark brown clayey loams.

The natural vegetation is steppic in the low-lying areas, with classic Mediterranean oak woodland and maquis in the hills (Dewdney, 1971). At present, little of this natural vegetation remains since much of the land is deforested and under cultivation.

10.3. Geomorphological Sequences

10.3.1. LATE PLEISTOCENE PRELUDE TO THE HOLOCENE

The geomorphological evidence for Late Pleistocene environmental and landscape changes comes primarily from the northern edge of the Urfa Plain (Fig. 10.1) at the base of the hills which ring the valley. Here, deep canal excavations have exposed sections that are tens of meters deep (Büyük Kanal Section) (Fig. 10.2, Table 10.1). At the time of these investigations, only the uppermost 14 m of the sections were left uncovered by a cement facing. Another long exposure, the Yaşilvadi Section (Table 10.2) (Fig. 10.3), is located in the low hills, five km north of the town of Şanlıurfa. Apart from these two major exposures, most of the Late Pleistocene material is exposed only at the base of pits excavated for irrigation pipes, or at the bottom of wadi cuts, with sometimes several meters of Holocene fill above them.

The Late Pleistocene sections of Büyük Kanal and Yaşilvadi are described here in order to draw a contrast between the environments of the Pleistocene and those

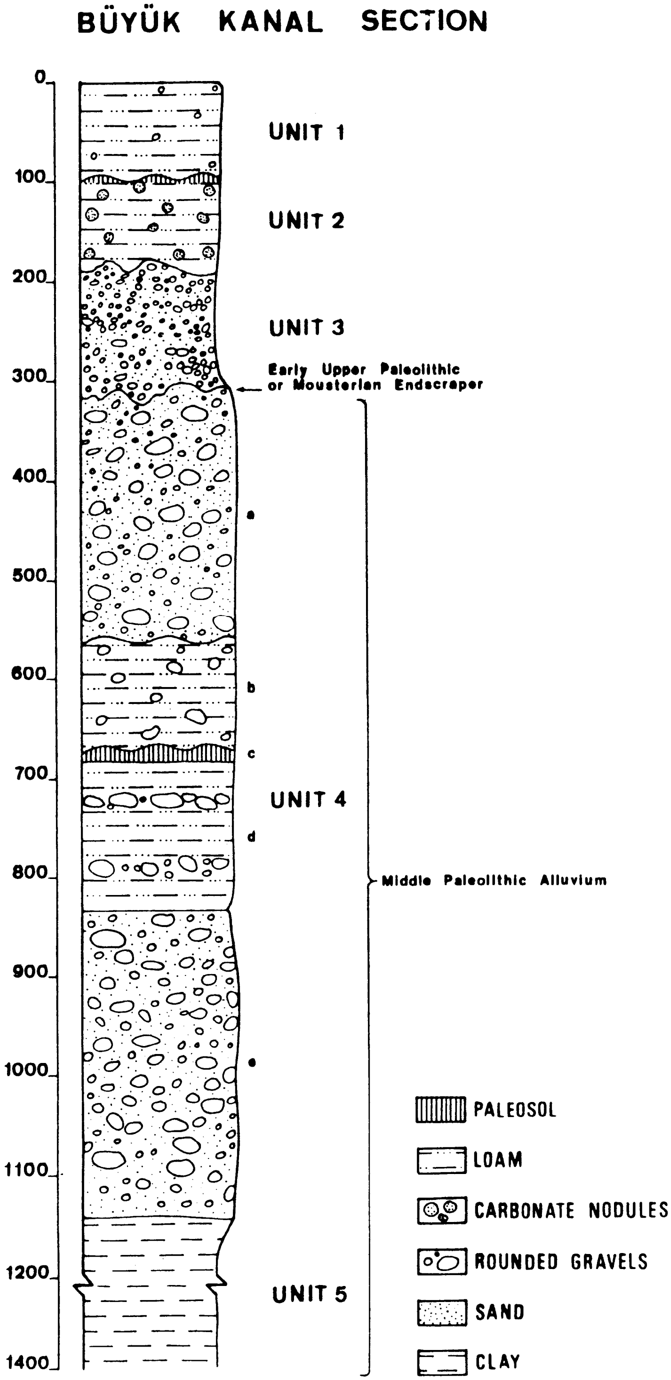

Figure 10.2. Schematic geological section of Pleistocene deposits in the Büyük Kanal Section. Scale in meters.

Table 10.1. Description of sediment units from the Büyük Kanal Section, Urfa Plain, southeastern Turkey

Unit	Depth cm	Color	Texture	Structure	Inclusions	Boundary	Depositional Env.
1	0–100	Yellowish red (5YR 4/6, m)	Clay loam	Angular blocky	Ca. 10% 1 mm–5 cm pebbles	Abrupt and wavy	Fine-grained alluvium
2	100–190	Yellowish red (5 YR 4/6, m)	Silty clay	Angular blocky	40% $CaCO_3$ nodules; mn stains	Abrupt and wavy	Paleosol on fine-grained alluvium
3	190–343	Yellowish red (5 YR 5/6, m) matrix	Moderately well-sorted pockets of 1–2 cm pebbles	Massive; uncemented	Basalt, limestone, flint	Abrupt and irregular	Channel gravels
4a	343–631	Reddish yellow (5 YR 6/6, m) matrix	Strongly indurated, poorly sorted, subrounded, 5–10 cm gravels	Massive	Limestone, basalt and flint, carbonate cement	Abrupt and wavy	Channel deposits from torrential stream flows
4b	631–751	Strong brown (7.5 YR 4/6, m)	Indurated sets of sandy silt and lenses of poorly sorted gravels	Graded bedding	Limestone, basalt and flint, carbonate cement, Fe and Mn stains	Abrupt and wavy	Channel deposits from seasonal stream flows
4c	751–754	Yellowish red (5 YR 5/6, m)	Silt	Massive	Mn nodules	Abrupt and smooth	Truncated floodplain
4d	754–814	Brown (7.5 YR 5/4, m)	Sets of sandy silt and 15 cm lenses of poorly sorted gravels	Graded bedding	Limestone, basalt and flint, carbonate cement, Fe and Mn stains	Abrupt and smooth	Channel deposits from seasonal stream flows
4e	814–1150	Brown (7.5 YR 5/4, m)	Strongly indurated, poorly sorted gravels and cobbles up to 40 cm in diameter with pockets of well-sorted gravels	Massive	Limestone, basalt and flint	Abrupt and smooth	Channel deposit
5	1150–1400+	Dark reddish brown (5 YR 3/4, m)	Fine-grained clay alluvium	Massive	1% 1 mm pebbles, mn stains, clay skins	Not visible	Fine-grained floodplain deposit

Table 10.2. Description of sediment units from the Yaşilvadi Section, southeastern Turkey

Unit	Depth cm	Color	Texture	Structure	Inclusions	Boundary	Depositional Env.
1	0–300	Brown (7.5 YR 4/6, m	Sandy loam	Angular blocky	$CaCO_3$ nodules, 20% 1 cm pebbles, flint blades (UP?) and debitage,	Abrupt and wavy	Paleosol on fine-grained alluvial floodplain
2	300–450	Light gray (5 Y 7/2, m) and Yellowish brown (10 YR 5/8, m)	Sandy marly clay	Massive	Gley, Fe stains, occasional pebbles	Abrupt and wavy	Small lake or pond
3	450–470	Reddish yellow (7.5 YR 7/6, m)	Upper part of unit sand, lower composed of poorly sorted sub-rounded gravel	Massive	Limestone, basalt, flint, ripped-up-clasts of clay, Mn and Fe stains	Abrupt and wavy	Channel of seasonal stream
4	470–865	Pink (7.5 YR 7/4 m)	Silt with pockets of very well-sorted sands and small pebbles	Graded bedding	Limestone, basalt, and flint pebbles	Abrupt and wavy	Channel deposit of well-sustained, low-energy aggrading stream. Continuation of hydrological regime evident in Unit 6
5	865–900	Brown (7.5 YR 4/4, m)	Poorly sorted gravels (<20 cm in diam)	Massive	Limestone, basalt, and flint	Abrupt and wavy	Channel deposit of high-energy torrential stream flow
6	900–1020	Very pale brown (10 YR 7/4, m)	Indurated sandy silt with pebble lenses	Massive	Gley, lenses of limestone, basalt and flint	Not visible	Channel deposit of well-sustained, low-energy aggrading stream

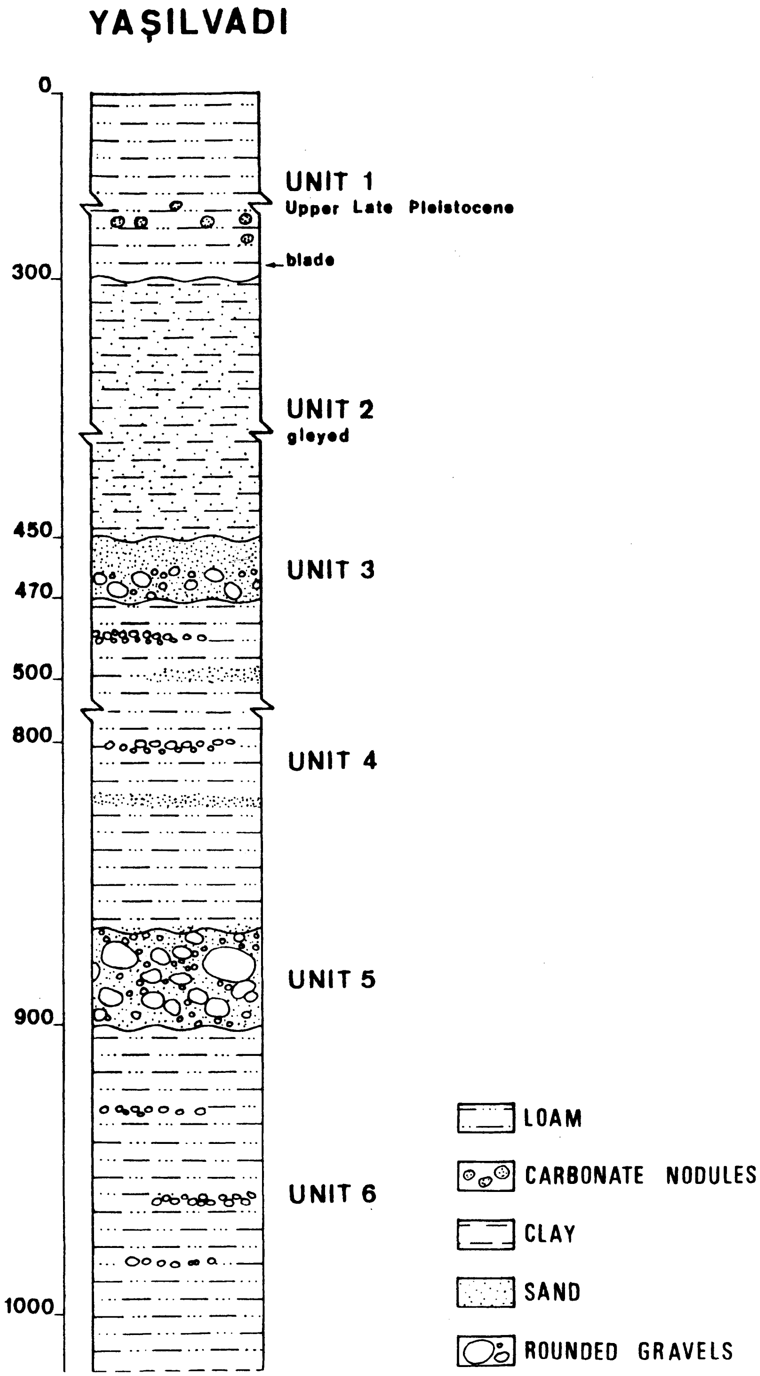

Figure 10.3. Schematic geological section of Pleistocene deposits in the Yaşilvadi Section. Scale in meters.

of the Holocene. The sediments from these exposures attest to three different environmental phases including: a) climatic stability under moist environmental regimes with well-developed clay-rich paleosols, b) dry episodes with major erosional cycles, and c) massive gravel deposition in fast-moving streams draining the hilly regions onto the Urfa Plain. Such a sequence is evident in the Büyük Kanal Section, located at the interface of the hills and the plain where streams rapidly loose competency and drop their bed load. The lowermost exposed unit (Unit 5) is a clayey overbank deposit overlain by a massive conglomerate (Unit 4). This conglomerate contained Levallois cores and flakes, and Mousterian period side scrapers. An endscraper with early Upper Paleolithic or Mousterian affinities was located at the top of the unit. It is therefore possible to assign a Mousterian-period date to this unit. There is a marked erosional break between Units 4 and 5 and the upper portion of the section. A tentative date for the sediments of these upper units (Units 3–1) is most likely Upper Paleolithic. Unit 3 represents a more gentle stream flow which eventually either ceased to flow or abandoned its bed in this location allowing a floodplain soil to develop in Units 2 and 1.

Further upstream at Yaşilvadi, the stream is bounded within a narrow valley. There is a different sequence of Late Pleistocene channel deposits and paleosols. The lowermost units (Units 3–6) are lightly cemented, gleyed stream deposits, some indicating fast-moving stream channels (Unit 5) but most are typical of slower, more steady flows with deposition in bars and levées. These indicate high water-tables and more moisture, but less erosive runoff. This stream subsequently became a small lake or pond as indicated by the fine-grained deposits and gleyed-mottled appearance of the Unit 2 sediments. The ponding was most likely caused by the blockage of the relatively narrow outlet from these hills onto the Urfa Plain to the south but also depended on a greater supply of water than is available in the area at present. The uppermost Unit 1 corresponds in lithology and stratigraphic position to the uppermost two units of the Büyük Kanal section. At Yaşilvadi, Unit 1 contains a flint blade within the section and a scattering of lithic debitage deflated from the top of the unit, which have affinities to Upper Paleolithic or Epipaleolithic assemblages. The two sections at Büyük Kanal and Yaşilvadi are equivalent to Wilkinson's Terrace III from the upper Euphrates valley (Wilkinson, 1990).

These and other sections in the vicinity attest to significantly moister Late Pleistocene climates than those prevalent throughout most of the Holocene. There were long periods of landscape stability with well-developed soils, competent stream flows, high water-tables, and periodic ponding. These landscape features imply a continuous ground cover of vegetation. All of this amounts to a rich environment for exploitation by hunter/gatherers, although few Mousterian or Upper Paleolithic sites have thus far been found in this region. One reason might be that sites in the hilly zones would have eroded away with the soil mantle, while sites on the plain would have been buried by Holocene alluvium.

10.3.2. EARLY HOLOCENE

Geomorphological and hydrological evidence for the Early Holocene is as yet identified in only a few locations on the Urfa Plain. In many localities, Mid-Holocene sediments rest directly on Late Pleistocene deposits, either conglomerates associated with Mousterian artifacts, or later paleosols. In all instances, the contact is markedly unconformable and the top of the Pleistocene units display pronounced erosion. This points to a distinct erosional phase, perhaps corresponding to the dry phase of the Terminal Pleistocene Younger Dryas episode.

A similar erosive event has been documented in the Northern Syrian desert (Courty, 1994). Although dated to the Early Holocene, it is attributed to a very intense and long-lived dry episode resulting in a reduction in the vegetation cover accompanied by a significant drop in water tables plus stream incision. Geomorphological work on the Konya basin of central Anatolia likewise attests to a dry alluvial plain in the Terminal Pleistocene, this also being attributed to a long dry phase (Roberts, 1982).

The pollen record of eastern Turkey is in the process of re-evaluation. The original pollen profiles of van Zeist and Woldring from Lake Van showed a strong dominance of non-arboreal pollen indicating dry environmental conditions from *ca.* 11,300 to 8,000 BP (uncalibrated) (van Zeist and Woldring, 1978). However, the dating of this dry episode is more likely to be earlier, again corresponding with the Terminal Pleistocene Younger Dryas. A new varve-dated core from Van combines pollen counts with data from Mg/Ca ratios and O^{18} values. It indicates a generally dry Late Pleistocene with intense aridity in the Terminal Pleistocene at 11,600–10,460 BP (*ca.* mid-10th to mid-9th millennia BC) (Wick *et al.*, 1997). The late Epipaleolithic and perhaps earliest Neolithic periods fall within this apparent dry phase. Few sites from these periods have as yet been recorded in southeastern Anatolia, but it is still unclear if this lack of archaeological data is a true indicator of settlement or if sites of these ages have been buried by more recent alluvial fills

Most of the early sites from this region date from the latter part of the Aceramic Neolithic (beginning in the mid-8th millennium BC) or the equivalent of the Levantine Pre-Pottery Neolithic B period (Yakar, 1994). The new data from Lake Van indicate that this was a moist episode in eastern Turkey beginning after 10,400 BP (*ca.* mid-9th millennium BC). An Early Holocene moist episode is also indicated by marshy alluvium on the Konya Plain in central Turkey dated to approximately 9000 calibrated BC (Roberts, 1997).

10.3.3. MID-HOLOCENE

On the Urfa Plain the sediment sections dated to this period consistently display a lithology suggestive of a moist hydrological regime. They are typically composed of heavy dark clays sometimes gleyed, though often with a signature of reworked carbonate nodules, presumably eroded from the Late Pleistocene paleosols during the previous dry phase. These deposits represent fine-grained overbank sediments and the microenvironments of seasonal marshes. The stream channel facies are typified by small well-sorted pebble deposits indicating small streams with well-sustained flows draining this marshland (Rosen, 1997a). Such sediments are widespread in the upper Urfa Plain; and similarly appear in the Euphrates basin, near the site of Titriş (Fig. 10.4, Table 10.3).

These sections are dated by the archaeological material found within, and in two instances this includes *in situ* archaeological remains. In section G-18 (Table 10.4, Fig. 10.5) from the site of Kazane Höyük, an alluvial fill (Units 3 and 4) has buried an *in situ* Pottery Neolithic structure with ceramics similar to the forms from the 'burnt village' (Level 6 or slightly earlier) at Sabi Abyad in Northern Syria (Bernbeck, 1997). Akkermans dates this period to the early 6th millennium BC (approximately 5900 BC calibrated) (Akkermans, 1993:111–115). If this alluvial fill is indicative of a moist phase, then a more humid episode might have begun some time shortly after this Pottery Neolithic occupation. However, active alluviation might have ceased before the the Halaf period occupation at Kazane in the 6th millennium BC, since Units 3 and 4 contain ceramics exclusively from the Pottery Neolithic period in spite of there being an Halaf settlement only a few tens of meters away (Bernbeck, 1997). All the same, further investigations are required before one can rule out the possibility that the alluviation was suspended due to a shift in the channel rather than a decrease in flow. In Unit 2 there is a change in lithology to fine silty clay alluvium with reworked $CaCO_3$ nodules and Chalcolithic and Early Bronze Age ceramics indicating renewed sedimentation in the fourth to the third millennium. Sedimentation ceased when the Early Bronze Age city wall blocked off natural alluvial deposition. The sequence in G-18 tentatively suggests a termination of intensive alluvial activity during the Early Chalcolithic Halaf period, followed by a renewal of moist conditions in the Mid-Late Chalcolithic periods and the Early Bronze Age. But since the early part of this sequence has as yet only been detected in this one section, it is difficult to draw definite conclusions about environmental change overall within that time span.

The upper portion of the Mid-Holocene fill has been better documented, as it appears in a number of sections in both the Urfa Plain and the Turkish Euphrates basin. Among them are two sections with *in situ* Late Chalcolithic deposits dated to 4000 BC (Kazane: GS-VIII, and the Esrüciye section, approximately 1 km south of Kazane); and several more with archaeological inclusions (Kazane: GS-XI, G-18, Unit 2; Titriş: GT-V) (Rosen, 1997a; Rosen and Goldberg in Algaze *et al.*, 1995). This fill is lithologically distinctive, being characterized by heavy dark brown clays containing numerous reworked $CaCO_3$ nodules derived from the Late Pleistocene paleosols. In most of the available sediment exposures, this fill lies directly above truncated Late Pleistocene paleosols. An example of this is in Section GS-VIII

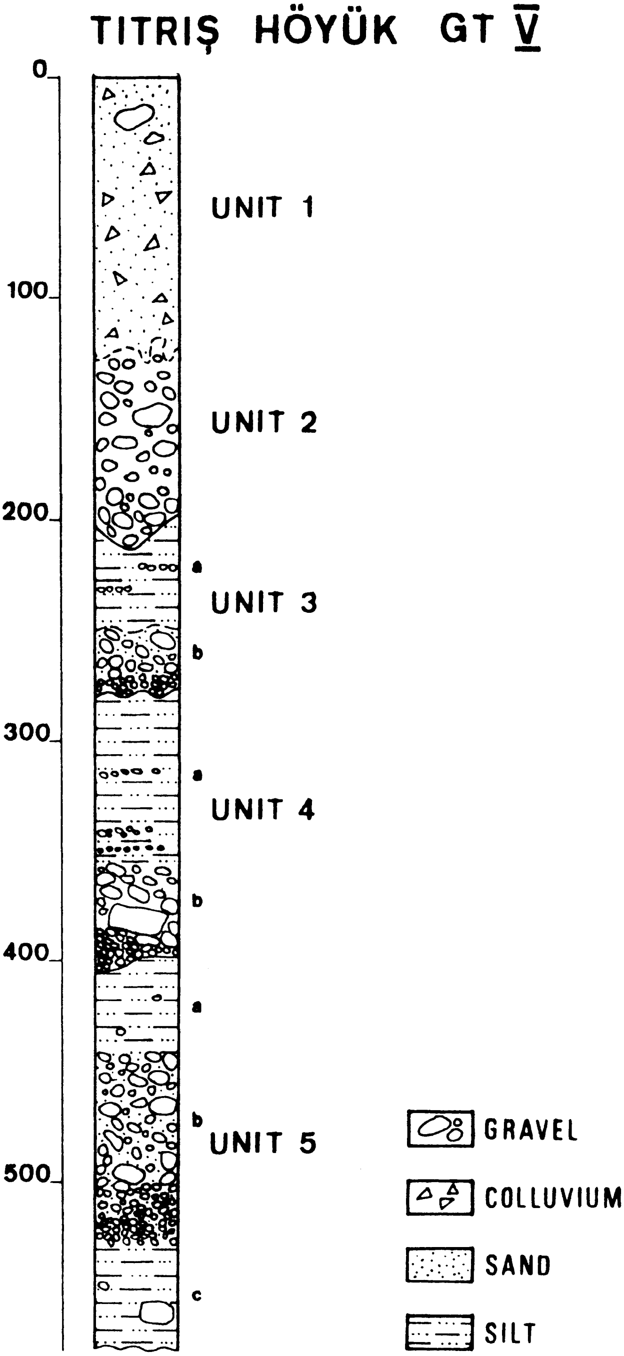

Figure 10.4. Schematic geological section of Mid-Holocene alluvium in GT-V from Titriş Höyük (Rosen and Goldberg in Algaze *et al.*, 1995). Scale in meters.

Table 10.3. Description of Mid-Holocene, Early Bronze Age fill from Section GT-V at Titriş Höyük, southeastern Turkey

Unit	Depth cm	Color	Texture	Structure	Inclusions	Boundary	Depositional Env.
1	0–130	Brown (10 YR 5/3, m)	Loam matrix with 30% diffuse angular and rounded gravels	Massive	Limestone, chalk and flint, sherds rootlets, krotovinas	Graded and wavy	Colluvial slope wash from tell
2	130–212	Light brownish gray (10 YR 6/2, m)	Poorly sorted, well-rounded discoidal gravels < 15 cm in diameter	Massive	Limestone and flint	Abrupt and irregular	Channel deposit from torrential flow
3a	212–254	Light brownish gray (10 YR 6/2, m)	Loam with occasional thin gravel lenses	Massive	Ca. 5% 3 cm pebbles, charcoal flecks	Graded and wavy	Fine-grained alluvium
3b	254–284	Light brownish gray (10 YR 6/2, m)	Very poorly sorted rounded and sub-angular gravels in sandy silt matrix; lowermost 10 cm is a lens of well-sorted 0.5–1 cm pebbles	Massive	Limestone, flint and marl with some EBA sherds	Abrupt and wavy	Channel deposit
4a	284–360	Light brownish gray (silts) (10 YR 6/2, m) and brown (sand) (10 YR 5/3, m)	Sets of thin 2 mm pebble lenses, silty sand and clayey silts	Graded bedding	Limestone, chalk, marl and EBA sherds	Abrupt and wavy	Channel deposit of well-sustained aggrading seasonal stream
4b	360–410	Light brownish gray (10 YR 6/2, m)	Moderately well-sorted angular, rounded, and discoidal gravels < 18 cm with lens of well-sorted 2 cm gravels at base	Graded bedding	Fe concretions, marl, limestone and flint	Abrupt and irregular	Channel deposit of high-energy flow
5a	410–444	Light yellowish brown (10 YR 6/4, m)	Clayey silt	Massive	Rare sherds; rare gravel	Abrupt and smooth	Fine-grained alluvial overbank deposit
5b	444–536	Light yellowish brown (10 YR 6/3, m)	Poorly sorted rounded and angular gravel (4 cm in diameter in mid-unit) and pebbles (0.5 cm in diameter) at base	Massive	Limestone, flint, marl, Fe concretions, EBA sherds	Abrupt and smooth	Channel deposit of high-energy flow
5c	536–580	Light brownish gray (10 YR 6/2, m)	Thin pebble lenses grading to sandy loam and clayey silt	Graded bedding	EBA sherds, 1% 5–8 cm gravels, one 20 cm cut block of limestone	Not visible	Channel & overbank deposits of well-sustained aggrading seasonal stream

Table 10.4. Description of sediment units from Square G-18, Kazane Höyük, Urfa Plain, southeastern Turkey

Unit	Depth cm	Color	Texture	Structure	Inclusions	Boundary	Depositional Env.
1	0–26/36	Yellowish brown (10 YR 4/4, m)	Silt loam	Massive	Ca. 5% sherds, 5% 3–5 cm pebbles, krotovinas, with lense of gravel and sherds at Unit 1/2 boundary	Graded and wavy	Post-Early Bronze Age colluvium
2	26/36–95	Dark grayish brown (10 YR 4/2, m)	Silty clay	Massive	Ca. 20% reworked $CaCO_3$ nodules, 10% 3 cm stones and sherds	Abrupt and smooth	Mid-Holocene (Chalcolithic-EB) fine-grained back-swamp alluvium
3	95–119	Yellowish brown (10 YR 5/4, m)	Loam	Massive	Burnt construction collapse, brick or pisé fragments, charcoal, sherds, concentration of small pebbles at base of unit	Very abrupt and wavy	Wash of burnt occupation debris
4	119–200	Dark brown (10 YR 4/3, m)	Clayey silt	Massive	Sherds, Charcoal, Bone, Flint, Fe stains	Abrupt and smooth	Fine-grained alluvium including PN occupation debris from nearby settlement
5	200–226	Dark brown (10 YR 4/3, m)	Clayey silt matrix	---	In situ PN structure with burnt clay surface or hearth and sherds	Not visible	Archaeological settlement deposit

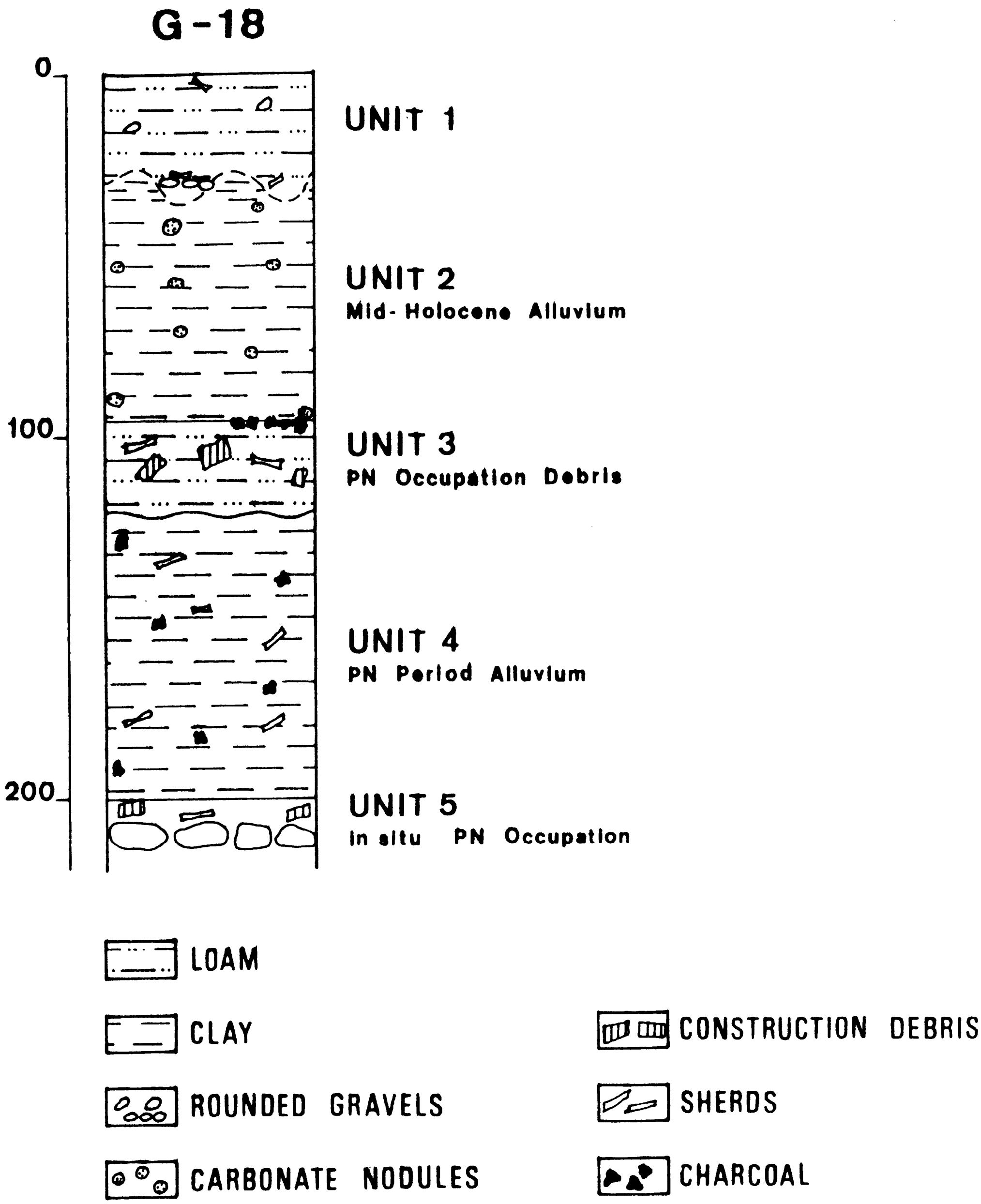

Figure 10.5. Schematic geological section of Mid-Holocene alluvium in G-18, Kazane Höyük. Scale in meters.

(Fig. 10.6, Table 10.5) which includes an *in situ* archaeological line of single stones and Late Chalcolithic sherds dating to *ca.* 4000 BC (H. Wright, personal communication) in the lowermost portion of the section, just 15 cm above the Late Pleistocene paleosol. There was no Pottery Neolithic period alluvium at this location, although it could have been eroded away before deposition of the Chalcolithic alluvium.

Section G-XI (Figure 10.6, Table 10.6) is another Mid-Holocene deposit which had an unconformable erosional contact with Late Pleistocene paleosols below. A Middle Chalcolithic sherd (*ca.* 5th millennium BC) was collected from the base of this section in Unit 3. This unit was characterized by well-sorted channel deposits, overlain by the marshy overbank sediments of Unit 2. The uppermost portion of this section contained third millennium BC Early Bronze Age sherds. There is a subsequent break in alluviation until the Late Classical and Medieval periods. It is quite significant that there have been no second millennium Middle Bronze Age sherds recovered from any of these sections. This strongly suggests that the alluvial hydrological regime had ended sometime within or at the end of the third millennium BC. With the second millennium there began a long phase of stream incision, probably related to a drop in water tables and decrease in stream flow.

10.4. Early to Mid-Holocene Landscapes and Environments

The above evidence suggests a scenario of climatic and hydrological change which we can tentatively compare to the still limited but ever-growing body of site settlement data. Geomorphological information indicates that the Late Pleistocene period fluctuated between very wet conditions in the Middle Paleolithic period to moist conditions in the Upper Paleolithic with long intervals of dry erosive phases. A dry episode in the Terminal Pleistocene appears to have been responsible for the truncation of Late Pleistocene paleosols and stream incision.

Tentative evidence for a renewal of alluvial activity in the Pottery Neolithic period (eartly 6th millennium BC), is afforded by Section G-18. However, this particular fill has been preserved only at this one location and therefore cannot be used to sustain a broad interpretation. Even if this section does represent increased stream activity, this appears to cease within the Halaf period in the 6th millennium BC. Holocene alluviation is renewed broadly during the Mid-Holocene (at the time of the Middle Chalcolithic); and it continues through the Early Bronze Age with possible minor breaks, but little channel erosion throughout. Lithologically this fill is very characteristic of sediments deposited in a moist microenvironment. The dark brown color of the fine-grained sediment and the signs of gleying point to marshy floodplains, while the well-sorted small channel gravels indicate small streams with well-sustained flows for at least part of the year. This moist situation continued until well into the Early Bronze Age (3rd millennium BC) but ended

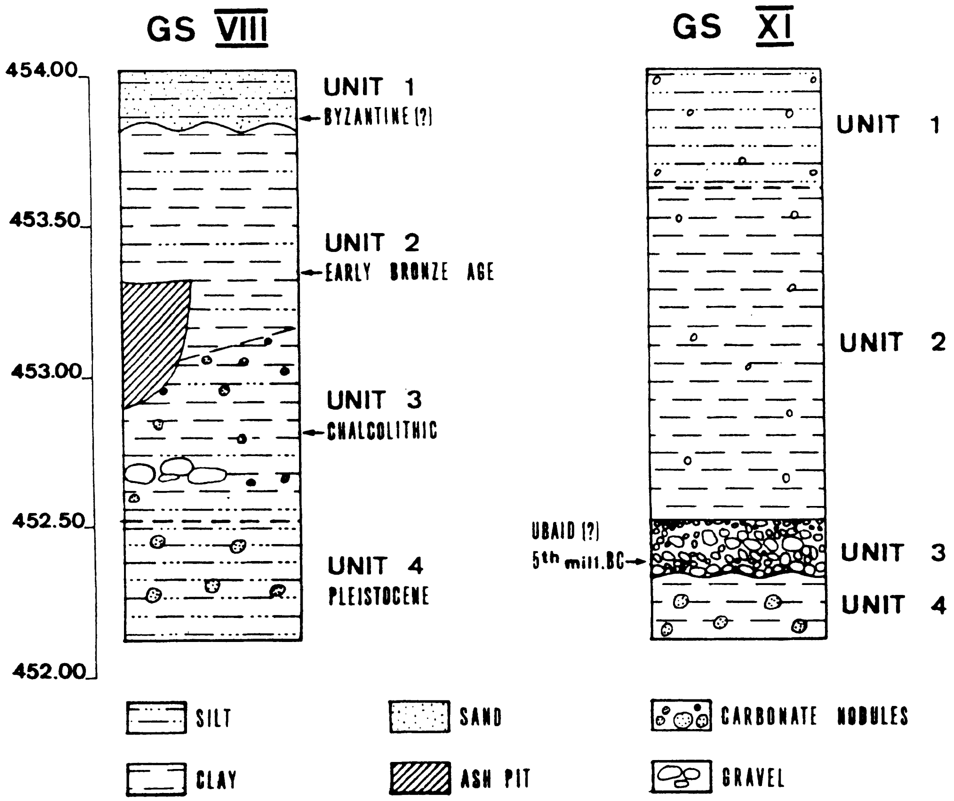

Figure 10.6. Schematic geological sections of Mid-Holocene channel and backswamp deposits in GS-VIII and GS-XI, Kazane Höyük (from Rosen, 1997). Scale in meters.

Table 10.5. Kazane Höyük Geological Section VIII (GS-VIII), Mid-Holocene (Chalcolithic/Early Bronze Age) flood plain (from Rosen, in press, Table 5)

Unit	Depth cm	Color	Texture	Structure	Inclusions	Boundary	Depositional Env.
1	0–20	Dark brown (10 YR 4/3, m)	Sandy silt loam	Massive	20% 5 mm–5 cm gravels, sherds, worked flint, charcoal, krotovinas	Wavy and abrupt	Colluvial wash
2	20–100	Dark brown (7.5 YR 3/4, m)	Silty clay	Massive	Large Chalcolithic and EBA sherds and a retouched flint blade, archaeological pit	Smooth and graded	Fine-grained flood plain deposit contemporary with archaeological material within
3	100–150	Dark brown (7.5 YR 4/4, m)	Silty clay	Massive	Sherds, exclusively Late Chalcolithic, anthropogenically laid stones	Smooth and graded	Fine-grained flood plain deposit which is contemporary with in situ archaeological material within
4	150–185	Dark brown (7.5 YR 4/4, m)	Clayey silt	Soil structure angular blocky	30 % 5–15 mm in situ $CaCO_3$ nodules, 10–20% Mn flecks	Not visible	Late Pleistocene paleosol (B/C-Horizon) on fine-grained flood plain alluvium

Table 10.6. Kazane Höyük Geological Section XI (GS-XI): Mid-Holocene channel, floodplain, and backswamp deposit (from Rosen, in press, Table 3)

Unit	Depth cm	Color	Texture	Structure	Inclusions	Boundary	Depositional Env.
1	0–40	Brown (7.5 YR 4/6, m)	Loam	Massive	5% 5 mm pebbles	Smooth and graded	Fine-grained flood plain deposit
2	40–150	Dark brown (7.5 YR 3/4, d)	Silty clay	Massive	5% 1–5 mm pebbles, some mn stains and gley	Smooth and semi-abrupt	Fine-grained flood plain backswamp deposit
3	150–170/200 (thickness varies)	Dark brown (7.5 YR 4/4, m)	Well-sorted 1–3 cm pebbles at base and 1–5 mm pebbles at top of unit; intebedded with mud lenses	Slight imbrication, graded bedding	Limestone, flint and basalt	Unconformable; wavy and abrupt	Medium energy stream channel
4	170–190	Brown (7.5 YR 5/4, m)	Silty clay	Massive	15% reworked 5 mm $CaCO_3$ nodules	Not visible	Fine-grained flood plain deposit

some time before or at the beginning of the second millennium Middle Bronze Age. No Middle Bronze Age artifacts have been found in these Mid-Holocene fills. From the Middle Bronze Age until Late Antiquity, there was no evidence of alluviation. Water tables dropped and the streams began to incise their beds, clearly indicating a dry phase with reduced rainfall and runoff.

Other paleoclimatic studies from Turkey and northern Syria corroborate portions of this scenario. Pollen diagrams from Lakes Van and Söğütlü show a very cool and arid phase in the Late Pleistocene (*ca.* 9600–8500 BC), marked by a dominance of non-arboreal pollen. This is followed by a moist amelioration (after *ca.* 8500 BC) with the movement of tree species into the area (van Zeist and Woldring, 1978; Wick *et al.*, 1997). A recently published ^{18}O and Mg/Ca record from Lake Van likewise shows cool/dry conditions in the Late Pleistocene between 12,000–10,460 BP (10th to mid-9th millennia BC, varve dates) and a Mid-Holocene moist phase from 8,200–5,050 BP (*ca.* 6200–3050 BC). The humidity slowly began to decrease after this phase until a very dry period set in from 4,190 BP–3,040 BP, or the end of the third millennium BC (Lemcke and Sturm, 1997).

Pedological and geomorphological work at a number of sites in northern Syria (Courty, 1994) yielded another paleoenvironmental sequence, dated primarily by archaeological finds, and presumably reported as calendrical dates. In this scenario, extensive alluvial deposits and soil formation in the Early Holocene signify a warm moist regime (Phase 1) before *ca.* 6000 BC, which might correspond to the Pottery Neolithic fill observed at Kazane Höyük. Phase 2 from 6000–5000 BC is interpreted as a dry episode in view of an influx of wind-blown sands. This arid phase appears to correspond with the beginning of the Halaf archaeological period, which may explain as well the apparent absence of Halaf-period alluvium on the northern Urfa plain. In Phase 3, dated to *ca.* 5000–3000 BC there is renewed alluvial buildup and soil formation attesting to a return to moist environmental conditions. This period includes archaeological assemblages from the end of the Early Chalcolithic (Halaf) until the Early Bronze Age. It is contemporary with the Mid-Holocene alluvium and backswamp deposits on the northern Urfa plain and in the vicinity of Titriş Höyük in the Turkish Euphrates valley. Phase 4 (3000–1800 BC) registers a slow shift to drier conditions with a decrease in alluvial sedimentation throughout much of the later Early Bronze Age. Finally Phase 5 (1800–1500 BC) was marked by a severe drought, probably corresponding to the stream incision recorded near Kazane and Titriş Höyüks.

Geomorphological work on the Konya plain in Central Anatolia has also resulted in a sequence of Holocene environmental events interpreted from records of sediment cores and raised beach deposits (Erol, 1978; Kuzucuoğlu *et al.*, 1997; Roberts, 1983). It has been summarized as follows by Kuzucuoğlu *et al.* (1997). From 23,000–17,000 BP (uncalibrated), there was a freshwater lake extending across the plain, typically 30 m deep. Between 17,000 and 13,000 BP (approximately 18200 to 13500 BC, calibrated), the lake receded leaving the plain completely dry until a moister phase from 12,000 to 11,000 BP (*ca.* 12000 to 11000 BC, cal.) in which a series of five shallow lakes developed. After 11,000 BP a dry episode returned in which these small lakes dried out and disappeared. This phase lasted until 7,000 BP (5800 BC, cal.). The Neolithic occupations of Can Hasan and Çatal Höyük took place at this time, ending with the renewal of moister conditions after 6700 BP (5600 BC, cal.).

Kuzucuoğlu's findings at the Karapınar Lake closely concur with this sequence: a complete drying of the lake from 17,000 to 13,800 BP (*ca.* 18200 to 14600 BC, cal.); a moist phase with freshwater marshes until 11,200 BP (*ca.* 11200 BC, cal.); a drier episode from 11,200 to 7,000 BP (*ca.* 11200 to 5800 BC cal.); a renewal of more humid conditions with an increase in marshes from 7,000 BP to 6,500 BP (*ca.* 5800 to 5400 BC, cal.); a short wet phase from 6,500 to 5,000 BP (*ca.* 5400 to 3800 BC, cal) when the lake returned; and then severe drought conditions tentatively dated by thermoluminescence from *ca.* 5,000 to 4,500 BP (Kuzucuoğlu *et al.*, 1997).

It is difficult to set precise dates to the beginning and end of wet, moist, and dry episodes from these data, and then in turn compare them with the archaeological evidence. However, it is possible to present a loose interpretation of a general climatic sequence as follows. A very wet period is roughly concurrent with the latter part of the Upper Paleolithic. There is a dry phase throughout much of the Terminal Pleistocene early and middle Mesolithic or Epipaleolithic period. Moist conditions return for a short interlude at the beginning of the Levantine Natufian period. Then a marked dry phase by the end of the Natufian period corresponds to the Younger Dryas cool/dry period recorded in many localities in the Near East (Moore and Hillman, 1992).

The evidence for the very earliest part of the Holocene appears to differ. The lake data from the Konya plain indicate a dry period throughout most of the aceramic and ceramic Neolithic, but geomorphological and pedological data from both northern Syria and the Urfa plain suggest a moist interval coinciding with at least part of the Ceramic Neolithic period. This also coincides with a return by the forest to eastern Anatolia. Clearly there is much need for further work on the Early Holocene. Geomorphological data from Kazane Höyük and northern Syria hint at a possible dry phase within the Halaf period with conditions becoming moister near the end of this period.

Lacustrine sequences from the Konya plain, ^{18}O and Mg/Ca values from Lake Van, and the alluvial sequences from southeastern Turkey and northern Syria all point to a significant Mid-Holocene cool/wet phase beginning just before or somewhat after the commencement of the Middle Chalcolithic Ubaid period. In this period lakes expanded in Central Anatolia, while marshes formed in valley bottoms in southeastern Turkey. Streams were actively building floodplains and water tables were higher than they are today. This amelioration extended into the Early Bronze Age. By the Middle Bronze Age (in the second millennium BC) a pronounced dry episode was well underway with water tables dropping substantially and streams beginning to incise their beds in northern Syria and southeastern Turkey. Lake Van became enriched in salts and ^{18}O values are high in sediments from this period.

10.5. Early to Mid-Holocene Environmental Change and Settlement

The late Epipaleolithic and Aceramic Neolithic in Southeastern Turkey is the northernmost manifestation of contemporary Levantine/Syrian Natufian and Pre-Pottery Neolithic A complexes which extend from the Sinai in the south to the Turkish Euphrates basin in the north. There are few sites from these periods. But two notable ones are Hallan Çemi Tepesi (late 11th millennium BC, cal.) in the Tigris basin (Rosenberg and Davis, 1992); and the lowest levels at Çayönü Tepesi (9th millennium BC, cal.) in the upper Tigris basin (Yakar, 1994). Both sites are characterized by round structures, and possible semi-permanent occupation. Although pulses such as lentils and bitter vetch were recovered from Hallan Çemi, there were no cereals and no evidence for any kind of cultivation. The earliest Neolithic occupation corresponds to the very end of the Younger Dryas arid episode characterized by dry steppe and parkland that would have included stands of wild cereals (Moore and Hillman, 1992).

The earliest full-fledged farming communities began to settle in southeastern Turkey in the later Aceramic (Levantine PPNB). These settlements were presumably more vulnerable to climatic fluctuations than the previous occupants because of their sedentism and reliance on cultivated cereals. They were established for the most part in river valleys on stream terraces or close to other permanent water sources (Yakar, 1991). In both the southern Levant and northern Syria, there is evidence for a wet episode corresponding to the PPNB settlement (Courty, 1994; Goldberg and Bar-Yosef, 1982), though this may not be the case in Anatolia. The Lake Van ^{18}O and Mg/Ca readings indicate a trend to moister conditions after the very dry Younger Dryas although it apparently still remained drier than the present (Lemcke and Sturm, 1997). Quite possibly though, the uncertainties of dry-farming were offset by some simple form of water manipulation or planting on moist alluvial soils.

The settlement pattern for the Neolithic period in Southeastern Turkey consists of dispersed communities in both the Aceramic and Ceramic phases. Aceramic sites in the region include eighth millennium BC Nevali Çori (Hauptman, cited by Yakar, 1994), Gritille (mid-8th/early 7th millennium BC) (Voigt, 1986), and Hayaz Höyük (mid-8th millennium BC) (Roodenberg, 1979–80). These sites have yielded evidence for domesticated flocks, primarily sheep and goats along with smaller numbers of pig and cattle. Also the inhabitants cultivated emmer together with einkorn wheat and lentils.

The Ceramic Neolithic site of Kumartepe is located on the Turkish Euphrates within Urfa province. It is 6 hectares in area although the population estimate at the site is relatively low. One radiocarbon date from the site fixes it within the 7th millennium BC (6743 cal. BC) (Roodenberg *et al.*, 1984; Wilkinson, 1990:87). Notably, the faunal remains from Kumartepe are dominated by domestic pig with a smaller proportion of sheep/goat and cattle. If, as tentatively suggested above, this period represents a moister phase within a gradually ameliorating climate, then we could expect there to be an increase in microenvironments suitable for the raising of pigs. The increase in forest cover and the spread of moist marshy habitats would lend themselves well to this pursuit.

The settlement patterns of the succeeding Early Chalcolithic or Halaf period (6th millennium BC) changed significantly from those of the previous Neolithic periods. The Halaf peoples expanded to the west and north of their centers on the north Mesopotamian plain. Although there were large Halaf centers in the region, the sites in Wilkinson's survey are small and more numerous than those of the preceding periods. They tend also to be situated in more diverse geographic localities, occurring on the high Euphrates terraces as well as the lower ones. Presumably they took advantage of 'wet point' locations where springs or water seeps were available, as well as ecotonal zones between woodland and steppe. These areas are suitable for both arable land and pasturage (Wilkinson, 1990:89–90). Whether or not these smaller dispersed settlements exploiting varied ecological zones represent an adaptation to a dry phase with reduced alluvial activity, is something which needs to be tested in future research in this region.

By the Middle Chalcolithic period (fifth millennium BC) environmental conditions were greatly improving. The Oak/Pistachio forests had recolonized the mountainous zones replacing the previous dry steppic vegetation. Surrounding the Urfa Plain and in the Turkish Euphrates zones, hillslopes were probably forested as well, as indicated by charcoal remains at Titriş and Kurban (Schlee in Algaze *et al.*, 1995; Miller in Marfoe *et al.*, 1986). Also the valley bottoms were aggrading with seasonal marshes beginning to form on the valley floors. It was a rich environment for the cultivation of cereals in moist alluvial soils as well as the raising of pigs, sheep, goats, and cattle. Yet survey and excavation data for this period indicate a decrease in population with fewer and smaller settlements than in the preceding Halaf period. The Euphrates valley sites moved away from the higher terraces and closer to the Euphrates floodplain (Wilkinson, 1990:91). At Kazane Höyük on the Urfa plain this period is known only from sherd scatters, and no architectural remains have yet been found (Wattenmaker and Mısır, 1993).

The Late Chalcolithic period (approximately spanning the fourth millennium BC) saw a major increase in settlements with a widespread expansion in towns ranging from 10–20 hectares. Some sites such as Tel Brak in northern Syria surpassed 50 hectares. This trend laid the foundations for the subsequent network of Early Bronze Age polities (Wilkinson, 1990). By this time the environmental conditions were optimal for farming and pastoral activities. Geomorphological evidence from tributaries to the Turkish Euphrates as well as from the Urfa Plain show that stream flow had increased and floodplains were aggrading with seasonal marshes and backswamps a prevalent part of the landscape. This means that prime agricultural soils were renewed with silts and clay in the winter seasons and remained moister throughout the growing season (Rosen and Goldberg in Algaze *et al.*, 1995; Rosen, 1997a). Such a situation would permit increased agricultural productivity for subsistence allowing surplus supplies for storage against drought years, as well as redistribution to non-farming craftspeople and administrators. At Kazane Höyük on the Urfa Plain, and Kurban Höyük on the Euphrates, the faunal remains were again dominated by pigs. But there were many cattle as well as sheep and goats (Nicola, 1995).

This period is also notable for the appearance in the north of goods in southern Mesopotamian Uruk styles, in some cases representing the movement of southern Mesopotamian populations, who established farming and perhaps trading colonies among local populations (Algaze, 1993). An environment rich in wood, arable land and pasturage at this time might have been one factor which encouraged foreign settlement in this region.

According to Wilkinson's Euphrates catchment survey data, the first Early Bronze Age hamlets in the early third millennium BC were not only dispersed, but small, no larger than one hectare in area. By the middle through to the end of the third millennium, some of these settlements developed into large polities dominant over a network of smaller satellite communities (Wilkinson, 1990). Some sites such as Kurban Höyük, Titriş Höyük and Kazane Höyük had grown to urban proportions supporting populations numbered in the thousands. These sites were located on the northernmost margins of the northern Mesopotamian sphere of influence, and were thus intimately connected to the Near Eastern world at large by political ties and international trade networks. Yet still, as in previous periods, their subsistence economies were based on locally produced agricultural products. Botanical remains from Kurban, Kazane, and Titriş indicate that these were mainly barley and emmer wheat presumably for home consumption. However, the abundance of grape seeds at Titriş also suggests a cottage industry of wine making, possibly as a trade item. Meanwhile abundant wood charcoal is important evidence for the continued existence of forests in the vicinity of these sites (Schlee in Algaze *et al.*, 1995; Albert and Rosen, n.d.).

The paleobotanical evidence for forested hillslopes in concert with the geomorphological indications of moist alluviating floodplains at Titriş (Rosen and Goldberg in Algaze *et al.*, 1995) and Kazane (Rosen, 1997a) throughout most of the Early Bronze Age suggest a climate not only moister than the present but probably with more evenly distributed rainfall and higher water tables. This, combined with opportunities for engaging in floodwater farming on second and third order streams, would have allowed higher grain yields each year than could be expected from an equivalent level of technology under present conditions. More importantly, though, is the stability in the yearly yield that the environmental conditions would have ensured. This stability can be critical for the maintenance of a stratified social system in which subsistence farmers also support a large number of non-producers.

At the end of the third millennium BC this Early Bronze Age network of large urban centers collapsed and many sites underwent major abandonments. These archaeological events coincide with much proxy evidence for a shift to a drier regime (see above). Not only does there appear to have been a major reduction of rainfall with droughts at this time, but the once alluviating streams reverted to a regime of downcutting and entrenchment of their beds, thus curtailing the possibilities for further floodwater farming. Diminished and probably more erratic winter rainfall acting in conjunction with stream incision would have seriously cut grain yields and put under enormous stress the agricultural base of the Early Bronze Age societies of southeastern Turkey and northern Syria. Some large sites were abandoned while others, such as Titriş Höyük suffered significant reductions in population. However, sites such as Kazane Höyük on the Urfa Plain and Tells Brak and Mozan in northern Syria seem to have been less affected by this environmental change. Within Wilkinson's survey area along the Turkish Euphrates, the large site of Kurban Höyük was abandoned for a short period, then resettled by a much smaller population. Although there may have been some reduction in the population of the region at large, the survey data indicate that once again there was an increase in dispersed smaller sites, along with at least one relatively large new village (Site 8) (Wilkinson, 1990:105). This settlement pattern represents a major shift in adaptation strategy. It might be attributable to the new environmental conditions, but was probably also related to shifting nodes of political control that may account for some of the variability in population responses at sites such as Kurban versus Kazane.

In previous works, it has been argued that severe climatic changes in antiquity do not lead automatically to the total collapse of settlements and civilizations; and that interesting questions arise from study of the various methods more resilient societies employ to cope with intervals of major agricultural stress (Rosen, 1995; 1997b). At present, there is not enough botanical, faunal and geomorphological data from the major and minor settlements in southeast Turkey and northern Syria to resolve these research problems effectively. Hopefully, however, future investigations at these sites will lead closer to an understanding of the interplay of social systems, technology, and environment which allowed some towns to continue to exist while others suffered collapse and abandonment.

10.6. General Inferences

The most significant element revealed by the comparison of settlement data with evidence for paleo-environment and climate change is that no simple correlation can be drawn between settlement evolution on the one hand and wet or dry trends on the other. In southeastern Turkey, large Aceramic and Ceramic period communities developed along floodplains at the very beginning of climatic amelioration, though not during periods appreciably wetter than the present. Conversely, moist environments during the mid-Holocene are left underexploited for hundreds of years within the middle Chalcolithic and earliest Early Bronze Age. Furthermore, late Early Bronze Age Kazane faced what must have been severe agricultural stress during the shift to a drier regime at the end of the third millennium BC. Yet it persisted, maintaining a relatively large population and complex social hierarchy well into the second millennium.

A reduction of human-environmental relations to a simple formula of wet=occupation/ dry=abandonment (Weiss *et al.*, 1993) may obscure important lessons about the multiplicity of human responses and adaptations to dynamic landscape and environmental change. One may thus overlook critical options such as technological flexibility, as in a switch from floodwater farming to canal irrigation; the adaptation of such social institutions as those for food redistribution; and flexible agricultural strategies, *e.g.* conversion to more drought resistant crops. Alterations in these and other factors can make the difference between the survival and dissolution for ancient agricultural communities under environmental stress.

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References

Akkermans, P.M.M.G. (1993) Villages in the Steppe: Late Neolithic Settlement and Subsistence in the Balikh Valley, Northern Syria. International Monographs in Prehistory, Ann Arbor.

Albert, R.M., and Rosen, A.M. (n.d.) Phytolith indicators of agriculture and paleoenvironments at Kazane Höyük, S.E. Turkey. (In preparation.)

Algaze, G. (1993) *The Uruk Expansion, World Systems and the Origins of Mesopotamian Civilization.* University of Chicago, Chicago.

Algaze, G., Goldberg, P., Honça, D., Matney, T., Rosen, A.M., Schlee, D., and Somers, L. (1995) Titriş Höyük, A small Early Bronze Age urban center in Southeastern Anatolia: The 1994 season. *Anatolica* **21**, 13–64.

Bernbeck, R. (1997) Unpublished summary of excavation trench G-18, Kazane Höyük.

Courty, M.-A. (1994) Le cadre paleogeographique des occupations humaines dans le bassin du Haut-Khabur (Syrie du Nord-Est). Premiers resultats. *Paléorient* **20**, 21–59.

Dewdney, J.C. (1971) *Turkey.* Chatto & Windus, London.

Erol, O. (1978) The Quaternary history of the lake basins of central and southern Anatolia. *In* W.C. Brice (ed.), *The Environmental History of the Near and Middle East since the Last Ice Age*,Academic Press, London, pp. 111–139.

Goldberg, P., and Bar-Yosef, O. (1982) Environmental and archaeological evidence for climatic change in the southern Levant and adjacent areas. *In* J.L. Bintliff and W. van Zeist (eds.), *Paleoclimates, Paleoenvironments and Human Communities in the Eastern Mediterranean Region in Later Prehistory*, British Archaeological Reports, Oxford, pp. 399–414.

Kuzucuoğlu, C., Karabiyikcoğlo, M., Fontugne M., Pastre J.-F., and Ercan T. (1997) Environmental changes in Holocene lacustrine sequences from Karapınar in the Konya Plain (Turkey). *In* H.N. Dalfes, G. Kukla and H. Weiss (eds.), *Third Millennium BC Climate Change and Old World Collapse*, Springer-Verlag, Berlin, pp. 451–463.

Lemcke, G., and Sturm, M. (1997) $\delta^{18}O$ and trace element measurements as proxy for the reconstruction of climate changes at Lake Van (Turkey): Preliminary results. *In* H.N. Dalfes, G. Kukla and H. Weiss (eds.), *op. cit.*, pp. 653–678.

Marfoe, L., Algaze, G., Ataman, K., Ingraham, I., McDonald, M., Miller, N., Snow, C., Stein, G., Verhaaren, B., Wattenmaker, P., Wilkinson, T.J., and Yener, K.A. (1986) The Chicago Euphrates Archaeological Project 1980–1984: An interim report. *Anatolica* 13, 37–148.

Moore, A.M.T., and Hillman, G.C. (1992) The Pleistocene to Holocene Transition and human economy in Southwest Asia: The impact of the Younger Dryas. *American Antiquity* 57, 482–494.

Nicola, J. (1995) Economic organization and local development in the Late Chalcolithic at Kazane Höyük, 3800–3100 BC. Paper on file, Department of Anthropology, University of Virginia, Charlottesville.

Roberts, N. (1982) A note on the geomorphological environment of Çatal Höyük, Turkey. *Journal of Archaeological Science* 9, 341–348.

——. (1983) Age, palaeoenvironments and climatic significance of Late Pleistocene Konya lake, Turkey. *Quaternary Research* 19, 154–171.

Roberts, N., Boyer, P., and Parish, R. (1997) Preliminary results of geoarchaeological investigations at Çatalhöyük. *In* I. Hodder (ed.), *On the Surface: Çatalhöyük 1993–95*, Ankara, British Institute of Archaeology at Ankara, pp. 19–40.

Roodenberg, J.J. (1979–80) Premier resultats des recherches archéologiques à Hayaz Höyük. *Anatolica* 7, 3–19.

Roodenberg, J.J., Wilkinson, T.J., and Bayri-Baykan, S. (1984) Survey and soundings at Kumartepe: An interim report. *Anatolica* 11, 1–16.

Rosen, A.M. (1995) The social response to environmental change in Early Bronze Age Canaan. *Journal of Anthropological Archaeology* 14, 26–44.

——. (1997a) The geoarchaeology of Holocene environments and land use at Kazane Höyük, S.E. Turkey. *Geoarchaeology* 12, 395–416.

——. (1997b) Environmental change and human adaptational failure at the end of the Early Bronze Age in the Southern Levant. *In* H.N. Dalfes, G. Kukla and H. Weiss (Eds.), *Third Millennium BC Climate Change and Old World Collapse*, Springer-Verlag, Heidelberg, pp. 25–38.

Rosenberg, M., and Davis, M.K. (1992) Hallan Çemi Tepesi, an early Aceramic Neolithic site in eastern Anatolia: Some preliminary observations concerning material culture. *Anatolica* 18, 1–18.

Tolun, N. (1975) *The Geological Map of Turkey, 1:500,000.* Ankara, Maden Tetkik ve Arama Enstitüsü Yayinlarindan.

van Zeist, W., and Woldring, H. (1978) A postglacial pollen diagram from Lake Van in east Anatolia. *Review of Palaeobotany and Palynology* 26, 249–276.

Voigt, M.M. (1986) Excavations at Gritille. *Anatolica* 15, 215–227.

Wattenmaker, P., and Mısır, A. (1993) Kazane Höyük—1992. *Kazı Sonuçları Toplantısı* 5, 177–187.

Weiss, H., Courty, M.-A., Wetterstrom, W., Guichard, F., Senior, L., Meadow, R., and Curnow, A. (1993) The genesis and collapse of third millennium North Mesopotamian civilization. *Science* **261**, 995–1004.

Wick, L., Lemcke, G., and Sturm, M. (1997) Vegetational response to Late-Glacial and Early-Holocene climatic changes in eastern Turkey: A high-resolution pollen record from Lake Van. Paper presented at INQUA, Ankara, April 1997.

Wilkinson, T.J. (1990) *Town and Country in Southeastern Anatolia: Settlement and Land Use at Kurban Höyük and Other Sites in the Lower Karababa Basin.* The University of Chicago, Chicago.

Yakar, J. (1991) P*rehistoric Anatolia: The Neolithic Transformation and the Early Chalcolithic Period.* Tel-Aviv University, Tel-Aviv.

——. (1994) Supplement No. 1, *Prehistoric Anatolia: The Neolithic Transformation and the Early Chalcolithic Period.* Tel-Aviv University, Tel-Aviv.

Chapter 11

Some Considerations on Climatic Changes, Water Resources and Water Needs in the Italian Region South of 43°N

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Nihil sub sole novum

11.1. Introduction

Whatever their causes might be, climatic changes inexorably affect the water cycle. Along with trends in mean values in all aspects of the weather (precipitation, temperature, evaporation, *etc.*), one can expect variations in the frequency of exceptional events, and also in the hydrological regime. There may be major consequences in terms of economy and quality of life. This has often occurred in the past, and it would be strange if does not happen again in the future. Besides the climate's natural variability, there are good reasons to think that man's activities these last two hundred years are leading towards a considerable rise in air temperature, the well-known anthropogenic greenhouse effect. So it is important to attempt to define in advance the environmental scenarios which can reasonably be expected to take place.

This study, which is essentially concerned with the part of Italy which is between 37° and 43°N (Fig. 11.1), presents information regarding variations in climate and water yield starting from about 1000 BC, and also makes an analysis of the current situation. Certain hypotheses are made regarding the average yearly water yield to be expected over the next 55 years (*i.e.* for the time interval 1996–2050) and a set of reasonable future scenarios are constructed, using the temporal analog approach (*cf.* Arnell, 1995:393–396).

Table 11.1 gives the meanings of the acronyms used in the text. The locations of the places mentioned are given in Figure 11.1

A. S. Issar and N. Brown (eds.), Water, Environment and Society in Times of Climatic Change, 241–271.

Table 11.1. Acronyms and abbreviations used in the text

AOGCM	–	Atmosphere-Ocean General Circulation Model
CLIMAP	–	Climate Long-Range Investigation Mapping and Prediction
CNR	–	Consiglio Nazionale delle Ricerche (Italian National Research Council)
GCM	–	General Circulation Model
IPCC	–	International Panel on Climatic Change
LIA	–	Little Ice Age
MPI	–	Max Planck Institute for Meteorology (Germany)
s.l.	–	Level of significance of linear fit according to the F test
WMP	–	Warm Mediaeval Period

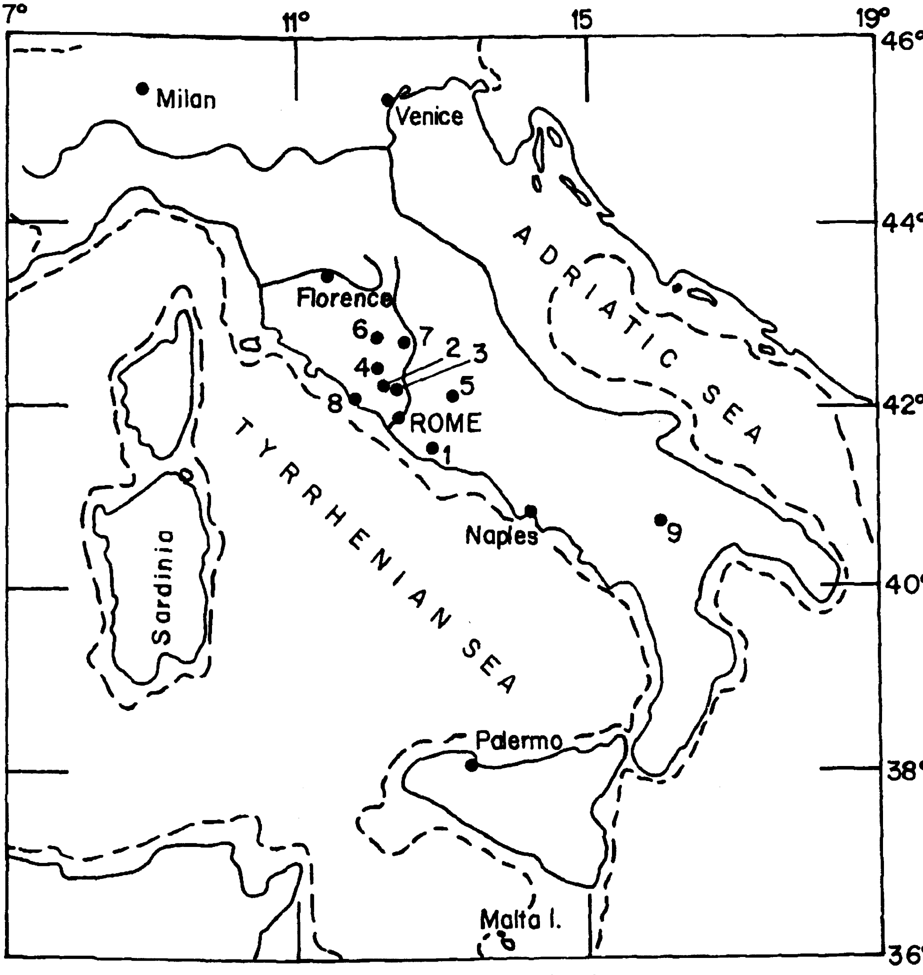


Figure 11.1. Map of Italy and locations of places mentioned in the text.
Lakes: 1) Castel Gandolfo; 2) Bracciano; 3) Martignano; 4) Bolsena and Mezzano; 5) Fucino. Meteorological stations: 7) Perugia; 8) Civitavecchia; 9) Potenza. The hatched line represents approximately the coast line during the last maximum glacial phase.

11.2. The Situation between 1000 BC and the Present

11.2.1. INTRODUCTION

In this paper climatic conditions in different periods of time are expressed in terms of wet or dry regimes (*i.e.* high or low water yield). The information presented here has been checked as thoroughly as possible, and should be considered to be the best data when differing from that presented elsewhere (for instance Dragoni, 1996). In the original papers, where dating was done in the traditional archaeological way (Late Bronze Age, First Iron Age, *etc.*), an absolute dating was attempted according to recent published comparative tables (Peroni, 1994:214–215).

Here information is given only on average lake levels for time spans longer than 50 years or so. Other sets of data reflecting the climatic situation will not be examined in any detail: for instance, the frequency of floods for a number of Italian rivers starting from the 4th century BC will not be considered (*cf.* for instance Natoni, 1944; Le Gall 1953; Lamb, 1972:427; Pavese *et al.*, 1992). This choice was based on the opinion that, although floods are connected with climatic conditions, there is also a very strong random element to them which could distort our perception of the 'average' climatic situation of the period during which such an event took place. It must also be pointed out that the reports which have reached us regarding floods almost never give information on contingent factors which may affect the course of the events. For example, a 'normal' flood may cause very serious damage (and thus be recorded as a great flood) due to incidental circumstances, such as the poor state of repair of embankments or the chance obstruction of the river bed. Moreover, particularly in the more remote past and/or in times of war or famine, it was always possible that the local occurrence of a great flood passed unrecorded or that records were destroyed. This implies that flood data from ancient times can be considered tolerably reliable only when a high frequency of flooding with very high water levels is reported for one or more periods (*i.e.* great floods regardless of incidental situations) and when these periods are more or less synchronous for two or more rivers. It appears that such a situation has been documented in Italy only for the flooding of various rivers in the second half of the 15th century (Camuffo and Enzi, 1995:448–449). The same considerations are valid also for the frequency of landslide movements and other natural hazards, as shown by Grove (1985) in Norway for the centuries between 1500 and 1900 AD (in Del Prete, 1993:13).

Table 11.2 reports some information about the lakes mentioned in the text.

11.2.2. LEVELS OF THE LAKES IN CENTRAL ITALY

11.2.2.1. *Castel Gandolfo Lake*

Castel Gandolfo Lake (some times called also Albano Lake) is located close to Rome and is of volcanic origin. The lake has no natural outlet so, without man's interven-

Table 11.2. Approximate data regarding the lakes mentioned in the text

Lake	Water body	Area (km^2) basin	Elevation (m a.s.l.)	Comments
Bracciano	57	149	164	natural outlet
Bolsena	115	159	305	natural outlet
Castel Gandolfo	6	10	293	no natural outlet, one artificial
Fucino	150	840	670	Roman outlet/the lake was drained in 1875
Martignano	2.5	6	207	Roman outlet
Mezzano	0.5	1	452	natural outlet
Trasimeno	125	309	257	no natural outlet, at least 2 artificial outlets

tion, its level would be entirely dependent on the climatic situation. Presumably in order to control the maximum levels of the lake, and for irrigation during the dry season, the people living in the area dug an outlet tunnel, which can be dated at around 400 BC (Castellani and Dragoni, 1991:58). According to Guidi (1986:241), at a depth of 8–10 m below the average present level of the lake (and below the bottom of the tunnel), there are remnants of Late Bronze settlement (1200–1000 BC). Given that the tunnel was designed to carry water with a free surface, and that no pumps were available, it must be supposed that the elevation of the tunnel bottom is several meters lower than what the lake's average level was at the time it was made. This means that between 1000 and 400 BC the average level of the lake rose many meters. At the moment there is no other information regarding the level of Castel Gandolfo Lake. However it is known that, in the last few centuries, water was flowing through the Roman tunnel (it was used to run some water-mills). In the last 10 years, the level of the lake has dropped dramatically after a period with low rain and increased pumping for water supply. At present the level is below the bottom of the Roman tunnel. During the LIA the level was higher than it was around 1000 BC.

11.2.2.2. *Bracciano Lake*

Bracciano Lake is a large lake located north of Rome. At least two prehistoric villages are well below the present level of the lake (Fugazzola-Delpino and Lombardi, 1982a:228–230). The first (Vicarello) has provided material belonging to the Ancient Bronze Age (2300–1700 BC). The second (Vigna Grande), located about 60–70 meters from the present-day shore, has provided material from the Middle Bronze Age and the Iron Age, the latter being dated around 1000–900 BC.

11.2.2.3. *Martignano Lake*

Martignano Lake is a small lake, located north of Rome. In the year 2 BC an aqueduct was built to carry water to Rome (Aqua Alsietina). The aqueduct begins with a tunnel, which today is about 12 meters above the surface of the water. As with Castel Gandolfo Lake, it must be supposed that the elevation of the bottom of the tunnel is at least 1–2 meters below the average lake level at the time of the digging of the tunnel. At a depth between 20 and 30 meters below the bottom of the Roman tunnel (and more than 15 meters below the present level), there are remnants of trees on the shore of an old coastline. Carbon dating tells us that the trees were alive between 200 AD and 500 AD (Mocchegiani-Carpano, 1976, in Liberati-Silverio, 1986:74). This means that between 2 BC and about 200 AD, the level of the lake dropped approximately 30 meters, with a shift towards a drier climate. This regime persisted at least until 500 AD. In order to reclaim some land, at the beginning of the last century some hydraulic works were constructed in the area. Therefore the relationship between the average level of the lake in 2 BC and the present average level cannot be attributed to climatic variations alone.

11.2.2.4. *Bolsena Lake and Mezzano Lake*

Bolsena Lake is located north of Rome, in a volcanic caldera; and the small Mezzano Lake is found nearby. Extensive archaeological investigations around Bolsena Lake, begun by A. Fioravanti in 1959 (*cf.* Tamburini, 1986), revealed important prehistoric and historic settlements and related remains that are today under the water of the lake. A series of absolute calibrated datings have been carried out by a team from La Sapienza University of Rome and the CNR during the last 20 years. The results of this important work are summarized in a paper by Belluomini *et al.* (1993), and are as follows:

a) Between about 1150 and 800 BC, the level of Bolsena Lake was about 3.5 m lower than today. The large number of samples covering this time span indicates that, during this interval, there were no variations in the lake level.

b) At about 500 AD, the level was 6 m lower than the present level.

c) Around 1000 BC, the level of Mezzano Lake was a few meters lower than today.

Samples of wood stakes, taken from different spots on the Bolsena lake bed (below the present lake level), are from a period extending roughly from 1050 to 1640 AD. However, these samples do not indicate a lower level of the lake. By about 1000 AD in Bolsena Lake, as in most other Italian lakes, a fishing technique was used which involved putting wood fascines into the water. These served as a shelter and a pasture for the fish, which crowded into the zone around the fascines and could be caught easily. The actual fishing was done by means of nets around the fascines; the nets were tied to wooden poles driven into the soft bottom of the lake, maybe as much as 10 or 12 meters below the surface of the water (*cf.* Gambini, 1996).

11.2.2.5. *Lake Fucino*

The large Lake Fucino was entirely drained in 1876. Apart from a few small karstic sinks, Lake Fucino had no natural outlet.

Around 1000–900 BC, the level of the lake was a few meters lower than it was between the 1st century BC and 52 AD. This is proved by some proto-Villanovian settlements located below the lake's surface in Roman times and, indeed, below the average level during the last 5–6 centuries (Irti, 1991:77, 80).

In order to gain land for farming, in the year 52 AD a Roman underground outlet (planned in the 1st century BC by Julius Caesar) was completed on the western shore of Lake Fucino. The median shoreline of the lake during this time is known from the remnants of this old Roman outlet as well as from geomorphological evidence (Giraudi, 1994:20–21). However, the limited technical capabilities in ancient times made practically impossible the digging of tunnels with sections and flow capacities sufficient for controlling the levels of lakes with basins and areas of this size. This implies that, even after the building of the Roman tunnel, the fluctuations in the levels of Lake Fucino were due essentially to natural causes (Castellani and Dragoni, 1991). Roman historians recount that the tunnel did not work well and attribute this, with some reason, to poor execution of the work. Later on (around the end of the 1st century AD and the first half of the 2nd century AD), the tunnel was reportedly restored by Emperors Trajan and Hadrian. The latter lowered the level of the intake gate, and new farmland was reclaimed. As mentioned earlier, the lowering could not have been caused by water drained out by the tunnel. So it must be concluded that it was caused mainly by a 'dry' period.

According to D'Amato (1980:181), the level remained low until the 4th century AD. However, other authors report that the Roman outlet was working (*i.e.* the lake was low) until the 6th century AD, though no evidence is given in support of this opinion (D'Amato, 1980:182). For certain in 782 AD the lake was at approximately the same level as it was in the beginning of the 1st century AD. This is proven by a document reporting the donation by Ildebrando, duke of Spoleto, of property in the county of the Fucino to the Montecassino monastery. This document mentions the donation of land together with "…. some fishermen of Lake Fucino, with the harbor named Adrestina". D'Amato (1980:187) noticed that the locality Adrestina should correspond to the place today known as Arestina, located at about the same level as that in the beginning of the 1st century AD. This observation rules out the opinion (founded on rather vague reasons impossible to check) of Brisse and De Rotroù (1876:238), according to whom the level of the lake was low during the 7th, 8th and 9th century AD.

According to Muratori (in D'Amato, 1980:41), in 1177 AD there was an exceptional rise in the water level. This suggests that during the so-called Warm Mediaeval Period the level of the lake was low, and that around 1177 AD the climate was beginning to shift again towards wetter conditions. In 1239–1240, Emperor Frederick II Hohenstaufen ordered the Roman outlet be put back into operation so as to lower the lake level, but the attempt was a failure. This order by the Emperor, recorded in Brisse and De Rotroù (1876:240–241), suggests that the people living on the shore of the lake were complaining about a recent rise in the lake level. which should have occurred not many years before 1239 AD. Since then, the level of the lake remained on the average high until it was drained in the last century. During the last three centuries, some variations in the lake level are very well documented. These can be correlated with periods of low/high rain locally or else with great volcanic eruptions, such as those of Tambora and Krakatoa (Giraudi, 1994:30; Dragoni, 1996:203).

11.2.2.6. *Lake Trasimeno*

The large Lake Trasimeno is a closed lake, one with no natural outlets. An underground outlet was probably dug by the Etruscans or the Romans, but the date of any such construction is unknown (Dragoni, 1982:193; Gambini, 1995:60–61). At the end of the last century an outlet tunnel was built to control the highest rises of the lake. For this reason, the lake level is at least 2.5 meters lower today than the average for the last four to five centuries (Dragoni, 1982; Gambini, 1995).

Many ancient underwater remains have been found and described. However, the most ancient (pre-Middle Age) have not yet been studied in sufficient detail. They are generally considered to be from the Etruscan-Roman period. But since that lasted for many centuries, they cannot be used to deduce old water levels with acceptable approximation. During the 1980s the remnants of a road of unknown age, but a few centuries at the very least, were found underwater. In March 1994 some ceramic remains from a settlement dated 15th–11th century BC were found; the dating was done on an archaeological basis (personal communication by G. Carancini, who studied the remains). These remains were a few meters lower than the natural level of the lake during the last century. So we must conclude that around 15th–11th century BC the lake level was very generally lower than during the 13th–19th centuries AD.

It appears that during most of the 12th and 13th century AD, the area of the lake was smaller than it was after 1300–1400 AD. According to Cialini (1991:7–9), this is confirmed by old documents giving the names of villages which later disappeared (presumably under water), and by the amount of wheat produced annually by farming land which is now flooded. All this is supported by the presence of medieval ruins, today under water, though still visible in exceptionally dry periods in the last century. It is certain that in 1274–1280 AD the Monastery of San Martino della Vena stood in an area which during the 15th century and after was very often under water, in locations perhaps 150–200 meters from the present-day mean shoreline: its remains were described in the 1950s and 1970s (Gambini, 1995:71–73). Between 1300 and 1400, frequent flooding (and a corresponding rise in the average lake level) had devastating effects on the area's economy, and around 1420 an underground outlet was dug in a fruitless effort to control the level. Around 1490 some streams that emptied into the lake were diverted to nearby Lake Chiusi (Dragoni, 1982:193). This attempt also failed. So the fluctuating lake level remained higher than it was in MWP or is today. There were particularly high levels in 1602, between 1762 and 1773, and between 1810 and 1820, the latter two phases coinciding with similar variations in the level of Lake Fucino and with particularly cold periods recorded throughout Europe (Lamb, 1977:468; Dragoni, 1996:204).

Fig. 11.2 integrates the above information. It appears that, to an approximation of one hundred years and simply in terms of 'high' and 'low' level, the positions of the levels were synchronous. There is abundant evidence (sedimentological, archaeological and historical) that the periods with 'high level' and those with 'low level' correspond respectively to periods 'cool and more rainy' and 'warm and less rainy', not only to 'cool' and 'warm' phases. In addition to the well known works of Lamb (1972, 1977, 1985), a recent review paper on the subject, covering roughly the interval 1100 BC–800 AD has been presented by Veggiani (1994). During the LIA an increase of precipitation, compared to the previous WMP, is also generally accepted in the area here considered. At present, for the lakes in central Italy, the

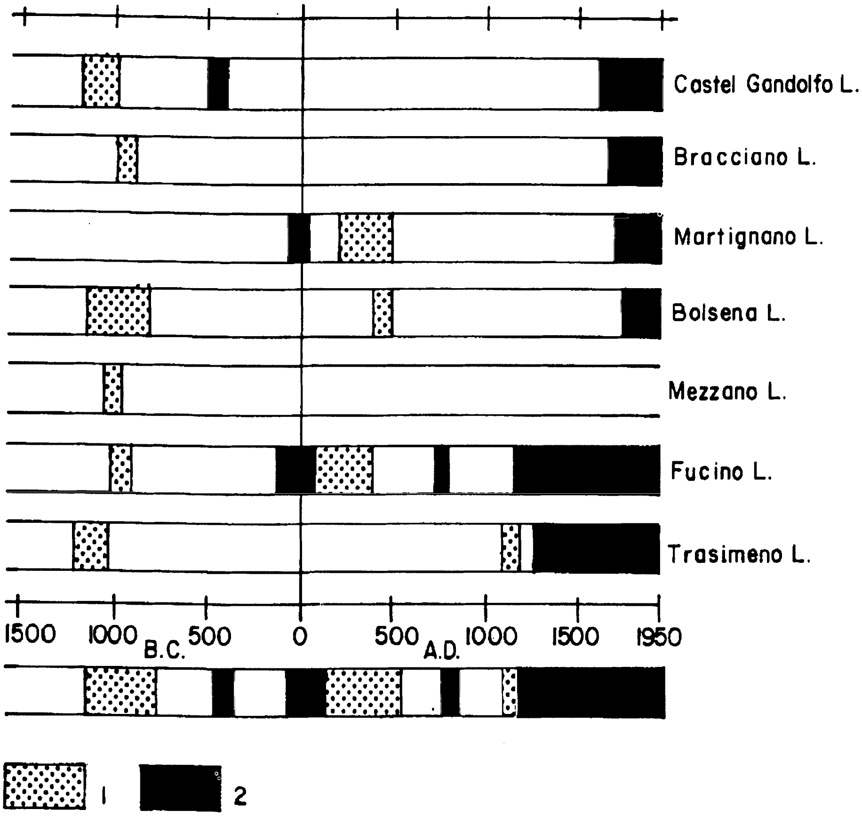


Figure 11.2. Comparison among the information obtained from the levels of lakes in central Italy: 1) low level; 2) high level. The bottom line represents all the information combined.

level variations are strongly linked to the yearly rainfall. The link with the mean annual temperature is weaker (Dragoni, 1982; Rufini, 1989).

11.2.3. LEVEL OF LAKES IN CENTRAL ITALY AND OTHER PROXY DATA

Fig. 11.3 compares the information given above to some other climatic proxy data sets, taken from several sources (Dansgaard *et al.*, 1971; Thompson and Mosley-Thompson, 1992; Mintzer, 1992; Whyte, 1995; Pinna, 1996).

When looking at Fig. 11.3, it should be kept in mind that (aside from the isotopic record) the dating approximation is around 100 years. The comparison with the other proxy information shows that warm periods correspond to a large extent to low lake levels (situation warm = dry = low rainfall), while cool periods correspond to high lake levels (situation cool = wet = high rainfall). *Sensu stricto,* this conclusion is valid only for the area were the lakes considered above are located, *i.e.* the western part of central Italy. However, using a sedimentological and archaeological approach, similar findings (cool = wet and warm = dry) were obtained recently for southern Italy and Sicily (Ortolani and Pagliuca, 1994; Giano and Guarino, 1996). To a large extent, the cool and warm phases are respectively synchronous in both areas. This correspondence is supported by some recent works which demonstrate that, at least at present, the Mediterranean area considered here is one region climatically (Conte and Giuffrida, 1991; Conte *et al.*, 1991; Balafoutis and Arseni-Papadimitriou, 1996).

11.3. Present Situation and Foreseeable Scenarios

11.3.1. INTRODUCTION

According to the great majority of authors who have studied the climatic fluctuations in this area, during the last 100–120 years there has generally been a decrease in rainfall and a slight increase in the average temperature (Melicchia 1942:78; Giuffrida and Conte, 1991:341; Rapetti and Vittorini, 1991:472–473; Conte, 1994:34,36,39; Mangianti and Beltrano, 1995:42; Dragoni, 1996; Montanari *et al.*, 1996). That is consistent with a very recent and statistically accurate analysis carried out by Piervitali *et al.* (1997a, 1997b) on 60 different rainfall stations across the N.W. Mediterranean in the time span 1951–1995. These authors observed in general an average decrease in rainfall of about 15–20% and an increase in temperature over the last 45 years. As far as I know, papers which give a different result are based on short data sets (for temperature: for example, Jones, 1994, considered only the period 1951–1980). Here I examined the rainfall and temperatures recorded at the stations in Perugia, Rome, Potenza and Palermo (Fig. 11.1). These data sets were chosen because they cover a large area, are rather extended in time

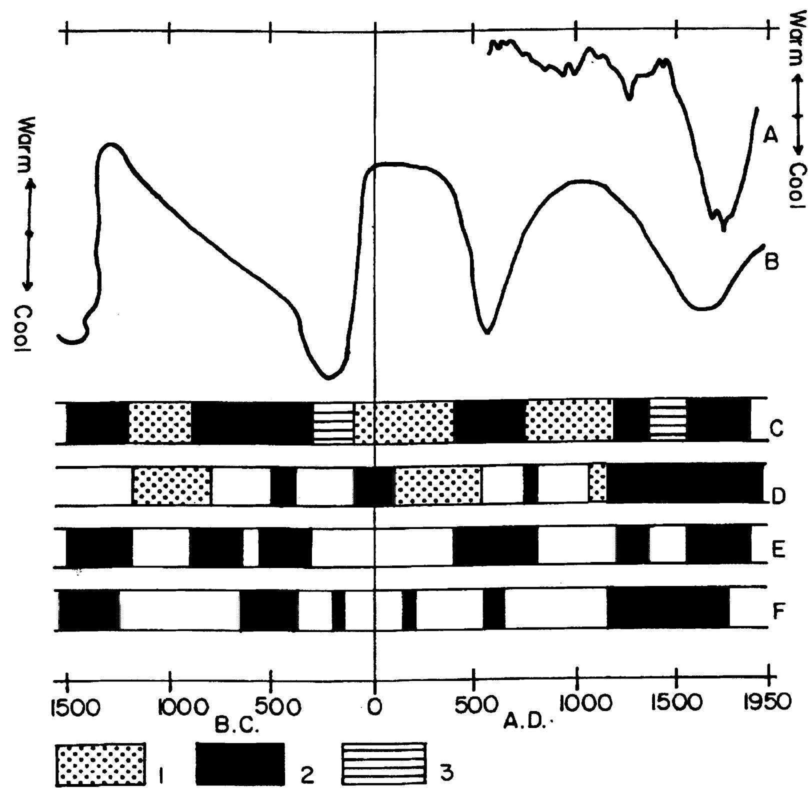


Figure 11.3. Comparison among different palaeoclimatic data sets.

A – Isotopic record ($\delta^{18}O$) from the Quelccaya Ice Cap, Perù, 200 years moving average (computed from the data given in Thompson and Mosley-Thompson, 1992).

B – Isotopic record ($\delta^{18}O$) from Camp Century, Greenland (redrawn after Dansgaard *et al.*, 1971, as reported in Rein and Negendank, 1993:165).

C – Climatic changes in western-southern Europe, according to a general review of existing proxy data due to Pinna (1996:118).

D – Situation in central Italy, according to the information on lake levels in central Italy reported in this paper.

E – Advances of Fernau Glacier in the Alps, after Pinna (1996:119)/

F – Advances of Jostedalsbreen ice cap in Norway, after Mintzer, 1992 (in Whyte, 1995:44).

Meaning of the symbols:

A and B (isotopic record): the isotopic records are not on the same scale: they must be seen only in qualitative terms.

C (Pinna's data): 1 = warm, 2 = cool, 3 = transition. The term 'transition' groups together the terms 'mild' and 'fresh' used in the original paper by Pinna (mite and fresco in Italian).

D (lake levels): 1 = dry, 2 = wet. Blank areas mean 'no information'.

E and F (Glaciers): 1 = major advances; blank areas = minor oscillation, or stable, or retreat.

and include both temperature and rainfall. For Perugia, Rome and Potenza, monthly values were available; for Palermo, only yearly values.

In the Perugia set there are no missing data, though some are missing from the other sets. Whenever possible, the missing data have been reconstructed through multiple regression with the closest stations. In the data set of Rome the temperature data from 1951 to 1960 and from 1989 to 1992 are missing, and have not been reconstructed as non reliable methods were found. Altogether Perugia, Rome and Potenza data seem to be highly reliable, but there is some uncertainty regarding that from Palermo. However, as the set for Palermo results to be the longest and most continuous in all of Sicily, it has also been taken into consideration, keeping in mind that the indications furnished by it must for the moment be utilized with due caution.

11.3.2. TRENDS DURING THE 20TH CENTURY

The data sets were analyzed by simply fitting a linear trend, the significance of which was tested using the F test (Tables 11.3., 11.4; Figs. 11.4, 11.5, 11.6, 11.7). Only trends having an s.l. ≤ 0.1 have been considered to be significant.

The fits between the average yearly rainfall and the average yearly temperature are shown in Figs. 11.8, 11.9, 11.10, and 11.11. In all the stations but Palermo there is a significant negative linear relation. This suggests that in this region if the temperature should increase, very probably the rainfall will decrease. These results are consistent with some earlier findings (Dragoni, 1996) and with a very recent work (Dragoni and Valigi, 1998).

Table 11.3. Average yearly rainfall and gradient for the stations considered

Station	Period (years)	Average (mm)	Gradient (mm y^{-1})
Perugia	1900–1995	868	−2.47
Rome	1880–1995	784	−2.67
Potenza	1926–1995	778	−2.82
Palermo	1936–1992	589	−1.83

Table 11.4. Average yearly temperature and gradient for the stations considered

Station	Period (years)	Average (°C)	Gradient (°C y^{-1})
Perugia	1900–1995	13.2	+0.012
Rome	1880–1995	16.0	+0.0082
Potenza	1926–1995	12.5	no trend
Palermo	1926–1992	18.4	no trend

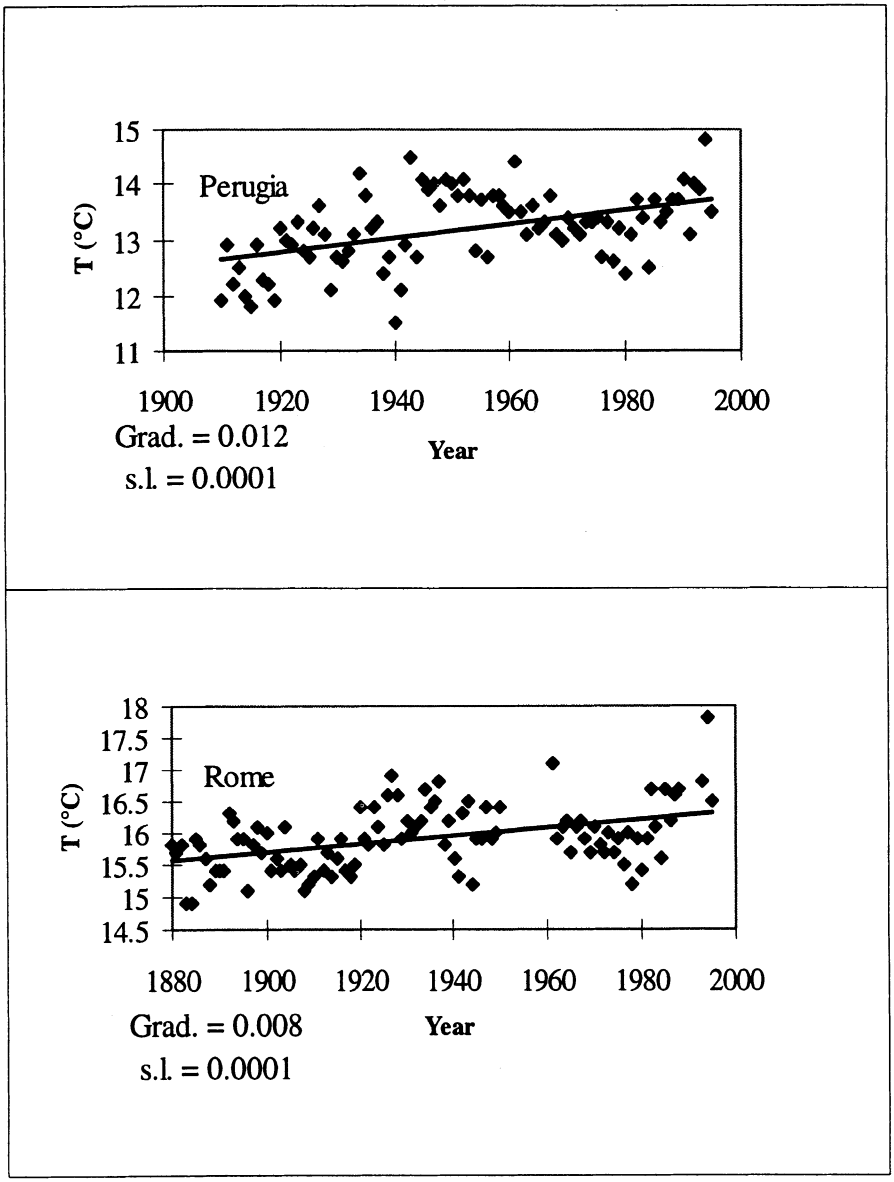


Figure 11.4. Perugia and Rome. Average yearly temperature data and linear fit.

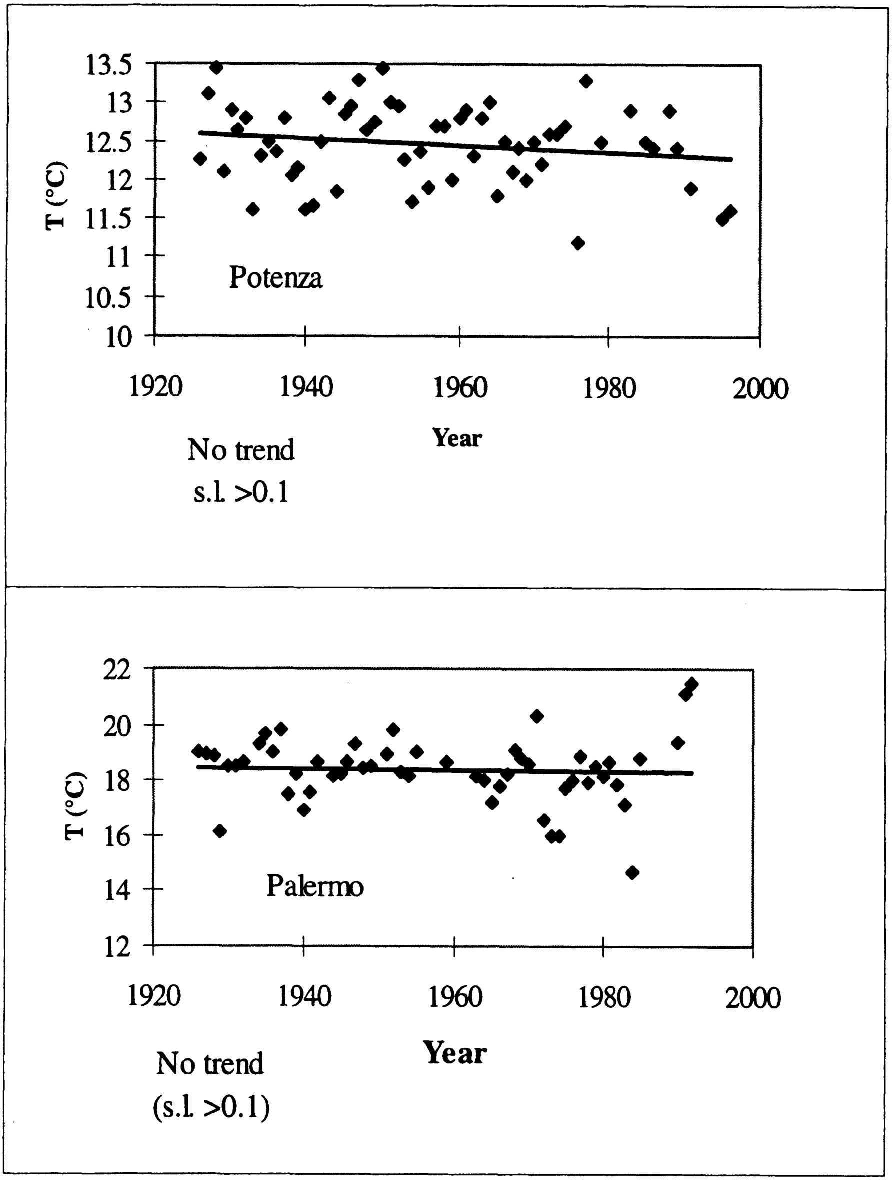
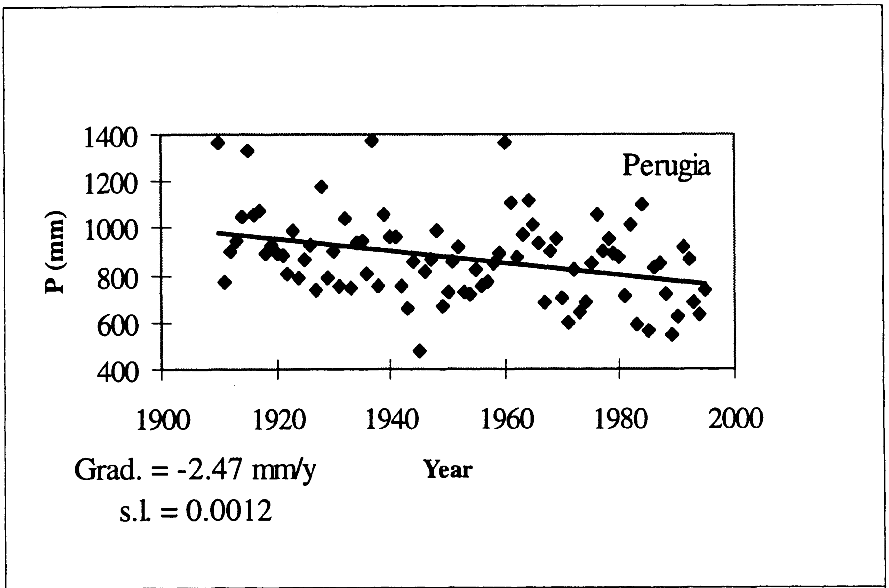


Figure 11.5. Potenza and Palermo. Average yearly temperature data and linear fit.



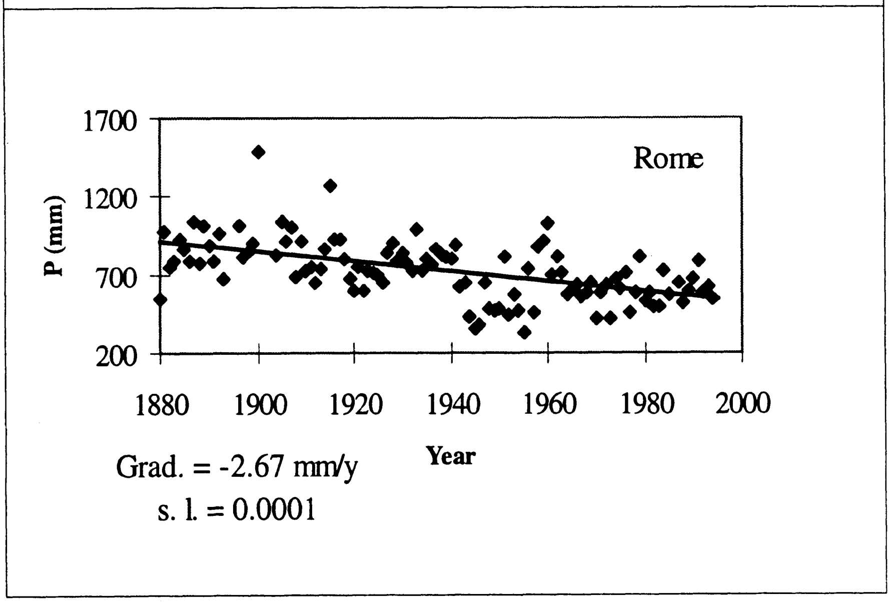


Figure 11.6. Perugia and Rome. Average yearly rainfall data and linear fit.

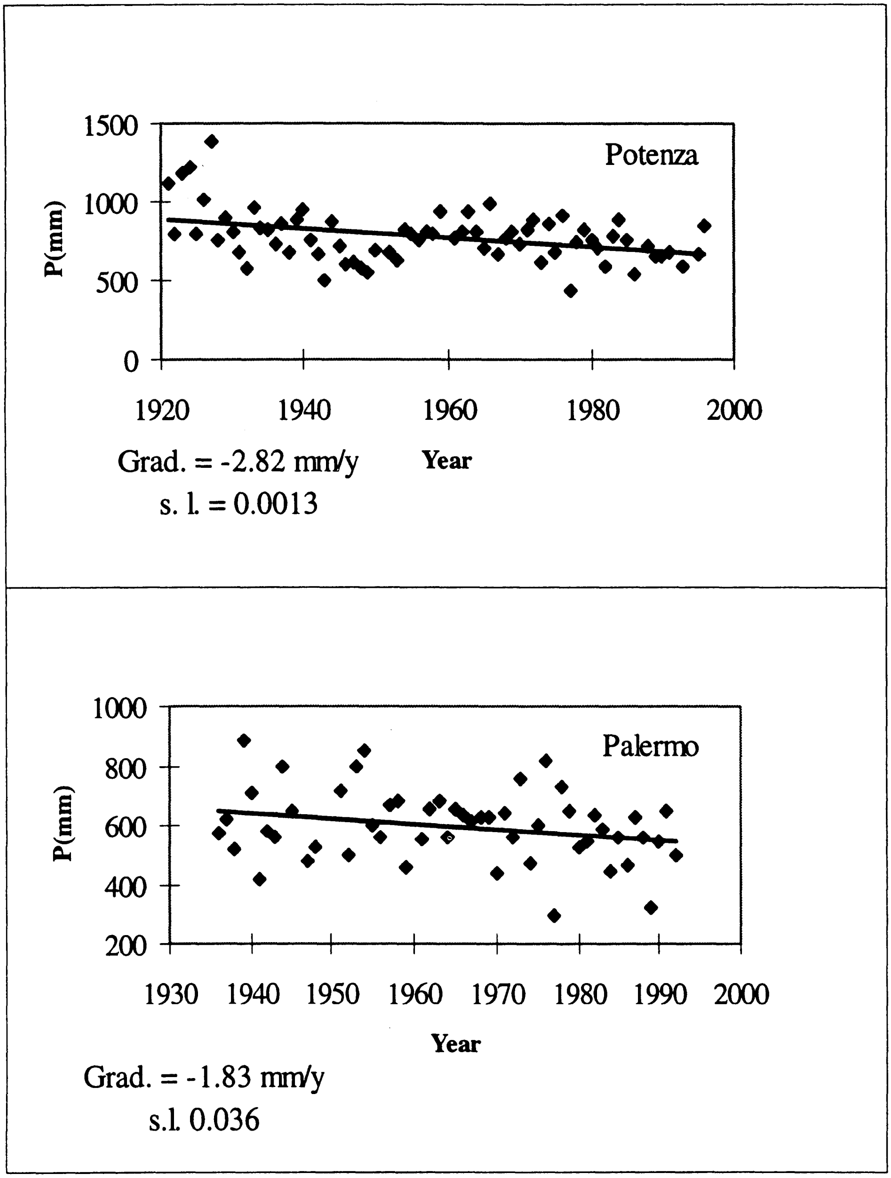


Figure 11.7. Potenza and Palermo. Average yearly rainfall data and linear fit.

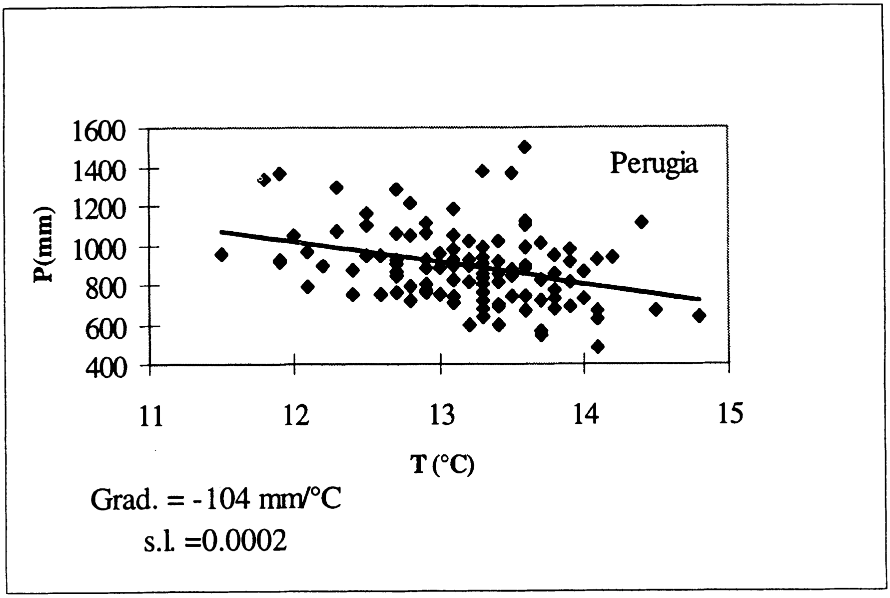


Figure 11.8. Perugia. Average yearly rainfall versus average yearly temperature.

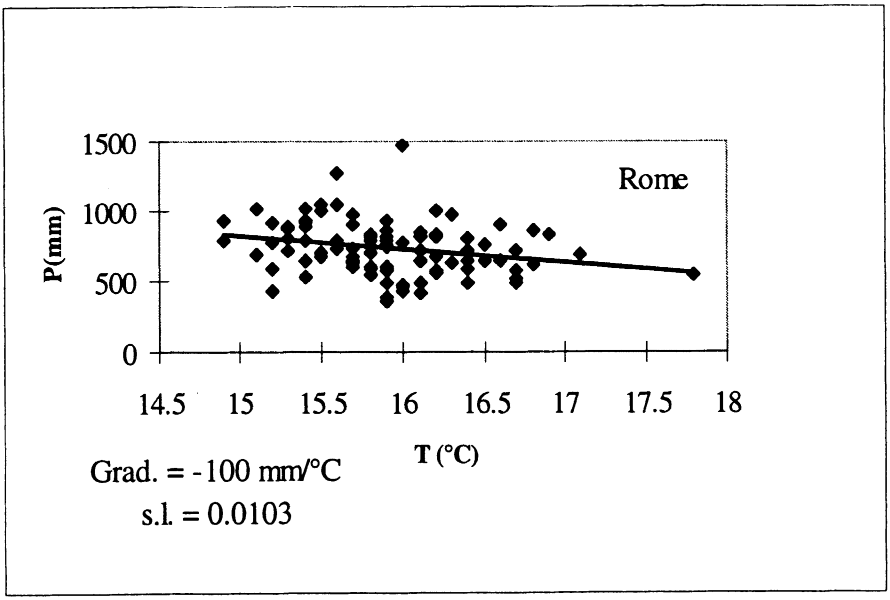


Figure 11.9. Rome. Average yearly rainfall versus average yearly temperature.

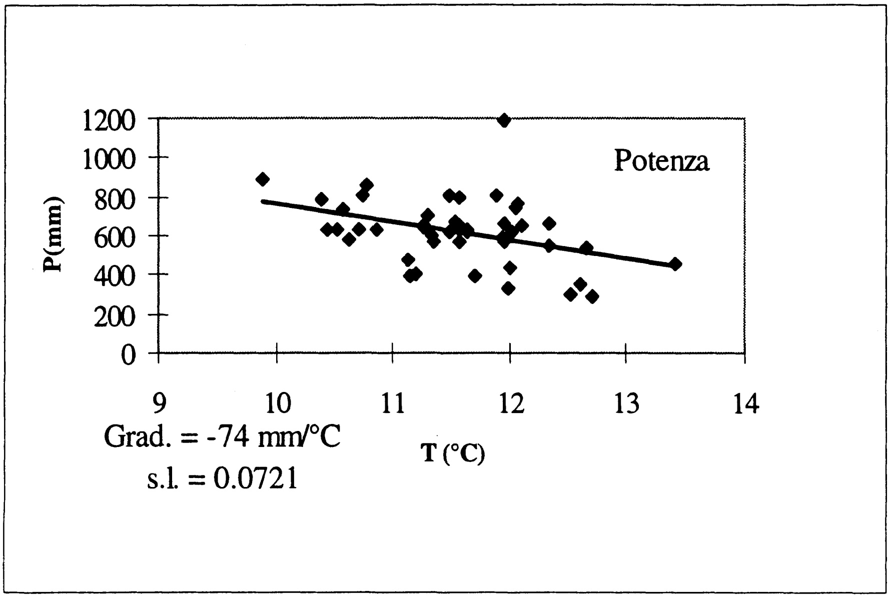


Figure 11.10. Potenza. Average yearly rainfall versus average yearly temperature.

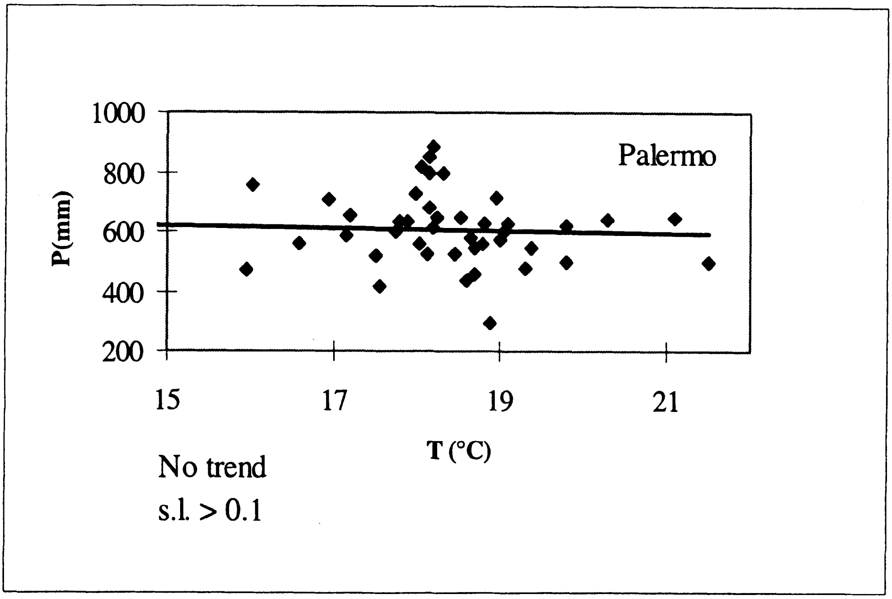


Figure 11.11. Palermo. Average yearly rainfall versus average yearly temperature.

In order to check further whether the climatic behavior is similar within the area considered, and to have an overall view of the recent evolution of the rainfall pattern, the monthly rainfall data from Perugia, Rome and Potenza were smoothed out and plotted together (Fig. 11.12). The smoothing was done in such a way that in Figure 11.12 the average rainfall is given for each consecutive ten-year moving average, with each starting one month later. For instance, the first data plotted for Perugia represent the average yearly rainfall of the decade beginning on 1 January 1901 and ending on 31 December 1910; the second represents the average yearly rainfall of the decade beginning on 1 February 1901 and ending on 31 January 1911, and so on. The results are consistent with the previously mentioned papers on the climatic homogeneity of the Western Mediterranean (Conte and Giuffrida, 1991; Balafoutis and Arseni-Papadimitriou, 1996). The figure indicates also that some cyclicity is present, and suggests that, if the apparent cyclic or pseudo-cyclic sequence continues, inside an overall decreasing trend, higher rainfall than in the recent past is to be expected for the years immediately ahead.

Otherwise the analysis shows that:

a) If there is a longer-term trend, it is toward increasing temperature and decreasing rainfall.

b) In three cases out of four there is a significant inverse relation between temperature and rainfall: on a yearly average, the higher the temperature, the lower the rainfall.

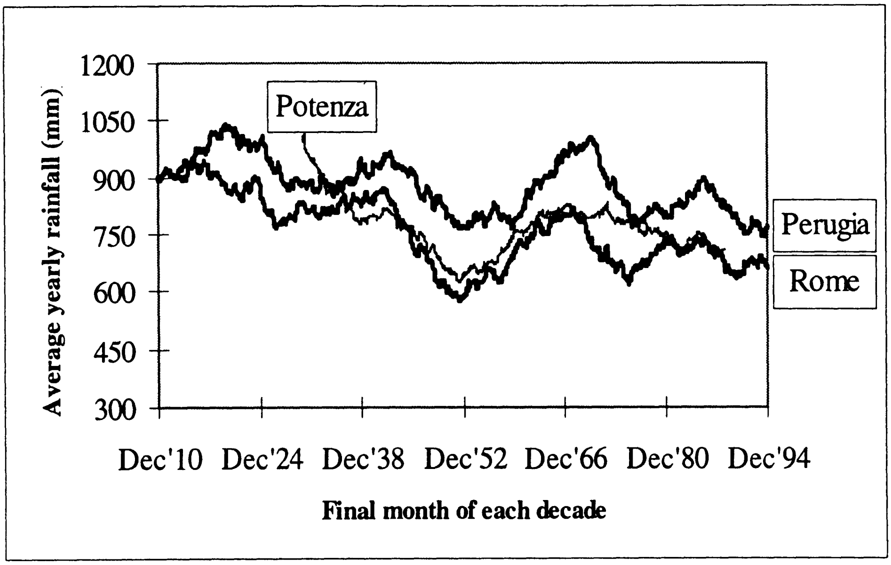


Figure 11.12. Moving average of rainfall computed for three different stations decade by decade (see text for explanations and comments).

c) The comparison between the smoothed series of Perugia, Rome and Potenza (Fig. 11.3.2.7) shows that the behavior is similar, *i.e.* the trends are consistent on a regional scale.

Moreover, the actual water yield of the hydrological systems throughout the area seems to be consistent with a decreasing trend (Dragoni, 1996). This is reflected also in the fact that the reservoirs built in the last decades are rarely filled to full capacity, and now some authors consider them to be generally oversized (Leone, 1996:91–92). The reason for this is that in many cases, one to two decades passed between the time of planning and the time of completion. Thus the calculation of water availability was done from old data sets, showing greater water resources than those actually existing today.

It is important to appreciate that the analyses done of the instrumental data (*i.e.* on the data of the last one hundred years) show that today the climate behaves the same as during the last 3,000 years: low temperature very consistently corresponding to high rainfall, and vice versa. This ought to mean that, for temperature variations not in excess of about ±1°C, the analog approach should be applicable to the prediction of tendencies in the hydrological cycle as a function of temperature variations.

11.3.3. FORESEEABLE SCENARIOS FOR THE NEXT FIFTY YEARS

For the region here being considered, the different GCM scenarios consequent upon a doubling of CO_2 have been exceptionally at variance. The 'forecasts' for rainfall range between ±30% of the present values, while, for temperature, increases of up to 3–4°C are considered possible (*cf.* Pinna, 1996; Wigley and Raper, 1992; Gaudioso *et al.*, 1995:63–64).

The studies carried out on changes in the hydrological cycle have suggested increased aridity and lower water yield (*cf.* Da Cunha, 1989, in Lins *et al.*, 1993:89; Gaudioso *et al.*, 1995:67). Indeed, in spite of the numerical differences, until about 1994–95 virtually all the GCMs agreed that the temperature would increase in this area. Later, at the regional level, eight out of nine AOGCMs suggested a winter rainfall increase (Kattenberg *et al.*, 1996:337). Recently in some more complex models, the cooling effects of the aerosols were also considered. These models point towards a cooling in summer and an increase in precipitation (Kattenberg *et al.*, 1996). How far these preliminary results are more reliable than the old ones is an entirely open question (*q.v.* Charlson and Wigley, 1994:36–38; IPCC, 1996:42–43; Kattenberg *et al.*, 1996:300, 305; Singer, 1996; Schwartz and Andreae, 1996; Hasselmann, 1997:915).

The problem is that no GCMs, no matter how complex or whether or not they consider the aerosols, do not incorporate all the relevant agents which control the climate. This is demonstrated by the actual data sets: considering instrumental data only, there are oscillations which are not simulated by the results of GCMs. On a longer scale, it does not seem that the cool and warm periods of the last 3,000 years can be simply explained by CO_2 variations, aerosols, Croll-Milankovitch cycles and by the physics embodied in the presently available GCMs.

Here one should stick to what the data say. During the last one hundred years there has been a slight increase in temperature and a decrease of rainfall. This behavior is consistent with what already happened during the last 3,000 years, when a typical warm or cool period lasted for up to few hundred years. In a broad sense, this evidence is more similar to the scenarios given by the older models, *i.e.* to the ones that forecast an increase in temperature and a decrease in rainfall. So it is important I present next what could reasonably happen if warming should continue, and if the decrease in rainfall should continue as well.

In projecting the future scenarios, no specific hydrological systems will be taken into consideration. The prognoses will be made solely on the basis of the mean precipitation and temperature values which are representative of the conditions in broad regions, for time periods of fifty years or more. In this manner, it is possible to consider 'water yield' as equal to 'water or moisture surplus', as defined by Thornthwaite and Mather (1957:187, 193), and to consider both terms to be synonyms of 'renewable water resources'.

In other words, from this point on the following is considered to be valid:

$$S = P - E_a \tag{11.1}$$

where the symbols stand for:

S = average yearly water surplus = average yearly water yield (mm)

P = average yearly rainfall (mm)

E_a = average yearly actual evapotranspiration (mm).

The estimation of Ea was done simply using the Turc formula (1954):

$$E_a = P/(0.9 - (P/(300 + 25T + 0.05T^3))^2)^{0.5} \qquad (11.2–3)$$

where:

T = average yearly temperature (°C).

Of course, from a theoretical point of view, and with more data available than hypothetical yearly rainfall and temperature, there are many better ways to compute E_a. However, let us remember that what is happening to the climate and to the environment (not to mention what is going to happen) is quite far from being definitely understood. This means that up until now, even the most complex models used to construct future scenarios will not provide a sound quantitative forecast, only a reasonable guess. In view of this, the Turc formula is routinely used, without too many complaints, to estimate the average yearly evapotranspiration when only rainfall and temperature are known (*cf.* Castany, 1982:14; Custodio, 1983:341; Singh, 1989:28; Shaw, 1993:241). On this basis, the method can also give reasonable scenarios for the average yearly water yield of a given area under different temperatures and rainfall amounts.

The graphs in Figs. 11.13, 11.14, 11.15, and 11.16 describe the water surplus for the stations selected as a function of the average yearly rainfall and temperature, according to equations (11.1) and (11.2–3). In every graph the water surplus is graphically represented as a function of the mean annual rainfall, plotted on the abscissa. Three lines are plotted, each one representing the behavior of S as function of P, given different values of T. The upper line in each graph represents the present mean annual temperature for the analyzed station, whereas the two lower lines are referred to an increase in T, with respect to the present, of 0.25°C and 0.5°C. Present average yearly rainfall and temperature are understood as the averages computed using the last 50 years of data available (see Tables 11.3, 11.4). Solid dots represent the present situation; empty dots represent the average S for the next 50/60 years if the detected trends should continue.

It is interesting to note that, when the yearly average water yield of a certain system computed with complex hydrological models, is plotted as a function of average yearly rainfall at constant temperature, the resulting graphs are quite similar to those here obtained (*cf.* Mimikou and Kouvopoulos, 1991:253–255; Nash and Gleik, 1993; in Loaiciga, 1996:17). This gives further support to the validity of the procedure here adopted.

Figures 11.13, 11.14, 11.15, and 11.16 show that:

a) changing the temperature alone will have a small effect on the water yield;

b) the combination of modest temperature increase and modest rainfall decrease, would have relevant effects on the water resources of the more arid areas (as in the cases of Palermo and Rome);

c) typically produce by 2050 an average decrease of between 20 and 30% of the present water yield.

11.4. Discussion

In the area here considered (the western Italian peninsula, south of 43°N, and Sicily, Fig. 11.1) the work done seems to indicate that:

1) During the last 3,000 years, the climate has alternated between warm and cool periods lasting a few hundred years each.

2) The warmer periods have corresponded to less rain; cooler periods to higher.

3) During the last 100–120 years there has been a shift towards warmer and, as has happened during the late Holocene, less rainy conditions.

4) Extrapolation to the next 55 years of the detected trends (trends compatible with the scenarios forecast by the majority of GCMs) shows a decrease in wa-

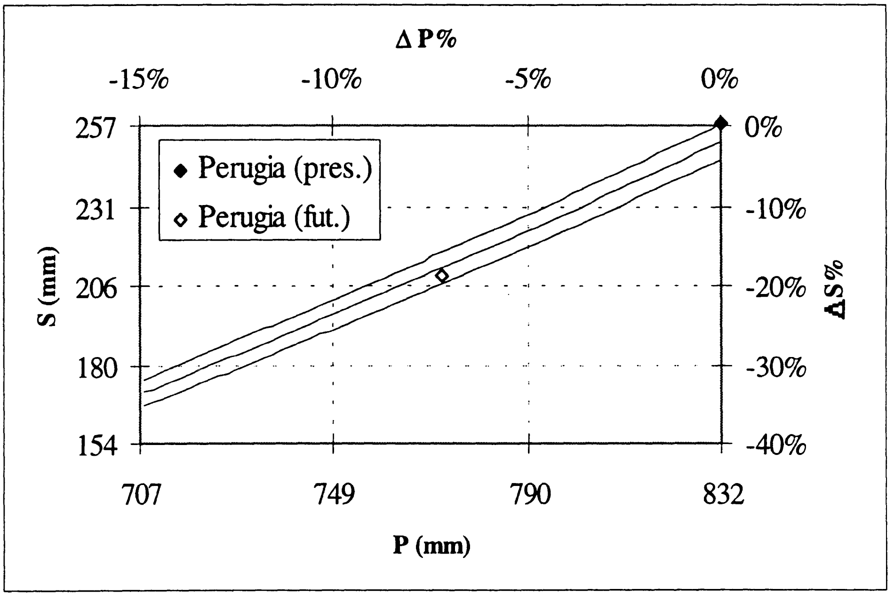


Figure 11.13. Foreseeable scenarios for Perugia, according to different rainfall decreases and temperature increases, for the period 1996–2050; average present yearly temperature is 13.5°C. See text for explanation.

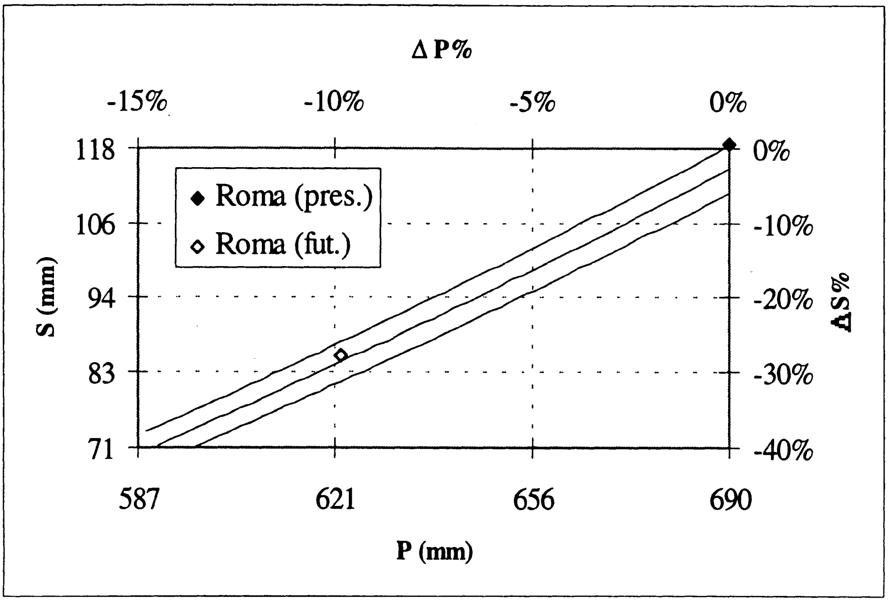


Figure 11.14. Foreseeable scenarios for Rome, according to different rainfall decreases and temperature increases, for the period 1996–2050; average present yearly temperature is 16.3°C. See text for explanation.

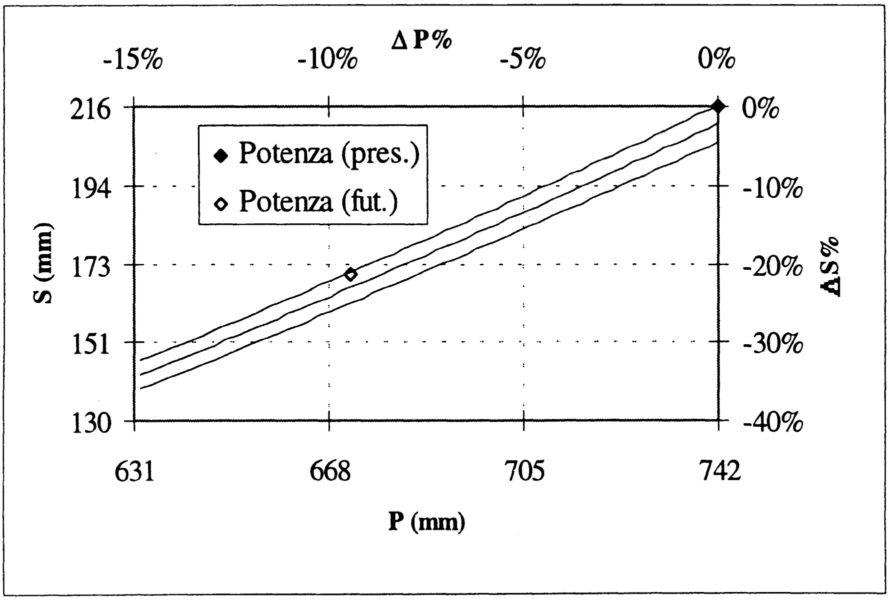


Figure 11.15. Foreseeable scenarios for Potenza, according to different rainfall decreases and temperature increases, for the period 1996–2050; average present yearly temperature is 12.5°C. See text for explanation.

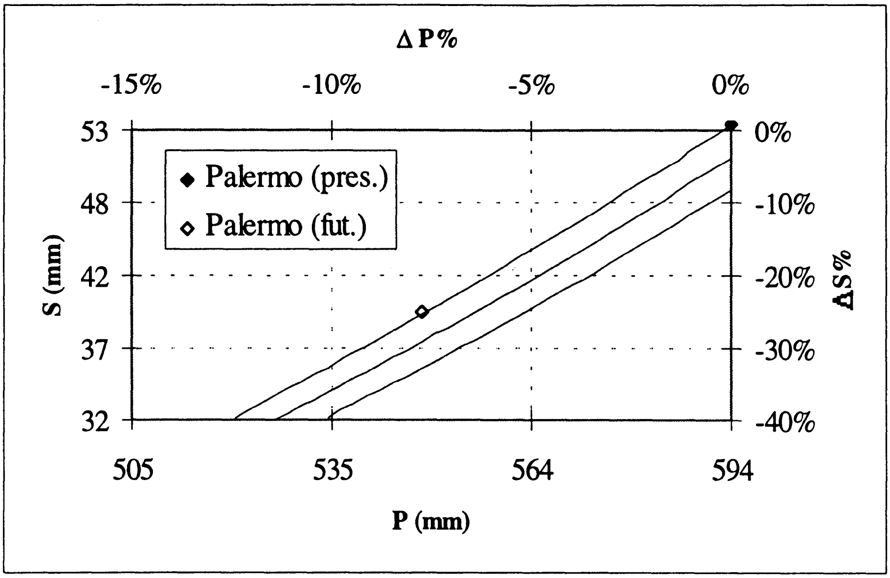


Figure 11.16. Foreseeable scenarios for Palermo, according to different rainfall decreases and temperature increases, for the period 1993–2047; average present yearly temperature is 18.4°C. See text for explanation.

ter yield of 20 to 30% in the time interval 1996–2050, compared with the present. Greater decreases should be expected if the extreme forecast values of some GCMs actually occur.

However, each of the points made above gives rise to certain rejoinders:

Regarding point 1), not everybody accepts that during the late Holocene the climate has fluctuated notably on a global level: for instance, serious doubts have been raised about the actual existence of the WMP (Serre-Bachet, 1994:222; Hughes and Diaz, 1994:124–125, 136; Briffa *et al.*, 1990, in Goudie, 1994:167). Of course, if the WMP did not exist, it is difficult to accept also the LIA, as this is defined as a deterioration of the climate after the 14th–15th centuries. To discuss such an issue in depth (as well as the problem of the effect of climatic change on human history) is beyond the scope of this paper. However, the interested reader can easily find stimulating discussion in the literature, notably in works by Lamb (1972; 1977; 1985) and in some more recent books (Bell and Walker, 1992; Goudie, 1994; Brown, 1995; Whyte, 1995; Pinna, 1996). Sometimes the debate is rough. It has to be admitted that, in some cases, climatic change is presented as a sort of 'politically correct ecological belief'. Even so, it is not possible to deny all the data which support the existence of late Holocene climatic variations, no matter how many ambiguous and ill understood facts accompany such data. Regarding this, I wish to mention a little-known work by Monterin (1937), then the Director of the Royal Meteorological and Geophysical Observatory of Mt. Rosa. Well before climatic change became so fashionable and pressing an issue, this author brought to light a great amount of data and facts showing, at least for the Italian Alps, the existence of a warm period before the 16th century, as well as of some warming after the mid-19th century (in the 1930s the terms 'WMP' and 'LIA' did not exist).

Regarding point 2), a large number of studies demonstrate that, during the Pleistocene glaciations, the Italian peninsula and surrounding European regions had a dry climate compared to today (*cf.* Bonatti, 1966:984–985; CLIMAP, 1976:1132–1133; Folieri *et al.*, 1988; Zonneveld, 1996:104). This seems to contrast with the evidence about Holocene lake levels in central Italy as shown in Fig. 11.2. However, these contrasting relationships (cooler = dryer climate in the Pleistocene; cooler = wetter in the late Holocene), should not be seen as mutually exclusive. After all, the global circulation during a true glacial period and that during the last 3,000–4,000 years are quite different. The main differences are:

a) *Temperature:* during a Pleistocene glacial period, the average global temperature was at least 4–7°C lower than today (*q.v.* Budyko et al., 1988:2; Mélières et al., 1991:333; Goudie, 1994:116). In some regions in Europe, the temperatures of the coldest months could be as much as 10–15°C lower than at present (Pons, 1993:463, 465). For the late Holocene fluctuations there is general agreement that temperatures differed at most by ±2°C from present values (Dansgaard *et* *al.*, 1969:380; Bell and Walker, 1992:70–73). For the last millennium, two very recent estimates give a temperature fluctuation of less than ±0.5°C (Crowley and Kim, 1995, in Nicholls *et al.*, 1996:176–177). In other words, during the Pleistocene glacial phases the temperature was so low as to decrease considerably global evapotranspiration compared with the Holocene.

b) *Ice caps extension:* during the maximum glacial phases of the Pleistocene the total extension of the ice caps was some three times larger than at present (Embleton and King, 1967, in Goudie, 1994:56). Never during the coolest periods of the last 3,000 years were ice conditions reached at all comparable with those prevailing during a glacial maximum. The mean albedo and oceanic circulation during the glacial phases were entirely different than those of the last 3,000 years.

c) *Sea level:* during the Pleistocene glacial phases, the sea level was notably lower than at any time in the last 3,000 years. In particular, during the last glacial phase, the sea level was around 130 meters lower than today, and the shape of Italy was as shown in Fig. 11.1. For that reason alone, every locality considered here had a more continental climate. All in all, it is futile to expect parallelism between the climate during the glacial phases and that of the last 3,000 years.

Regarding point 3), not everybody agrees that at present a sustained warming is going on globally. Not long ago it was argued that, if there were global warming, the polar ice caps should melt: which seemed not to be happening (Fong, 1993). Other researchers found that some instrumental data sets did not show the warming forecast, in accordance with the greenhouse effect, for the Arctic regions (Kahl *et al.*, 1993:335–337). Some reasonable, if not definitive, answers to these objections are possible, as those given, for instance, by Walsh (1993:300–301). As to the lack of evidence of melting ice caps, the latest evidence is that Antarctic sea-ice decreased notably (de la Mare, 1997). May one point out that for central-southern Europe, aside from the previously mentioned statistical analysis on the instrumental data, many different events are actually taking place which, taken together, are quite difficult to explain in any way other than that of an average general warming, no matter how small, discontinuous and differentiated from one area to the next. For instance, according to a recent review paper, in the Italian Alps, since the end of the LIA, the glaciers have had a total surface reduction of more of 30%; in the Lombard Alps, nearly 30% of the 271 glaciers have become extinct since 1961. In particular, in the zone considered in the present paper, the Calderone glacier has lost nearly 70% of its surface area and 90% of its volume during the 20th century: the Calderone is the only glacier in the Apennines. It is located at 42°N, less than one hundred kilometers from the drained Lake Fucino. Besides this, still on the Italian side of Alps, the snow line has risen by up to few hundred meters (Laureti *et al.*, 1996:30). Another recent paper reports the spreading in Germany of many species of dragonfly which in the recent past were more typical of warmer regions. This is also coupled in Germany with the earlier flowering and ripening of plants (Ott, 1996). It is obvious that all these indirect signs of warming cannot be explained by urban heating.

Moreover, after admitting the warming, doubts remain as to whether the greenhouse effect is really responsible for the 0.3–0.5°C rise in the yearly average during the last 150 years. From a methodological point of view, it seems difficult to attribute for certain the present climatic variation to the anthropogenic greenhouse effect, as climatic changes have happened in the recent past without man's intervention. This difficulty of having positive and certain proof could remain even if, in the near future, statistical tools indicate that a variation larger than 'natural variability' of the instrumental data is occurring. The proxy data indicates that, during the last 3,000 years, non-anthropogenically caused climatic variations larger than those embodied in our instrumental data sets.

Regarding point 4), concern only about a decrease in water yield can be questioned—after all, we do not really know what is going on. Witness the recent GCMs, which incorporate the effect of aerosols, raising the possibility of a decrease in temperature coupled with some increase in rainfall for southern Europe (Kattenberg *et al.*, 1996). This scenario should not be allowed for too much in the planning of water management for the near future. This view is based on the following:

a) The actual data point towards a decreasing trend for rainfall and an increasing trend, or no trend, for temperature, *i.e.* towards decreasing water yields.

b) Considering the GCMs forecast in the area here studied, it could seem that the ones with aerosol forecast definitively wetter scenarios. However, if one analyses the indications of the GCMs with aerosols (Nicholls *et al.*, 1996:306–307), it is discovered that the changes to be expected on that basis by the years 2030–2050, compared to the present situation, are limited and hard to interpret, given the low spatial resolution. Thus the models give the same variations of rainfall and temperature between Perugia and Palermo despite their different climates (*cf.* Tables 11.3, 11.4). And regarding water yield, it would be easy to show that, according to the Thornthwaite and Mather method, in the interval 1995–2050 the conditions forecasted by GCMs with aerosol would produce on balance a negligible increase in the average yearly water yield. In any case the actual data affords no evidence of any long term increase of the average yearly rainfall or of the water yield.

c) Over the past 150 years, we have emerged from a cool and rainy period (the LIA). Historically, warm and dry periods have lasted up to 2–3 centuries (Fig. 11.3).

d) Yet in spite of all the objections, no matter how sophisticated and reasonable, it is hard to imagine sound reasons why, at least in the long run, the greenhouse effect should not take place. This means some increase in temperature, which in the past has corresponded to drier conditions.

e) Today there is already a shortage of water. Reservoirs built according to climatic data more than 50 years old appear oversized (Leone, 1996). Furthermore, the annual water needs of Italy in the early 1980s were estimated to be 40.8×10^9 m^3, while for the period 2000–2015 they are estimated to be 53.5×10^9 m^3 (Rusconi, 1994:115). Therefore, it would definitely be too risky to give credence to an optimistic interpretation.

In conclusion, it must be said the task which remains to be done in order to understand better what has happened and is now happening is big and important. The following points are particularly relevant, and could complement the results obtained by present and future GCMs.

- Research aimed at detecting more historical information (hidden within lakes all over Italy, in the archives and such) must continue. It is particularly necessary to increase the temporal accuracy of the information.
- Update and carefully check all the data sets presently available: if one verifies the original data recorded on paper, or compares the published series, one finds many discrepancies and uncertainties.
- Up until now, not all the Italian instrumental data sets have been examined. In particular, it is necessary to consider sets regarding mountain areas, which are still little covered. Together with other long data sets, these could change the picture presented here.
- Choose a large number of hydrological systems to be monitored with the specific objective of tracking any significant changes in the hydrological cycle, as regards both surface and underground water resources. Both total and seasonal water yields should be considered, as well any changes in the regimen.
- It is imperative to revise the plans for future water supply management, taking into consideration measures which up until now have been neglected in Italy, such as aquifer recharge and the development of underwater marine springs.

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References

Arnell, N.W. (1995) Scenarios for hydrological climate change impact studies. *In* H.R. Oliver and S. Oliver (eds.), *The Role of Water and the Hydrological Cycle in Global Change*, Springer-Verlag, Berlin Heidelberg, Nato ASI Series, **1**(**31**), pp. 389–407.

Balafoutis, C. and Arseni-Papadimitriou, A. (1996) Analysis of the temperature data over the Mediterranean basin. *In* B. Obresbska-Starkel and T. Niedzwiedz (eds.), *Proc. Int. Conf. on Climate Dynamics and the Global Change Perspective*, Cracow, October 17–20, 1995, Universitas Iagellonica, Acta Scientiarum Litterarumque, Schedae Geographicae-Fasciculus CII; Schedae Instituti Geographici Universitatis Iagellonicae-Fasciculus, **CXXIV**, pp. 321–327.

Bell, M. and Walker, M.J.C. (1992) *Late Quaternary Environmental Change-Physical and Human Perspectives.* Longman Scientific and Technical.

Belluomini, G., Manfra, L., and Proposito, A. (1993) Lago di Bolsena: Datazione con il Carbonio-14 di reperti sommersi e fluttuazioni oloceniche del livello delle acque. *Il. Quaternario* **6**, 265–270.

Bonatti, E. (1966) North Mediterranean climate during last Wurm glaciation. *Nature* **209**, 984–985.

Brisse, A and De Rotroù, L. (1876) *Dessechément du Lac Fucino.* Reprint 1985 in Italian, U.M. Pallanza (ed.), *Avezzano, Tipolitografia Nobile Paolini.*

Brown, N. (1995) The impact of climate change: Some indications from history, AD250–1250, OCEES Research Papers No 3.

Budyko, M.I., Golitsyn, G.S., and Izrael, Y.A. (1988) *Global Climatic Catastrophes.* Springer-Verlag.

Camuffo, D. and Enzi, S. (1995) The analysis of two bimillenary series: Tiber & Po river floods. *In Climatic Variations and Forcing Mechanism of the Last 2000 Years*, Springer-Verlag, NATO–ANSI Series, **41**, pp.433–452.

Castany, G. (1982) *Principes et Méthodes de l'Hydrogéologie.* Dunod Université, Paris.

Castellani, V. and Dragoni, W. (1991) Opere Arcaiche per il controllo del territorio: Gli emissari sotterranei artificiali dei laghi albani. *In* Bergamini (ed.), *Atti del Convegno Gli Etruschi Maestri di Idraulica*, Perugia, febbraio 1991, ELECTA, **XX**, pp. 43–60.

Charlson, R.J. and Wigley, T.M.L. (1994) Aerosol di acido solforico e cambiamenti climatici. *Le Scienze* **308**, 30–38.

Cialini, G. (1991) *Memoria storico-artistica della Chiesa di Santa Maria di Ancaèlle.* Università degli Studi di Perugia–Collegio Pio della Sapienza, Perugia.

CLIMAP Members (1976) The surface of the Ice-age Earth. *Science* **191**, 4232, 1131–1137.

Conte, M. (1994) Situazioni meteorologiche del Mediterraneo ed andamento delle precipitazioni sul territorio italiano. *In* G. Rossi and G. Margaritora (eds.), *La Siccità in Italia 1988–1990*, Presidenza del Consiglio dei Ministri, Dipartimento della Protezione Civile, Roma, pp. 13–40.

Conte, M. and Giuffrida, A. (1991) L'oscillazione climatica mediterranea: Possibili effetti sulle precipitazioni in Italia. *In Atti della LXI Riunione SIPS, 23–26 ottobre 1991.*

Conte, M., Giuffrida, A. and Tedesco, S. (1991) L'oscillazione mediterranea, in Le variazioni recenti del clima (1800–1990) e le prospettive per il XXI secolo. *Proceedings of the Italian Geographical Society Conference,* Rome, 5–6 April 1990, *Memorie Società Geografica Italiana* XLVI, 115–124.

Custodio, E. and Llamas, M.R. (1983) *Hidrologia subterranea*, **1**. Ediciones Omega.

Da Cunha, L. (1989) Water resources situation and management in the EEC. *Hydrogeology* **2**, 57–69.

D'Amato, S. (1980) *Il primo prosciugamento del Fucino.* Avezzano.

Dansgaard, W., Johnsen, S.J., Moller, J., and Langway, C.C. (1969) One thousand centuries of climatic record from Camp century on the Greenland Ice Sheet. *Science* **166**, 377–381.

Dansgaard, W. and Langway, C.C. (1971) Climatic record revealed by the Camp Century Ice Core. *In* K.K. Turekian (ed.), *The Late Cenozoic Glacial Ages*, New Haven, Conn., pp. 37–56.

de la Mare, W.K. (1997) Abrupt mid-twentieth-century decline in Antartctic sea-ice extent from whaling records. *Nature* **389**, 57–60.

Del Prete, M., Giaccari, E., and Trisorio-Liuzzi, G. (1993) Rischio da frane intermittenti a cinematica lenta nelle aree montuose e collinari urbanizzate della Basilicata, Università della Basilicata, CNR–GNDCI, U. O. 2.37, rapporto 1992.

Dragoni, W. (1982) Idrogeologia del lago Trasimeno: sintesi, problemi, aggiornamenti. *Geografia Fisica Dinamica Quaternario* **5**, 192–206.

Dragoni, W. (1996) Response of some hydrogeological systems in central Italy to climatic variations. *In* A. Angelakis and A. Issar (eds.), *Diachronic Climatic Impacts on Water Resources with Emphasis on Mediterranean Region*, Springer-Verlag, NATO–ASI Series, **1**(36), pp. 193–229.

Dragoni W. and Valigi, D. (1998) Risorse idriche e variazioni climatiche nel Mediterraneo centro-settentrionale: stato delle conoscenze. *Proc. of Sympsium 'La georisorsa acqua nel bacino del Mediterraneo'*, Erice, 2–4 ottobre 1997 (in press).

Folieri, M., Magri, D., and Sadori, L. (1988) 250,000-year pollen record from Valle di Castiglione (Roma). *Pollen and Spores* **30**, 329–356.

Fong, P. (1993) Conclusive evidence of no global warming. *Abstracts* of the 1993 Spring Meeting, paper A12C-7, *EOS supplement*, April 20, 69.

Fugazzola-Delpino, M.A. and Lombardi, A.L. (1982) Scheda Lago di Bracciano (Roma), Catalogo della Mostra. *In* A. Aspes (ed.), *Palafitte: mito e realtà*, 8 luglio–31 ottobre 1982, Verona, Museo Civico di Storia Naturale, Verona, Italy.

Gambini, E. (1995) Le oscillazioni di livello del lago Trasimeno, *Quaderni del Museo della Pesca del Lago Trasimeno*. Grafiche Piemme, Perugia.

Gambini, E. and Pasquali, E. (1996) *I tori—la gran pesca del Medioevo al Lago Trasimeno*. Edizioni Guerra, Perugia.

Gaudioso, D., Contaldi, M., Sciortino, M., Dalla Costa, M., Ferrara, V., and Borrelli, G. (1995) First Italian national communication to the framework convention on climate change, Ministry of Environment, ENEA.

Giano, S. I. and Guarino, P.M. (1996) Trasformazioni ambientali in età storica del territorio di Oria (Brindisi). *Il Quaternario* **9**(1), 331–336.

Giraudi, C. (1994) Origine ed evoluzione geologica recente del bacino del Fucino. *In* Burri E. (ed.), *Sulle Rive della Memoria, Il Lago Fucino e il Suo Emissario*, CARSA.

Giuffrida, A. and Conte, M. (1991) L'evoluzione a lungo termine del clima italiano. In Le variazioni recenti del clima (1800–1990) e le prospettive per il XXI secolo, *Proc. of the Italian Geographical Society Conf.*, Rome, 5–6 April 1990. Memorie Società Geografica Italiana XLVI, 329–342.

Goudie, A. (1994) *Environmental Change*. Clarendon Press, Oxford.

Grove, J. M. (1972) The incidence of landslide, avalanches and floods in western Norway during the Little Ice Age. *Arctic and Alpine Research* **4**, 131–138.

Hasselman, K. (1997) Are we seeing global warming? *Science* **276**, 914–915.

Hughes, M.K. and Diaz H.F. (1994) Was there a "Medieval Warm Period", and if so, where and when? *In* M.K Hughes and H.F. Diaz (eds.), *Climatic Change*. Special Issue titled "*The Medieval Warm Period*" **26**(2–3).

IPCC (1996) Technical summary. *In* J.T. Houghton, L.G. Meira Filho, B.A. Callander, N. Harris, A. Kattenberg, and K. Maskell (eds.), *Climate Change 1995. Contribution of WG1 to the Second Assessment Report of the IPCC*, Cambridge University Press, pp. 9–49.

Irti, U. (1991) Gli insediamenti all'aperto del neolitico e dell'età dei metalli. *In* U. Irti, G. Grossi and V. Pagani (eds.), *Atti del Convegno di Archeologia Il Fucino e le aree limitrofe nell'Antichità*, 10–11 novembre 1989, Avezzano, Italy, Archeoclub d'Italia, Sezione della Marsica, pp.76–109.

Kahl, J., Charlevoix, D.J., Zaitsevat, N.A., Schnelt, R.C., and Serreze, M.C. (1993) Absence of evidence for greenhouse warming over the Arctic Ocean in the past 40 years. *Nature* **361**, 335–337.

Kattenberg, A., Giorgi, F., Grassl, H., Meehl, G.A., Mitchell, J.F.B., Stouffert, R.J., Tokioka, T., Weaver, A.J., and Wigley, T.M.L. (1996) Climate models—Projection of Future Climate. *In* J.T. Houghton, L.G. Meira Filho, B.A. Callander, N. Harris, A. Kattenberg, and K. Maskell (eds.), *Climate Change 1995*. Contribution of WG1 to the Second Assessment Report of the IPCC, Cambridge University Press, 285–357.

Jones, P.D. (1995) Maximum and minimum temperature trends in Ireland, Italy, Thailand, Turkey and Bangladesh. *Atmospheric Research* **37**, 67–78.

Lamb, H.H. (1972) *Climate: Present, Past & Future*. Methuen & Co Ltd, London, I.

——. (1977) *Climate: Present, Past & Future*. Methuen & Co Ltd, London, II.

—— (1982) *Climate History and the Modern World*. Methuen, reprint 1985.

Laureti, L., Sauro, U., Terranova, R., Brandolini, P., Firpo, M., Pirola, A., Piccazzo, M., Rossetti, R., and Smiraglia, C. (1996) Physical changes in the Mediterranean basin. *Bollettino Società Geografica Italiana* XII, 7–44.

Leone, G. (1996) Prospettive dell'approvigionamento idrico nell'Italia meridionale e nelle isole. *In Atti Convegno Nazionale "Verso la razionale utilizzazione delle risorse idriche: dalla legge 183/89 alla legge 36/94"*, Taormina, 2–3 dicembre 1994, CNR-GNDI pub. n. **1449**, 67–96.

Lins, H.F., Shiklomanov, I.A., and Stakhiv, E.Z. (1993) Impacts on hydrology and water resources. *In* J. Jager and H.L. Ferguson (eds.), *Climate Change: Science, Impacts and Policy, Proc. Second World Climate Conference*, Geneva, 29 October, 1990, pp. 87–97.

Loaiciga, H.A., Valdes, J.B., Vogel, R., Garvey, J., and Schwarz, H. (1996) Global warming and the hydrological cycle. *Journal of Hydrology* **174**, 83–127.

Mangianti, F. and Beltrano, M.C. (1995) *Le precipitazioni a Roma*. Ministero delle Risorse Agricole, Alimentari e Forestali, UCEA, Roma.

Melicchia, A. (1942) *Variazioni climatiche nell'Italia centrale e loro rapporto col regime del Tevere. Ricerche sulle variazioni storiche del clima italiano*, **II**, **9**, CNR, Comitato Nazionale per la Geografia, Bologna.

Mélières, M., Martinerie, P., Raynaud, D., and Lliboutry, L. (1991) Glacial-interglacial mean sea level pressure change due to sea level, ice sheet and atmospheric mass changes. *Global and Planetary Change Section* **89**, 333–340.

Mimikou, M.A. and Kouvopoulos, Y.S. (1991) Regional climate change impacts: I, Impacts on water resources. *Hydrological Sciences Journal* **36**, 3, 6, 247–257.

Mintzer, I.M. (1992) *Confronting Climate Change*. Cambridge University Press.

Montanari, A., Rosso, R., and Taqqu, M.S. (1996) Long rainfall records in Italy: Trend or long range dependance? *In Book of Abstracts of the International Congress on Environment and Climate*, March 4–8, 1996, Rome, Italy, p. 6.

Monterin, U. (1937) Il clima sulle Alpi ha mutato in epoca storica? *In Ricerche sulle variazioni storiche del clima italiano*, **II**, **2**, CNR, Comitato Nazionale per la Geografia, Bologna.

Nash, L. and Gleick, P. (1993) The Colorado River basin and Climatic Change. Rep. EPA 230-R-93-009, United States Environmental Protection Agency, Washington, D.C.

Natoni, E. (1944) *Le piene dell'Arno e i provvedimenti di difesa*. Accademia D'Italia, Commissione italiana di studio per i problemi di soccorso delle popolazioni, XII, Le Monnier F., Firenze.

Nicholls, N., Grza, G.V., Jouzel, J., Karl, T.R., Ogallo, L.A., and Parker, D.E. (1996) Observed climatic variability and change. *In* J.T. Houghton, L.G. Meira Filho, B.A. Callander, N. Harris, A. Kattenberg, and K. Maskell (eds.), *Climate Change 1995. Contribution of WG1 to the Second Assessment Report of the IPCC*, Cambridge University Press, pp. 133–192.

Ortolani, F. and Pagliuca, S. (1994) Variazioni climatiche e crisi dell'ambiente antropizzato. *Il Quaternario* **7**(**1**), 351–356.

Ott, J. (1996) Zeigt die Ausbreitung der Feuerlibelle in Deustschand eine Klimaveranderung an? Mediterrane Libellen als Indikatoren fur Anderungen in biozonosen. *Naturschutz und Landschaftsplanung* **28**(2), 53–63.

Pavese, M.P., Banzon, V., Colacino, M., Gregori, G.P., and Pasqua, M. (1992) Three historical data series on floods and anomalous climate events in Italy. *In* S. Bradley and P.D. Jones, (eds.), *Climate Since A.D. 1500*, Routledge, pp. 155–170.

Peroni, R. (1994) *Introduzione alla Protostoria Italiana.* Laterza, Bari.

Piervitali, E., Colacino, M., Conte, M. (1997a) *Le preciptazioni sul Mediterraneo centro-occidentale nel periodo 1951–1995. IFA R.I.*, 97–2, Roma.

Piervitali, E., Colacino, M., Conte, M. (1997b) Signals of climatic change in the Central-Western Mediterranean basin. *Theoretical and Applied Climatology* (in press).

Pinna, M. (1996) *Le Variazioni del Clima: dall' Ultima Grande Glaciazione alle prospettive per il XXI secolo.* Franco Angeli, Milano.

Pons, A. (1993) Climate changes in Europe during the last two climatic cycles with emphasis on temperate post-optimum evolution and entering into glaciation. *In* I. Troen, (ed.), *Global Change: Climate Change and Climate Change Impacts*. Proc. of the symposium held in Copenhagen, Denmark 6–10 September 1993, pp. 455–467.

Rapetti, F. and Vittorini, S. (1991) Studio di serie pluviometriche secolari in Toscana, in Le variazioni recenti del clima (1800–1990) e le prospettive per il XXI secolo. *Memorie Società Geografica Italiana* **XLVI**, 457–478.

Rein, B. and Negendank, J.F.W. (1993) Organic carbon contents of sediments from Lake Schalkenmeherener Maar: A palaeoclimate indicator. *In* J.F.W. Negendank and B. Zolitschka (eds.), *Paleolomnology of European Maar Lakes*, Springer-Verlag, Heidelberg, pp. 163–171.

Rufini, E. (1989) Contributo alle conoscenze idrogeologiche del lago di Bolsena. Thesis, Perugia Univ.

Rusconi, A. (1994) *Acqua: Conoscenze ed utilizzo.* Editoriale Verde Ambiente, Roma.

Schwartz, S.E. and Andreae M.O. (1996) Uncertainty in climatic change caused by aerosols. *Science* **272**, 5265, 1121–1122.

Serre-Bachet, F. (1994) Middle Ages temperature reconstructions: A focus on northeastern Italy—Climatic change. *In* M.K. Hughes and H.F. Diaz (eds.), *The Medieval Warm Period,* Special Issue, **26**(2–3):213–224

Shaw, E.M. (1993) *Hydrology in Practice.* Chapman & Hall.

Singer, S. F. (1996) Climate change and consensus. *Science* **271**, 5249, 581–582.

Singh, V. P. (1989) *Hydrologic Systems, II, Watershed Modeling.* Prentice Hall.

Tamburini, P. (1986) Il villaggio del Gran Carro: Conoscenze attuali e proposte di ricerca. *In* G.L. Carancini (ed.), *Atti dell'Incontro di Acquasparta 1985* "Gli Insediamenti Perilacustri dell'Età del Bronzo e della Prima Età del Ferro: il Caso dell'Antico Lacus Velinus", Perugia, 15–17 nov. 1985, *Quaderni di Protostoria, Istituto di Archeologia, Università di Perugia*, 213–238.

Thompson, L.G. and Mosley-Thompson, E. (1992) *Tropical Ice Core Paleoclimatic Records—Quelccaya Ice Cap, Peru, A.D. 470 to 1984.* Byrd Polar Research Center, Ohio State University.

Thornthwaite, C.W. and Mather, J.R. (1957) Introduction and tables for computing potential evapotranspiration and the water balance. Drexel Institute of Technology, Laboratory of Climatology, *Publications in Climatology* X(3), 185–311, Centerton, New Jersey.

Turc, L. (1954) Le bilan d'eau des sols. Relations entre les precipitations, l'evaporation et l'ecoulement. *Ann. Agron.* **4**.

Veggiani, A. (1994) I deterioramenti climatici dell'età del ferro e dell'alto medioevo, Torriceliana. *Bollettino Società Torriceliana di Scienze e Lettere* **45**, 3–80.

Walsh, J. (1993) The elusive Arctic warming. *Nature* **361**, 300–301.

Whyte, I. (1995) *Climatic Change and Human Society.* Arnold.

Zonneveld, K.A.F. (1996) Palaeoclimatic reconstruction of the last deglaciation (18–8 ka B.P.) in the Adriatic Sea region; a land-sea correlation based on palynological evidence. *Palaeo* Elsevier, **122**, 89–106.

Chapter 12

Frequency of Extreme Hydroclimatically-Induced Events as a Key to Understanding Environmental Changes in the Holocene

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12.1. Introduction

Our understanding of past frequencies of extreme hydroclimatically-induced events such as floods and mass wasting is mainly based on direct or indirect interpretations of sediments, landforms and parallel biotic changes. These sources of information point to episodes with higher or lower frequency of extreme events, and indicate that some of these occurred under wetter and cooler conditions that permitted higher water storage (*cf.* Starkel, 1983). It also seems true that greater year-to-year and multi-year variability have characterized those times of transition when water storage or runoff drastically adjust (*e.g.* ice sheet decay in North America; Teller, 1995). In the arid zone, such instability can involve high frequency of cyclones, and thus wetter conditions; in the semi-arid zone, in contrast, greater instability may cause expansion of the desert. In the monsoonal areas, floods seem to become more prevalent during warming trends (*e.g.* the early Holocene; Kutzbach, 1983). Transitions and phases with frequent events have thus played a major role in shaping the physical and biotic environment during the Holocene.

Before identifying those episodes of the Holocene with higher frequencies of extreme hydroclimatically-induced events, it is necessary (a) to determine the sedimentary and geomorphic signature of present-day extreme events; and (b) to describe the climatic conditions of the Little Ice Age—a recent episode partly covered by instrumental records and characterized by a high frequency of extreme climatic and hydrologic events.

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12.2. The Sedimentary and Geomorphic Record of Extreme Hydroclimatically-induced Events

Extreme hydrological events may recur on time scales ranging from intra-seasonal, to multi-annual, to secular and beyond; their magnitude may range from the moderate (two- or three-year recurrence interval) to rare (the 20–100 year event), to the cataclysmic (*cf.* Starkel, 1976; Baker, 1987). Their duration may range from short-lived single storms to circulation anomalies that persist for months and be repeated during several consecutive years. Moreover, a given event may be one of a series of hydrological departures. Depending on the climatic zone in which it occurs, it may be considered frequent or extremely rare. Such events, whether short- or long-lived, may exceed thresholds in water storage, runoff, and slope/channel stability. They will thereby produce a sedimentary and/or geomorphic record.

Table 12.1 presents various types of extreme hydroclimatically-induced events, and the sedimentological and geomorphic responses that can be expected in various climatic zones (Starkel, 1976; Baker *et al.*, 1988; Maizels, 1995). It shows how various kinds of landforms and deposits are respectively characteristic of, and caused by specific types of events. Note that in the temperate zone debris flows are mainly instigated by a sudden heavy downpours (Kotarba, 1992) while in the humid tropics they are most common following a period of continual rain combined with a

Table 12.1. Types of extreme hydrological events

Extreme meteo-hydrological events	Hydrological response	Geomorphic sedimentological response
heavy downpours	detention, overland flow, local floods	slope wash, gullying, stratified deluvia, debris flows, earth flows.
continuous rains	infiltration, underground storage, floods rise of groundwater and fills	earthslides, channel changes, fluvial cuts and fills, rise overbank deposits (debris flows in humid tropics).
rainy phases (years)	deep infiltration in the ground, rise of groundwater and lake levels	rockslides, earth creep calcareous tuffa, peat growth, transgreding lacustrine deposits
snowy winters	snow storage	growth of snow and ice fields
rapid snowmelts	runoff over frozen ground, floods	slope wash, solifluction, ice jams, overbank deposits
rainy years-decades (multiannual)	higher groundwater and lake level	advances of glaciers reactivation of old landslides
droughts (multiannual)	groundwater and lake level drop, dessication of rivers and bogs	salt crusts, soil formation over fluvial and lacustrine deposits

torrential interlude (Starkel, 1972a). In the temperate zone such continuous rains may produce earth slides and rock slides (Starkel, 1996a). But, if a particular region of the temperate zone within a brief interval of time is found to have experienced an unusually large number of debris flows, earth slides, and rock slides (clustering) it may be inferred that the span in question was characterized by spells of continuous rains punctuated by heavy downpours. Confidence in such a conclusion is increased if those same conditions can be corroborated from other proxy records (*e.g.* glacier advances; lake-level rises, *cf.* Starkel, 1996b). In the tropical arid zone changes in the pattern of runoff or water storage are liable to result from a single extreme event (Starkel, 1972b). To arrive at a correct explanation in a particular instance, all available paleohydrological records need to be taken into consideration (Starkel, 1996b).

12.3. The Distribution Pattern of Extreme Events in the 20th Century: The Polish Carpathians and Darjeeling Himalayas

Records from the Polish Carpathians (Starkel, 1996a) and the Darjeeling sector of the Himalayas (Starkel, 1972a; Froehlich and Starkel, 1995) provide evidence about extreme hydrological events which occurred during the 20th century. In the Carpathians, with an annual precipitation of 700–1,500 mm, localized heavy downpours characteristically have a random distribution in space and time (Fig. 12.1). The heaviest recorded 24-hour rainfall in the highest of the mountains (the Tatras) is 300 mm. But it is the downpours of short duration, with an intensity exceeding 1 mm per minute, that instigate debris flows. Such events have occurred several times this century (Kotarba, 1992).

Persistent rainstorms, yielding 200 to 500 mm of precipitation in two to four days, occasionally cover large areas. They can cause great floods and instigate many landslides—typically shallow features, or else reactivate older slides. Such areally extensive rains and floods occurred in this century in 1903, 1925, 1934, 1948, 1958, 1960, 1970, and 1980 (Fig. 12.2). Altogether in 17 years there were in various parts of Carpathians significant damage on slopes and in the valley floors with recurrence interval in any one area of 10–20 years. In the smaller catchments (100–300 km^2) distinct changes in channel morphology were observed every 2–5 years (Baumgart-Kotarba, 1983).

On two occasions during the past century, floods occurred in clusters that extended over consecutive years. The first of these occurred in the western Carpathians during 1958, 1959, and 1960 (Zietara, 1968); the second along the Ropa Valley of eastern Carpathians between 1970 and 1974 (Soja, 1977). In both cases the effect was marked by changes in slope and channel morphology, and a new trend in their evolution. The other extremes of rainy years are reflected in the large and deep rockslides which occurred in 1907, 1913, and 1974 (Gil and Starkel, 1979). The distribution of these years in time appears random.

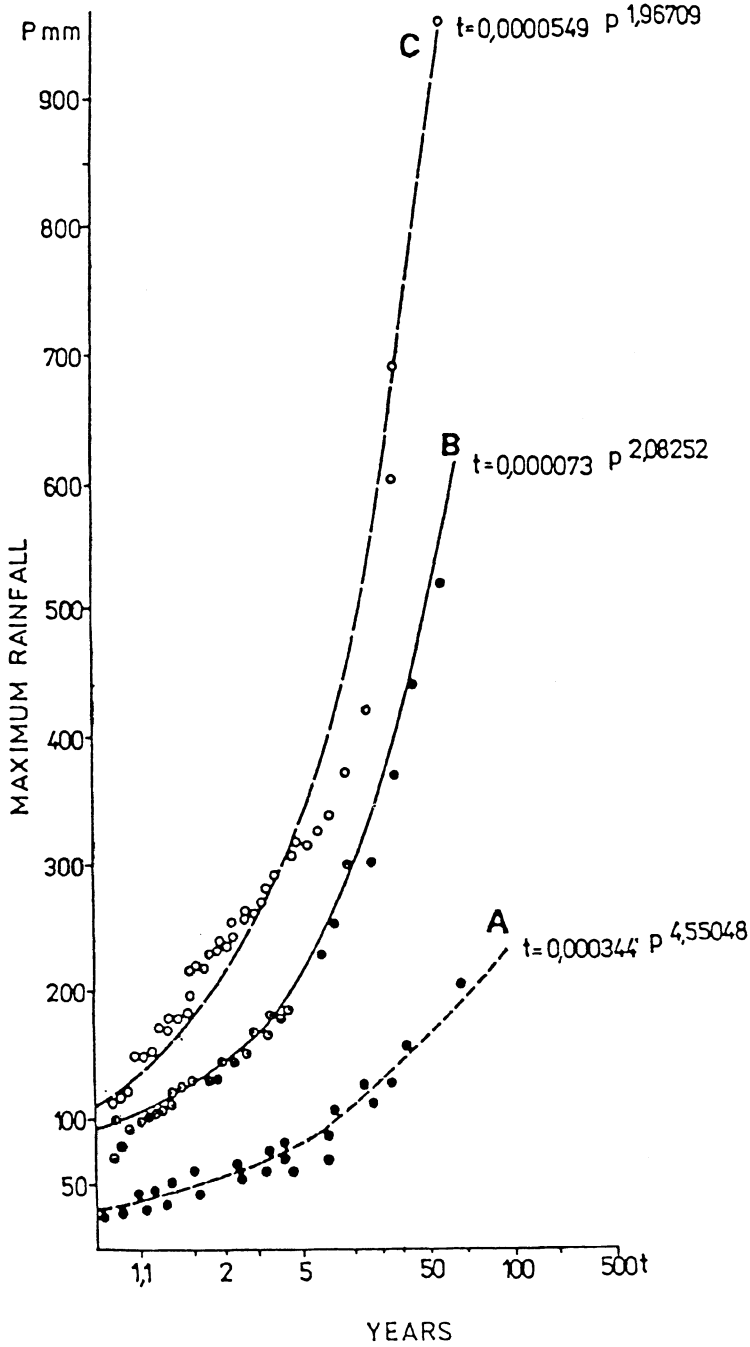


Figure 12.1. Recurrence interval for maximum rainfalls. A—24 hours rainfall in Labowa, Polish Carpathians (1969–1988), B—24-hours rainfall in Darjeeling, Sikkim Himalaya (1949–1986), C—three day continuous rain in Darjeeling (after Froehlich and Starkel, 1995).

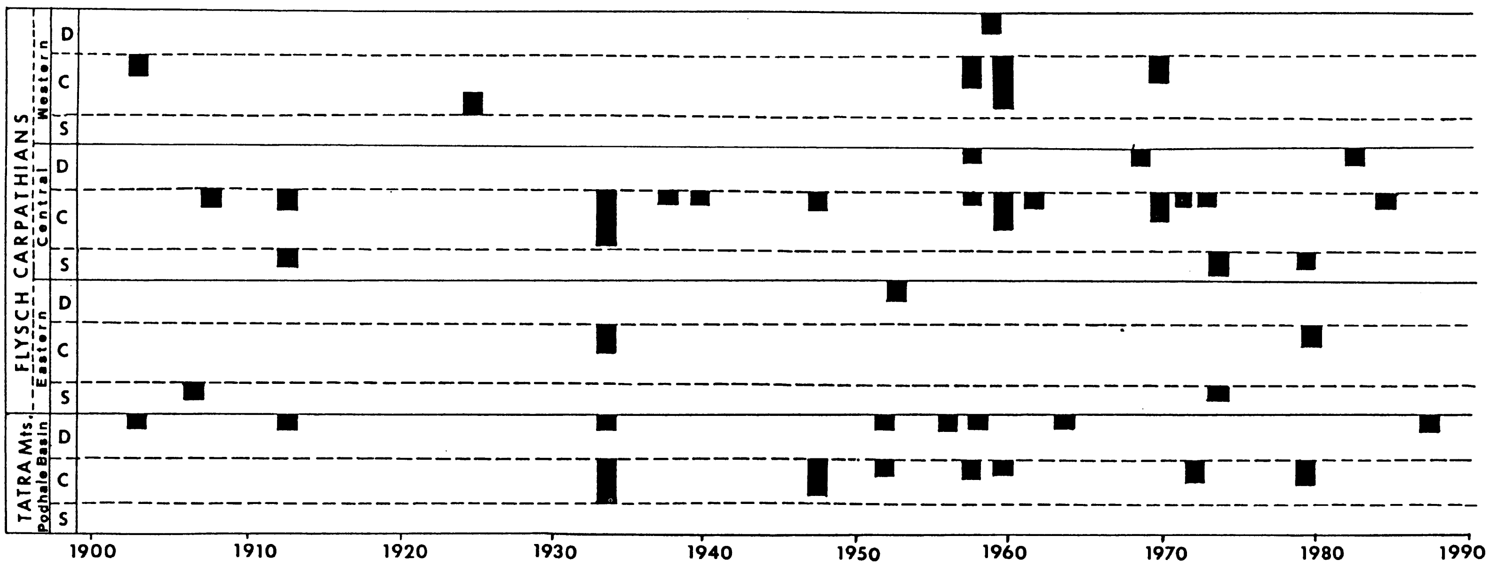


Figure 12.2. Extreme events in the 20th century recorded in various parts of the Polish Carpathians (after Starkel, 1996a, simplified). Signs: D—downpours, C—continuous rains, S—rainy seasons. 1. slope wash, 2. earthflows, 3. piping, 4. earthslides, 5. rockslides, 6. linear erosion, 7. increased channel erosion, 8. increased deposition—channel facies, 9. increased deposition—overbank facies, 10. debris flows.

The Darjeeling Himalayas, with a total annual precipitation of between 2,000 and 5,000 mm, is affected occasionally by local downpours that create mass movements on slopes. Very rarely more continuous rains reach 600–1,500 mm over periods of two to three days (Starkel, 1972a; Froehlich *et al.*, 1990). Such continuous rains, producing thousands of debris flows and catastrophic floods, were recorded in 1899, 1950, and 1968 (Fig. 12.1). Also here with only a century of instrumental record it is not possible to distinguish past episodes or trends. Luckily, the sedimentary and geomorphic records in many regions are clearly exposed. So in the future it should be possible to identify earlier episodes of extreme rainfall.

12.4. The Past 500 Years: Sedimentary and Geomorphic Responses to the Climate of the Little Ice Age

The written and, in due course, the instrumental records from the past several centuries show clearly that, relative to the past 150 years, the period between AD 1450–1850, the so-called Little Ice Age, was characterized by a markedly higher frequency of extreme hydroclimatically-induced events (Fig. 12.3). As documented in recent monographs (Grove, 1988; Borisenkov, 1988; Bradley and Jones; 1992), the Little Ice Age was a period of very unstable weather that produced frequent heavy rains and floods over much of the northern hemisphere (Pfister, 1992; Pavese *et al.*, 1992; Camuffo and Enzi, 1995; Lyakhov in: Borisenkov, 1988; Shvec, 1978; Baron *et al.*, 1992; Zhang and Wu, 1990; Zhang and Ge, 1990). Frequent floods through the 16th and 19th centuries in the Vistula Basin of Poland caused wholesale modification of the river, transforming it from a meandering to a braided channel (Klimek, 1974; Szumanski, 1977). Wider tree rings in the Mediterranean region and the Levant likewise point to times of frequent heavy rains (Guiot *et al.*, 1987; Liphschitz *et al.*, 1979; Serre-Backet *et al.*, 1992). In the high European mountains the thinner rings coincide with debris flows, avalanches and local floods (Bednarz, 1981; Grove, 1972; Kotarba, 1992). Unusual climatic events (heavy rains or whatever) were also recorded in the Tanaro Valley near Alessandria in the Italian Piedmont (Pavese *et al.*, 1992). These hydroclimatically-induced phenomena seem to have been concurrent throughout Europe, with multi-decade clusterings occurring immediately before AD 1600, around 1700, and between 1800 and 1820 (Fig. 12.3). These clusters seem to coincide with advances of mountain glaciers in Europe and elsewhere (*cf.* Grove, 1988) and with increases in the frequency of stronger El Niño events (Quinn and Neal, 1988). At the Quelcaya Ice Cap in Peru, the thicker annual laminae that accumulated during the 16th and 17th centuries coincide with extremely high annual ranges in the oxygen isotope values (Thompson, 1992). All of these data indicate that the Little Ice Age was characterized by a highly unstable climate with marked annual fluctuations and much clusterings of extreme events. Some paleoclimatologists draw attention to their coincidence with increased volcanic activity (Hammer *et al.*, 1980; Bradley and Jones, 1993).

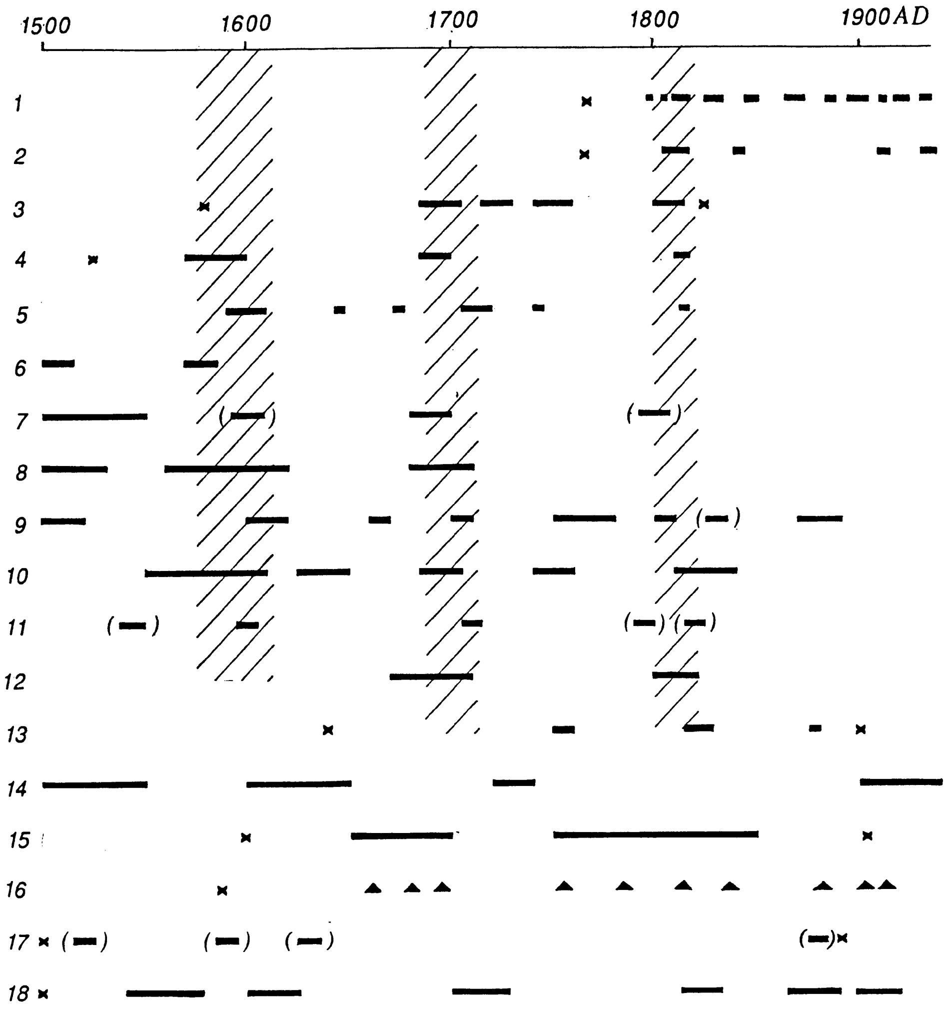


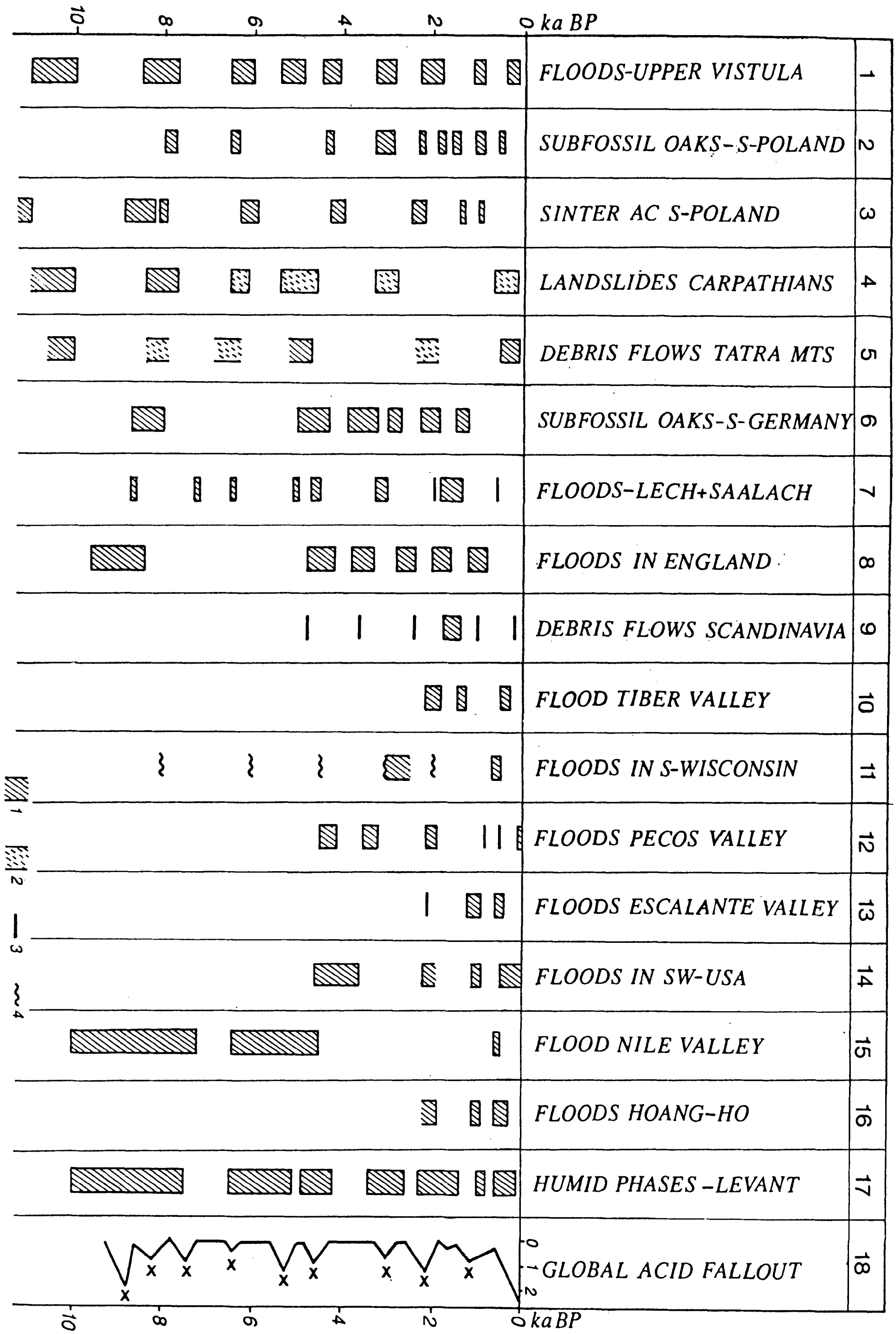
Figure 12.3. Rainfall anomalies during the Little Ice Age. 1. Frequent debris flows in Tatra Mts. (Baumgart-Kotarba and Kotarba, 1995), 2. Cool and wet years in the Tatra Mts.—based on tree-rings (Bednarz, 1981), 3. Frequent extreme events in Scandinavia (Grove, 1972, 1988), 4. Periods with frequent extreme rainfalls in Switzerland (Pfister, 1992), 5. Cool and wet years in Central Europe—based on tree-rings (Briffa and Schweingruber, 1993), 6. Floods in N. Italy (Pavese *et al.*, 1993), 7. Floods in Tiber Valley, Italy (Camuffo and Enzi, 1995), 8. Precipitation extremes in European Russia (Lyakhov, 1988), 9. Flood periods in the Dnieper valley (Shvec, 1978), 10. Wet periods in Mediterranean France based on tree-rings (Guiot, 1987), 11. Periods with heavy rainfalls in Morocco—based on tree-rings (Serre-Bachet *et al.*, 1993), 12. Periods with heavy rainfalls in Levant (Lipschitz, 1979), 13. Periods with frequent downpours and thunderstorms in Massachusetts (Baron, 1993), 14. Periods with extreme floods in China (Zhang and Wo, 1990; Zhang and Ge, 1990), 15. Half-centuries with frequent floods and long continuous rains in Japan (Yamamoto, 1971), 16. Volcanic eruptions—higher acid fallout (Hammer *et al.*, 1980), 17. Higher oscillations of ^{18}O and ice thickness in Quelecaya ice cap (Thompson, 1992), 18. Period with stronger El Niño (Quinn and Neel, 1993). In brackets—length of period not identified. Hatched-periods with frequent extreme rainfalls in Europe.

12.5. Other Holocene Episodes Characterized by a High Frequency of Extreme Hydrological Events

Scientists have long recognized there are episodes predating the Little Ice Age in which extreme hydroclimatically-induced events occurred at high frequency (Starkel, 1983, 1984; Knox, 1983; Bryson, 1989). These episodes are particularly well documented in western and central Europe, Eastern China, the southwestern and middlewestern United States, and the Nile Valley. The proxy records that document clustering of events remain sparse over time at least as compared to the scale of the Little Ice Age. In the Vistula Basin and other European basins, a number of different records can be used to reconstruct such events. These include rapid changes in fluvial systems (channel avulsion, a high rate of deposition on alluvial fans and point bars, frequent overbank deposition in slackwater areas and over peatbogs, concentrations of uprooted tree trunks buried in alluvial deposits), frequent landslides and debris flows, high growth rates of speleothems, *etc.* Such events might be further reflected in the rises of lake levels advances of mountain glaciers, high rate of deposition of calcareous tuffa and bogs and even in changes of plant communities (Starkel, 1996b, Fig. 12.4).

In the Vistula Basin the record of paleofloods is known in detail. Such activity clusters at the following intervals: 8,500–7,700 BP, 6,500–6,000 BP, 5,500–4,900 BP, 4,400–4,100 BP, 3,500–3,000 BP, 2,700–2,600 BP, 2,200–1,800 BP and dated by dendrochronological method 450–575 AD and 900–1100 AD (Starkel, 1983; Kalicki, 1991; Starkel, *et al.*, 1996). Channel avulsions tended to occur not through by one extreme event but during longer intervals of floods. In several localities we traced on both sides of the abandoned meandering system, the cut-off meanders formed during earlier stage of the phase of high flood frequency (Starkel *et al.*, 1991, 1996). After an avulsion the local flood tended to decline while a new meandering channel started to form (Fig. 12.5). At one locality in Zabierzów Bochenski, in the aban-

Figure 12.4. Phases with higher extreme rainfalls or flood frequency during the Holocene. 1. Higher fluvial activity in upper Vistula basin (Starkel, 1991, Starkel *et al.*, 1996; Kalicki, 1991); 2. Flood periods reconstructed from fossil oaks in upper Vistula basin (Krapiec, 1992); 3. Wet periods reconstructed from speleothems in S. Poland (Pazdur *et al.*, 1995); 4. Landslide phases in the Carpathians (Starkel, 1985, 1996c); 5. Debris flow phases in Tatra Mts. (Baumgart-Kotarba and Kotarba, 1993); 6. Flood periods reconstructed from fossil oaks in S. Germany (Becker, 1982); 7. Higher fluvial activity at the Alpine foreland (Schreiber, 1985 and others); 8. Higher fluvial activity in England (Needham and Macklin, 1992); 9. Debris flow phases in northern Sweden (Jonasson, 1993); 10. Flood periods in Tiber Valley (Camuffo and Enzi, 1995); 11. Discontinities and flood periods in S. Wisconsin valleys (Knox, 1983, 1993); 12. Floods in Pecos River, Texas (Kochel, 1988); 13. Floods in S.W. part of U.S.A. (Ely *et al.*, 1993); 14. Phases of frequent floods of Nile river (Said, 1993); 15. Phases of frequent floods in Huang-ho river (Zheng, 1985); 16. Humid phases in Levant (after Issar, 1995); 17. Global acid fallout—from Greenland (Hammer *et al.*,1980). Signs: 1. distinct phases; 2. less identified phases; 3. extreme events; 4. discontinuities in fluvial sequences.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FLOODS-UPPER VISTULA
SUBFOSSIL OAKS-S-POLAND
SINTER AC S-POLAND
LANDSLIDES CARPATHIANS
DEBRIS FLOWS TATRA MTS
SUBFOSSIL OAKS-S-GERMANY
FLOODS-LECH+SAALACH
FLOODS IN ENGLAND
DEBRIS FLOWS SCANDINAVIA
FLOOD TIBER VALLEY
FLOODS IN S-WISCONSIN
FLOODS PECOS VALLEY
FLOODS ESCALANTE VALLEY
FLOODS IN SW-USA
FLOOD NILE VALLEY
FLOODS HOANG-HO
HUMID PHASES -LEVANT
GLOBAL ACID FALLOUT
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doned channel three younger flood phases are recorded (Kalicki *et al.*, 1996). Some of the alluvium deposited during times of inferred flooding is choked with the trunks of oak trees. These particular deposits have been radiocarbon dated at 3,200–3,000 BP and dendroclimatically dated at 225 BC–325 AD, 425–575 AD, 900–1150 AD, 1200–1325 AD and from transition 15th to 16th century (Krapiec, 1992). At some sites it is possible to count more than 100 flood events within the interval 8400 to 7800 BP (Niedzialkowska *et al.*, 1977, Czyzowska and Starkel in: Starkel *et al.*, 1996). Also within an early medieval phase of oak deposition were discovered the clusterings of trunks in one or two decades interpreted as deposited by single extreme floods (Krapiec in: Starkel *et al.*, 1996). Studies on landslides (Starkel, 1985, 1996c) and debris flows (Kotarba, 1996; Baumgart-Kotarba and Kotarba, 1993) in the Carpathians have shown similar time of episodes higher frequency (Fig. 12.4). Also the dated speleothems from several caves in the Cracow Upland present a polymodal distribution in time (Pazdur *et al.*, 1995) with culminations after 8,500, about 6,500, between 5,000 and 4,500, and *ca.* 2,500 radiocarbon years BP.

On the Alpine foreland of southern Germany, Schreiber (1985) recognized that fluvial activity coincided with the deposition and burial of oak trunks (Becker, 1982). A similar coincidence is found elsewhere in Europe (Starkel, 1995) and British Isles (Needham and Macklin, 1992). In the Tiber Valley near Rome, as in the valleys of central Europe, intervals of frequent floods occurred from 2,050 to 1,800 BP, and from 1,450 to 1,250 BP (Camuffo and Enzi, 1995). Also in the mountains of Scandinavia were recorded several distinct episodes of snow avalanche activity (Blikra and Nemec, 1993) and rapid mass wasting (Jonasson, 1993).

In the midwest of North America, Knox (1983; 1993) documents paleofloods from the record of overbank deposits. According to his analysis, the incidence of flooding increased after 3,000 BP and continued till 1,800 BP and then increased again between 1,200 and 800 BP. In the desert southwest of that continent, the Pecos River of Texas shows clusters of floods beginning from 4,450 BP, after 3,200 BP, and around 2,000 BP (Kochel, 1988). Nearby, the Escalante River of Utah experienced clusters of extreme events around 2,100 BP, between 1,200 and 980 BP, and between 1350 and 1550 AD. In general in the dry southwest, high flood frequencies occurred between 4,600–3,600 BP, around 1,000 BP, and some time after 500 BP (Ely *et al.*, 1993).

The Nile River, whose flow and flood are dependent on monsoonal circulation, experienced a decline in flood frequency around 4,500 BP, but after historical records it increased during the period 1000–1200 AD and again between 1350–1450 AD (Said, 1993). On the Hoang-ho River of eastern China, flooding was frequent before 1,800 BP, between 1,140 and 900 BP, and between 670 and 270 BP (Zheng, 1985).

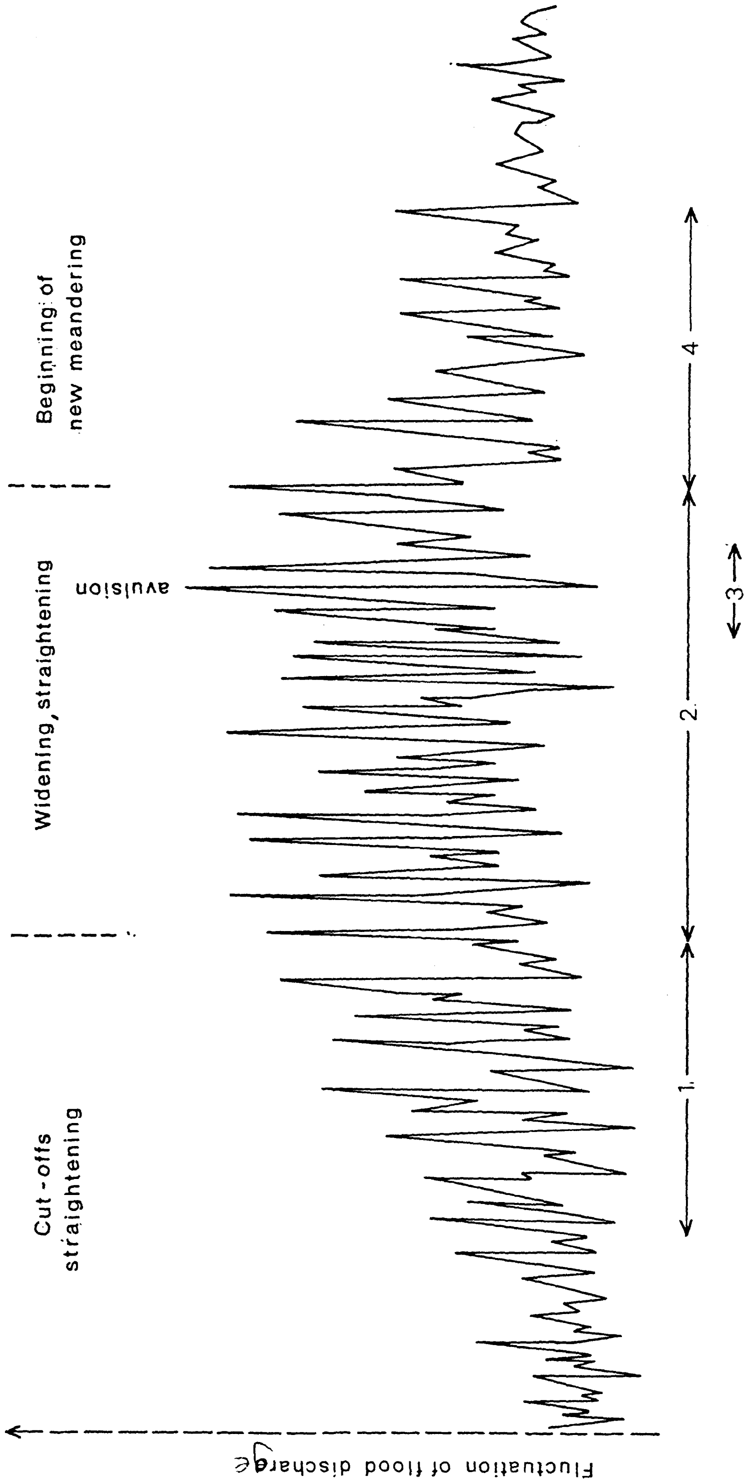

Figure 12.5. Model of phase with more frequent extreme floods during the Holocene, reflected in the evolution of the fluvial system (4 phases distinguished).

12.6. Conclusions

As one considers the phases with high frequency of heavy rainfalls and floods in various parts of the globe, it is remarkable how they not only reflect general changes in the climatic circulation pattern but also show a great concurrence of shorter episodes (Figs. 12.3, 12.4).

The first cause which should be considered are the fluctuations in the solar activity. Their minima coincide with positive ^{14}C anomalies (Stuiver *et al.*, 1991). At such times with the cooler and wetter climate, glaciers advance (Röthlisberger, 1986) and lake levels rise (Magny, 1993). The widespread sudden wetting of European raised bogs *ca.* 830 BC coincide with the advent of a solar minimum (Kilian *et al.*, 1995). But the extreme hydroclimatically-induced events do not correlate everywhere with these cooler and wetter phases. Overall as the second of the main causes of this coincidence, the volcanic activity may be more pronounced (Hammer *et al.*, 1980; Bryson, 1989; Nesje and Johannessen, 1992). Indeed, heavy rainfalls are known to have followed the great volcanic eruptions of recent centuries (Grove, 1988). Changes in the frequency of volcanic events, superimposed on the global and regional climatic fluctuations, may thus cause the exceeding of thresholds and disturbance of equilibrium of slopes and river channels as well as of ecosystems as a whole including the shifting of ecotones. Especially evident are the consequences of changes in the frequency of heavy rains in areas characterized by a deficit of water (Rognon, 1987; Said 1993; Baker *et al.*, 1995).

The investigation of the frequency of extreme hydrological events and their clusterings may serve in better understanding of mechanisms of environmental changes during the Holocene.

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References

Baker, V.R. (1987) Palaeoflood hydrology and extraordinary flood events. *Journal of Hydrology* 96, 79–99.

Baker, V.R., Bowler, J.M., Enzel, Y., and Lancaster, N. (1995) Late Quaternary palaeohydrology of arid and semi-arid regions. *In* K.J. Gregory, L. Starkel and V.R. Baker (eds.), *Global Continental Palaeohydrology*, J. Wiley & Sons, pp. 203–231.

Baker, V.R., Kochel, R.C. and Patton, P.C., eds. (1988) *Flood Geomorphology*. J. Wiley & Sons.

Baron, W.R. (1992) Historical climate records from the northern United States, 1640 to 1900. *In* R.S. Bradley and P.D. Jones (eds.), *Climate Since A.D. 1500*. Routledge, London and New York, pp. 74–91.

Baumgart-Kotarba, M. (1983) Kształtowanie koryt i teras rzecznych w warunkach zróżnicowanych ruchów tektonicznych (na przykładzie Wschodniego Podhala). *Prace Geograficzne IGiPZ PAN* 145, 1–33.

Baumgart-Kotarba, M. and Kotarba, A. (1993) Późnoglacjalne i holoceńskie osady z Czarnego Stawu w Tatrach. *Dokumentacja Geograficzna IGiPZ PAN* 4–5, 9–30.

Becker, B. (1982) Dendrochronologie und Paläoekologie subfossilen Baumstämme aus Flussablagerungen: Ein Beitrag zur nacheiszeitlichen Auenentwicklung im sudlichen Mitteleuropa. *Mitteilungen der Komission für Quarterforschung der Osterreichischen Akademie der Wissenschaften* 5, 1–120.

Bednarz, Z. (1981) Relationship of tree-ring widths in the Tatra Mountains to variations in monthly temperature and precipitation. *Mitteilungen Forstlichen Bundesversuchanstalt* 142, 367–389.

Blikra, L.H. and Nemec, W. (1993) Postglacial avalanche activity in western Norway: Depositional facies sequences chronostratigraphy and palaeoclimatic implication. *Paleoklimaforschung* 11, 143–162. Spec. Issue ESF Project European Palaeoclimate and Man, 6, G. Fischer Verlag, Stuttgart.

Borisenkov, Ye.P., ed. (1988) *Climate Oscillations of the Last Millennium*. Gidrometeozdat, Leningrad.

Bradley, R.S. and Jones, P.D., eds. (1992) *Climate Since A.D. 1500*. Routledge, London and New York.

Bradley, R.S. and Jones, P.D. (1992) Records of explosive volcanic eruptions over the last 500 years. *In* R.S. Bradley and P.D. Jones (eds.), *Climate Since A.D. 1500*, pp. 606–662.

Briffa, K.R. and Schweingruber F.H. (1992) Recent dendroclimatic evidence of northern and central European summer temperatures. *In* R.S. Bradley and P.D. Jones (eds.), *Climate Since A.D. 1500*, pp. 366–392.

Bryson, R.A. (1989) Late Quaternary volcanic modulation of Milankovitch climate forcing. *Theoretical and Applied Climatology* 39, 115–125.

Camuffo, D. and Enzi, S. (1995) Climatic features during the Sporer and Maunder Minima. *Paleoklimaforschung* 16, 105–124. Spec. Issue ESF Project European Palaeoclimate and Man, G. Fischer Verlag, Stuttgart.

Ely, L.L., Baker, V.R., and Cayan, D.R. (1993) A 5,000 year record of extreme floods and climate change in the south-western United States. *Science* 262, 410–412.

Froehlich, W., Gil E., Kasza I., and Starkel, L. (1990) Thresholds in the transformation of slopes and river channels in the Darjeeling Himalaya, India. *Mountain Research and Development* 10(4), 301–312.

Froehlich, W. and Starkel, L. (1995) The response of slope and channel systems to various types of extreme rainfall. A comparison between the temperate zone and humid tropics. *Geomorphology* 11, 337–345.

Gil, E. and Starkel, L. (1979) Long-term extreme rainfalls and their role in the modelling of flysch slopes. *Studia Geomorphologica Carpatho-Balcanica* 13, 207–220.

Grove, J.M. (1972) The incidence of landslides, avalanches and floods in western Norway during the Little Ice Age. *Arcitic and Alpine Research* 4, 131–138.

——. (1988) *The Little Ice Age*. Methuen.

Guiot, J., Pons, A., de Beaulieu, J.L., and Reille, M. (1987) A 140,000 year continental climate reconstruction from two European pollen records. *Nature* 338, 309–313.

Hammer, C.U., Clausen, H.B., and Dansgaard W. (1980) Greenland ice sheet evidence of post-glacial volcanism and its climatic impact. *Nature* **288**, 230–255.

Issar, A.S. (1995) Impacts of climate variations on water management and related socio-economic systems. IHP-IV Project H-2.1, Unesco, Paris, 1–97.

Jonasson, C. (1993) Holocene debris-flow activity in northern Sweden. *Paleoklimaforschung* 11, 179–195. Spec. Issue Project European Palaeoclimate and Man, **6**, G. Fischer, Verlag, Stuttgart.

Kalicki, T. (1991) The evolution of the Vistula river valley between Cracow and Niepotomice in Late Vistulian and Holocene times. *Geographical Studies* **6**, 11–37. Spec. Issue, Warszawa.

Kalicki, T., Starkel, L., Sala, J., Soja, R., and Zernickaya, V.P. (1996) Subboreal palaeochannel system in the Vistula valley near Zabierzów Bocheński (Sandomierz Basin). *Geographical Studies* **9**, 129–158. Spec. Issue, Continuo, Wrocław.

Killian, M.R., van der Plicht J., and van Geel B. (1995) Dating raised bogs: New aspects of AMS wiggle matching, a reservoir effect and climatic changes. *Quaternary Science Review* **14**, 959–966.

Klimek, K. (1974) The structure and mode of sedimentation of the flood-plain deposits in the Wisłoka-valley (South Poland). *Studia Geomorphologica Carpatho-Balcanica* **8**, 137–151.

Knox, J.C. (1983) Responses of river systems to Holocene climates. *In* H.E. Wright (ed.), *Late-Quaternary Environments of the United States* **2**, Univ. of Minnesota Press, pp. 26–41.

——. (1993) Large increases in flood magnitude in response to modest changes in climate. *Nature* **361**, 430–432.

Kochel, R., Craig (1988) Extending stream records with slackwater paleoflood hydrology: Examples from West Texas. *In* V.R. Baker, R.C. Kochel and P.C. Patton (eds.), *Flood Geomorphology*, J. Wiley, pp. 377–391.

Kotarba, A. (1992) High-energy geomorphic events in the Polish Tatra Mountains. *Geografiska Annaler* **48A**, 2–3, 123–131.

——. (1993–94) Zapis małej epoki lodowej w osadach jeziornych Morskiego Oka w Tatrach Wysokich. *Studia Geomorphologica Carpatho-Balcanica* **27–28**, 61–69.

——. (1996) Sedimentation rates in the High Tatra lakes during the Holocene—Geomorphic interpretation. *Studia Geomorphologica Carpatho-Balcanica* **30**, 51–61.

Krapiec, M. (1992) Late Holocene tree-ring chronologies of South and Central Poland. *Geologia* 18(3), 37–119, Wyd. AGH, Kraków.

Kutzbach, J.E. (1983) Monsoon rains of the Late Pleistocene and early Holocene: Pattern intensity and possible causes of changes. *In* A. Street-Perrot *et al.* (eds.), *Variations in the Global Water Budget*, D. Reidel Publ., Dordrecht, pp. 371-389.

Liphschitz, N., Waisel, Y., and Lev-Yadum S. (1979) Dendrochronological investigations in Iran. *Tree Ring Bulletin* **39**, 39–45.

Magny, M. (1993) Holocene fluctuations of lake levels in the French Jura and sub-Alpine ranges and their implications for past general circulation pattern. *The Holocene* 2, 79–84.

Maizels, J.K. (1995) Palaeohydrology of polar and subpolar regions over the past 20,000 years. *In* K.J. Gregory, L. Starkel and V.R. Baker, *Global Continental Palaeohydrology*, J. Wiley & Sons, pp. 259–299.

Needham, S. and Macklin, M.G., eds. (1992) *Alluvial Archaeology in Britain. Oxbow Monograph* 27, Oxford.

Nesje, A. and Johannesson, T. (1992). What were the primary forcing mechanisms of high-frequency Holocene glacier and their implications for past general circulation pattern. *The Holocene* 2, 79–84.

Niedzialkowska, E., Skubisz, A., and Starkel, L. (1977) Lithology of the Eo- and Mesoholocene alluvia in Podgrodzie upon Wisłoka river. *Studia Geomorphologica Carpatho-Balcanica* 11, 89–100.

Pavese, M.P., Banzon, V., Colacino, M., Gregori, G.P., and Pasaqua, M. (1992) Three historical data series on floods and anomalous climatic events in Italy. *In* R.S. Bradley and P.O. Jones (eds.), *Climate Since A.D. 1500*, Routledge, London and New York, pp. 155–170.

Pazdur, A., Pazdur, M.F., Pawlyta, J., Górny, A., and Olszewski, M. (1995) Paleoclimatic implications of radiocarbon dating of speleothems from the Cracow-Wieluń Upland, Southern Poland. *Radiocarbon* 37, 2.

Pfister, C. (1992) Monthly temperature and precipitation in central Europe 1525–1979 quantifying documentary evidence on weather and its effects. *In* R.S. Bradley and P.D. Jones (eds.), *Climate Since A.D. 1500*, Routledge, London and New York, pp. 118-142.

Quinn, W.H. and Neal, V.T. (1992) The historical record of El Niño events. *In* R.S. Bradley and P.D. Jones (eds.), *Climate Since A.D. 1500*, Routledge, London and New York, pp. 623-649.

Rognon, P. (1987) Aridification and abrupt climatic events on the Saharan northern and southern margins, 2,000 yrs BP to present. *In* W.H. Berger and L.D. Labeyrie (eds.), *Abrupt Climatic Change*, Dordrecht, Reidel, pp. 209–220.

Röthlisberger, F. (1986) *10000 Jahre Gletschergeschichte der Erde*. Aarau, Sauerlander.

Said, R. (1993) *The River Nile: Geology, Hydrology and Utilization*. Pergamon Press, Oxford.

Schreiber, U. (1985) Das Lechtal zwischen Schongau und Rain im Hoch-, Spät- und Postglazial. *Geolog. Institute Univ. Köln, Sonderveröfentlichungen* 58, 1–192.

Schwetz, G.I. (1978) Multi-centennial changes of the run-off of the Dniepr River. Gidrometeoidat, Leningrad (in Russian).

Serre-Bachet, F., Guiot, J., and Tessier, L. (1992) Dendroclimatic evidence from Southern Europe and northwestern Africa. *In* R.S. Bradley and P.D. Jones (eds.), *Climate Since A.D. 1500*, pp. 349–365.

Soja, R. (1977) Deepening of channel in the light of the cross profile analysis (Carpathian river as example). *Studia Geomorphologica Carpatho-Balcanica* 11, Kraków, 127–138.

Starkel, L. (1972a) The role of catastrophic rainfall in the shaping of the relief of the Lower Himalaya (Darjeeling Hills). *Geographia Polonica* 21, 103–147.

——. (1972b) The modelling of monsoon areas of India as related to catastrophic rainfall. *Geographia Polonica* 23, 151–173.

——. (1983) The reflection of hydrologic changes in the fluvial environment of the temperate zone during the last 15,000 years. *In* K.J. Gregory (ed.), *Background to Palaeohydrology: A Perspective*, J. Wiley & Sons, pp. 213–235.

——. (1984) The reflection of abrupt climatic changes in the relief and in the sequence of continental deposits. *In* N.A. Mörrner and W. Karlen (eds.), *Climatic Changes on a Yearly to Millennial Basis*, Reidel Publ. Comp., Dordrecht, pp. 135–146.

——. (1985) The reflection of the Holocene climatic variations in the slope and fluvial deposits and forms in the European mountains. *Ecologia Mediterranea* 11(1), 91–98.

——. (1991) The Vistula River Valley: A case study for Central Europe. *In* L. Starkel, K.J. Gregory and J.B. Thornes (eds.), *Temperate Palaeohydrology*, J. Wiley & Sons, pp. 171–188.

——. (1994) Frequency of flood during the Holocene in the Upper Vistula Basin. *Studia Geomorphologica Carpatho-Balcanica* 27–28, 3–13.

——. (1995) Palaeohydrology of the temperate zone. *In* K.J. Gregory, L. Starkel and V.R. Baker, *Global Continental Palaeohydrology*, J. Wiley & Sons, pp. 233–257.

——. (1996a) Geomorphic role of extreme rainfalls in the Polish Carpathians. *Studia Geomorphologica Carpatho-Balcanica* 30, 21–38, Kraków.

——. (1996b) Palaeohydrological reconstruction: Advantages and disadvantages. *In* J. Branson, A.G. Brown and K.J. Gregory (eds.), *Global Continental Changes: The Context of Palaeohydrology*, Geological Society Special Publications 115, pp. 9–17.

——. (1996c) Mass movements during the Holocene: Carpathian example and the European perspective. *Proceedings of ESF Symposium, Paleoclimate and Man*, Stuttgart.

Starkel, L., Gębica, P., Niedzialkowska, E., and Podgórska-Tkacz, A. (1991) Evolution of both the Vistula floodplain and late glacial-early Holocene palaeochannel systems in the Grobla Forest (Sandomierz Basin). Evolution of the Vistula river valley during the last 15,000 years, part IV. *Geographical Studies* 6, 87–99, Spec. Issue .

Starkel, L., Kalicki, T., Krapiec, M., Soja, R., Gębica, P., and Czyżowska, E. (1996) Hydrological changes of valley floor in the upper Vistulian Basin during Late Vistulian and Holocene. *Geographical Studies* 9, 7–128. Spec. Issue, Continuo, Wrocław.

Stuiver, M., Braziunas, T.F., Becker, B., and Kromer, B. (1991) Climatic, solar, oceanic and geomagnetic influences on Late-Glacial and Holocene atmospheric $^{14}C/^{12}C$ change. *Quaternary Research 35*, 1–24.

Szumański, A. (1977) Changes in the course of the Lower San channel in XIX and XX centuries and their influence on the morphogenesis of its floodplain. *Studia Geomorphologica Carpatho-Balcanica* 11, 139–154, Kraków.

Teller, J.T. (1995) The impact of large ice sheets on continental palaeohydrology. *In* K.J. Gregory, L. Starkel and V.R. Baker (eds.), *Global Continental Palaeohydrology*, J. Wiley & Sons, pp. 109–129.

Thompson, L.G. (1992) Ice core evidence from Peru and China. *In* R.S. Bradley and P.D. Jones (eds.), *Climate Since A.D. 1500*, Routledge, London and New York, pp. 517-548.

Yamamoto, T. (1971) On the climatic change in the XV and XVI centuries in Japan. *Geophysical Magazine* 35, 187–206.

Webb, R.H., O'Connor, J.E., and Baker, V.R. (1988) Palaeohydrologic reconstruction of flood frequency on the Escalante river, south-central Utah. *In* V.R. Baker, R.C. Kochel, and P.C. Patton (eds.), *Flood Geomorphology*, J. Wiley & Sons, pp. 403-420.

Zhang Peiyuan and Ge Quansheng (1990) Abrupt climatic change: Introduction and a case study. Beijing Internat. Symposium Climatic Change and their Impacts, reprint, Nat. Sc. Foundation of China, Beijing C-1.

Zhang Peiyuan and Wku Xiangding (1990) Regional response to global warming: A case study in China. *In* Yino Yuan Li (ed.), *World Laboratory and CCAST Workshop Series*, Gordon and Breach Scient. Pub. 5, pp. 26–41.

Ziętara, T. (1968) Rola gwaltownych ulew i powodzi w modelowaniu rzeźby Beskidów. *Prace Geograficzne IG PAN* 60, 116, Warszawa.

Chapter 13

The Impact of Climate Changes on Groundwater Regimes and Resources in Russia

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Changes in climate, always involving the water cycle, have occurred throughout the observable geological history of inner Eurasia. One peak was reached near the Pliocene/Pleistocene boundary (*ca.* 2 million years BP). Millions of square kilometers were then affected by glacial erosion or deposition. Successive advances or retreats of sheet ice repeatedly altered the hydrological network around this time and, of course, subsequently. The accumulation of ice diminished riverine run-off. Thawing naturally increased it, often very sharply. Pronounced upper valley erosion and lower valley aggradation could occur during thaws. Witness the valleys buried by aggradation in the White and Black Sea basins.

By the 'groundwater drainage base' is meant the altitude up to which rivers can be recharged from subterranean supply. In the middle Pliocene (*i.e.* from 6.0–4.5 million years BP) this altitude fell by over 325 meters in the Kama river basin (61°N; 55°E). Less extreme but still highly significant fluctuations have occurred throughout the Pleistocene and Holocene, the Quaternary period as they are together known. Palynological, isotopic, glaciological and geomorphological investigations have allowed us to reconstruct with some confidence trends during the Holocene (the last 10,000 years) in temperature, precipitation, lake levels, and humidity in different parts of Russia or the Former Soviet Union.

Such palaeo-reconstructions have very consistently exhibited the following regularities:

1. During times of general warming globally, this warming has tended to be most pronounced in high latitudes while seemingly negligible in low ones. It has also been manifested more strongly in winter than in summer.

A. S. Issar and N. Brown (eds.), Water, Environment and Society in Times of Climatic Change, 289–293.

2. During times of continental warming across Russia, increases in precipitation are recorded in the North and, indeed, all coastal territories. Decreases are observed in the Southern Interior.
3. The concentrations in the atmosphere of those gases (most notably, carbon dioxide, CO_2, and methane, CH_4) prone to produce a 'greenhouse effect' has coincided closely with times of general warming. The cause-and-effect relationship has yet to be elucidated fully.
4. Times of continental warming have witnessed high levels in the ground water and in lakes.

Since the abrupt acceleration in the rate of increase of atmospheric carbon dioxide *ca.* 1955–1960, upward trends are to be observed across Russia in groundwater levels and discharge rates. The corresponding trends in the winter run-off, as caused by groundwater discharge, have been especially marked in the drainage areas for tributaries of the Volga, mainly in the hills north-west of Moscow. In all cases, too, the graphs show a secular increase by comparison with the interwar years. In several instances, however, the evidence is of a slow down or reversal of trend from around 1950 or 1960 to perhaps 1970 or 1975, this being followed by a strong recovery. That arresting of the secular trend coincides quite closely with the retardation or, in the Northern Hemisphere, slight reversal of global warming long remarked as occurring between *ca.* 1940 and *ca.* 1965.

Analyses of tendencies in the main elements in the water balance have also been carried out for the number of karst areas on or peripheral to the Russian Platform: the Ishorsk and Ufimsk plateaux, the Sredne-Russian elevation, the Western Ukraine, Mountainous Crimea and Western Caucasus (Kovalevsky and Yefremenko, 1995). In all these areas, except the Crimea, upward trends are to be observed in respect of precipitation, surface run-off and groundwater discharge the last several decades. The downward tendency in the Crimea was evident, for example, in most of the yields from springs, 48 out of 52. Below ground in the central part of the Moscow region, there is a cone of dryness because confined reserves of groundwater are being intensively exploited. However, that little affects the wider picture.

As already implied, the exact causal relationships have still to be elucidated. Meanwhile, it is salutary to remember that changes in the concentration of the greenhouse gases have occurred throughout the history of life on Earth. Thus for much of the Archaean (*i.e.* 2.5–3.5 billion years ago) it was at several times today's level. Much nearer the present, a rise in global temperature of 1°C was reached during the Holocene optimum, 5–6 thousand years BP. One of 2°C was attained during the Pleistocene optimum: that is to say, during the Mikulinsk or Samsk interglacial some 120–130 thousand years ago. One of 3.0–3.5°C would match that of the Pliocene optimum, *ca.* 3.3–4.4 million years ago. As things stood in 1990, the first of these thresholds was due to be crossed in 2000–2010; the second around 2025–2030; and the third after 2050 (Budyko *et al.*, 1991).

For Russia a global warming of one degree can be expected to involve a temperature increase, in summer and winter, of 2–4° in the North; and 0.5–1.5°C in middle and southerly latitudes. Annual precipitation will increase 50–100 mm in the North while in the South it will decrease 25–50 mm.

By the year 2025, the carbon dioxide content of the atmosphere was liable to have doubled. So according to M.I. Budyko and colleagues *(ibid.)*, the consequent rise in mean air temperature of 3.0–3.5°C globally could involve a rise in parts of the Arctic basin by as much as 12–15°C in July and by 15–20°C in January. In middle Russia, it would be higher than now by 2–5°C in July and by 10–15°C in January. Thus the zero isotherm for January would retreat 10–15° northwards.

Statistically confirmed correlations can be used to model the hydrological implications, given that annual precipitation in the northern part of the country could increase by 200–600 mm and in the southern by 50–200 mm. Actual and potential evaporation would increase as well. The margin of precipitation over potential evaporation would be positive in high latitudes but negative in the South, causing humidification in the former and aridization in the latter. The increase in annual 'effective precipitation', that which contributes to surface run-off and groundwater recharge, may range from 200–300 mm in the North to between 0 and 50 mm in South Russia.

Time lags are among the additional parameters that need to be fed in to assess the impact of mankind, acting both locally and globally, on specific catchment areas, urban territories *etc.* The methodological aspects have already been well covered in the literature (Gavitch, 1980; Kovalevsky, 1974; Bredehoeft *et al.*, 1982; Anderson *et al.*, 1988).

Moreover, a prognostic review of transformations in groundwater resources across Russia as a whole has now been conducted (Kovalevsky and Maximova, 1988; Kovalevsky, 1994; Kovalevsky and Yefremenko, 1995). It shows that, with global warming reaching 1°C in the first decade of the next century, there will be no great changes in groundwater recharge (Kovalevsky and Maximova, 1988). In the main, the increases will not exceed 10%. Only in part of Eastern Siberia will they be 20–30% above the present norm. Nevertheless, signs will already be evident of a contrast between a groundwater increase in the North and a decrease in the South and South-West, a contrast well anticipated in existing long series of data.

A doubling of the level of atmospheric carbon dioxide has been calculated to increase groundwater resources by 20–30% overall. Should this happen, the most significant changes (up to 40%) are to be expected along the Arctic coast of Siberia and in Middle Asia. More focused predictions using models with concentrated parameters have been made for the Volga basin, home of roughly half the whole population of Russia. So have they been for six of the karst zones on or near the Russian Platform. Karst hydrology being especially sensitive to climate change.

Maximum increments were to be observed in two littoral Karst locations, the Izhorsk plateau and the Crimea. At the same time, the seasonally low rate of discharge for groundwater has been forecast to show, proportionately speaking, the biggest rise of all. The increases anticipated are typically between 10% and 35% while on the Izhorsk plateau, the rate could even double.

All of which points towards a need to refine further our ability to predict what changes are possible in groundwater regimes regionally and what impact these may make intra-annually. Probability graphs for monthly values for, say, groundwater levels or spring yields can be useful, especially if gauged in relation to human demand. The saltiness of the groundwater can become a very relevant criterion.

Regarding the Volga basin, prognoses of change in surface flow and groundwater drainage were carried out for eight catchment areas with varying geological and climatic circumstances, but all set against the background of a doubling in atmospheric carbon dioxide. Using a model with grouped parameters, it was confirmed that by no means all the effective precipitation would be spent in surface run-off. Between 28 and 80% would go towards the accumulation of groundwater. This would result in a general rise in groundwater levels of between 10 and 50 centimeters in sand and between 20 and 110 centimeters in the porous loams.

There would be diverse consequences, ecologically and economically. A CO_2 doubling might in itself raise agricultural productivity as much as 20% (Budyko *et al.*, 1978). The margins of stable agriculture could advance northwards by up to 5° or 7°. A reduction of about a third in the permafrost zone was to be expected.

Power consumption for heating would be reduced, a very important gain for Russia. A rise in river discharge in the low season would improve navigation as well as public water supply and the power output from small rivers. All these conditions would favor the settlement of the North.

On the other hand, an increase in groundwater through more vigorous recharge might cause a variety of particular problems in North Russia where, what with permafrost and old shield bedrock, the groundwater reserves usually extend to no great depth and where perennially water tables tend to be close to the surface and swampy conditions prevalent. Under these circumstances, water logging might be even more common with the consequent disruption of ecosystems. So may landslides, mudflows and subsidence. Soil productivity may be reduced through gleying. Such coastal locations as St. Petersburg and Archangelsk could be further threatened by rises through ice cap melt of mean sea levels.

Conversely, in the hot and arid South, a decrease in low water riverine discharges would be expected. So would a decrease in soil water by typically a third. So, too, would a spreading of deserts; and a rise, maybe of a third, in drought frequency. Saline water could likewise become more of a problem.

Yet none of these predictions can be made sufficiently definitive and accurate unless the General Circulation Models of the atmosphere and ocean are well constructed and well inputted. Yet they need to incorporate many natural variables and to take cognizance as well of all the vagaries of human interventions and human responses. Historical perspectives may be more than averagely important in the hydrological sphere.

References

Anderson, M.G. and Burt, T.P.,eds. (1988) *Hydrogeological Forecasting*. A. Wiley.

Budyko, M.I., Vinnikov, K.Y., Drozdov, O.A., and Jefimova, N.A. (1978) Forthcoming climate changes. *USSR AS Bul. Sect. Geography* 6, 5–20.

Budyko, M.I., Israel, Y.A., Krecken, M.S., and Khek, A.D., eds. (1991) *Forecasting Climate Changes.* Gidrometeoisdat Leningrad.

Bredehoeft *et al.*, eds. (1982) *Groundwater Models,* Vol. 1. The UNESCO Press, Paris.

Gavitch, I.K. (1980) *Mathematical Modeling in Hydrogeology: Theory and Practice.* Nedra, Moscow.

Kovalevsky, V.S. (1974) *Basis of Forecasting Natural Groundwater Regime.* Stoyizdat, Moscow.

——. (1994) *Impact of Hydrogeological Condition Changes on the Environment.* Nauka, Moscow.

Kovalevsky, V.S. and Maximova, N.G. (1988) On possible changes of river groundwater recharge in future due to climate changes. *Proceedings of Hydrogeological Congress 'Water Resources and Water Balance'*, Gidrometeoisdat, Leningrad, pp. 354–360.

Kovalevsky, V.S. and Yefremenko, A.V. (1995) Peculiarities of karst water resources in the Russian Platform and forecasting their changes due to human-induced climate transformations. *J. of Water Resources* 22(3), 372–383.

Chapter 14

Pollen Records of Past Climate Changes in West Africa since the Last Glacial Maximum

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14.1. Introduction

Environmental reconstruction since the Last Glacial Maximum in Northwestern Tropical Africa has been severely limited by the incompleteness of the record, owing primarily to major discontinuities in lacustrine sediments preserved in a predominantly arid climate. Only two long and complete pollen sequences, each from deep crater lakes, have been obtained to register the whole deglacial and Holocene history of the humid equatorial forest (Maley, 1989). Other continental sites, all closely subject to local hydrogeological conditions (Lézine and Casanova, 1989) provide detailed 'windows' on short periods of the Holocene. However, comparison with marine sedimentary sequences that provide continuous, well dated records over longer time spans allows one to reconstruct changes in past atmospheric patterns over West Africa and in vegetation distribution on sub-continental as well as local scales. I present here a review of pollen data from West Africa and the nearby Eastern Atlantic undertaken to understand the vegetation response to past climatic and hydrological changes on different time scales. Interpretations of pollen diagrams are based on numerous studies on both modern pollen deposition from the Equatorial evergreen and semi-deciduous forests to the south (Brenac, 1988; Elenga, 1992; Reynaud-Farrera, 1995) to the Sudanian, Sahelian and Saharan driest ecosystems to the north (Lézine and Edorh, 1991; Lézine and Hooghiemstra, 1990; Maley, 1972, 1981; Ritchie, 1987) (Fig. 14.1). Land-sea correlations are also made.

14.2. Present Environmental Setting

According to Leroux (1993; Fig. 14.1), the atmospheric circulation over West Tropical Africa is much influenced by air masses of polar origin: The Mobile Polar Highs

A. S. Issar and N. Brown (eds.), Water, Environment and Society in Times of Climatic Change, 295–317.

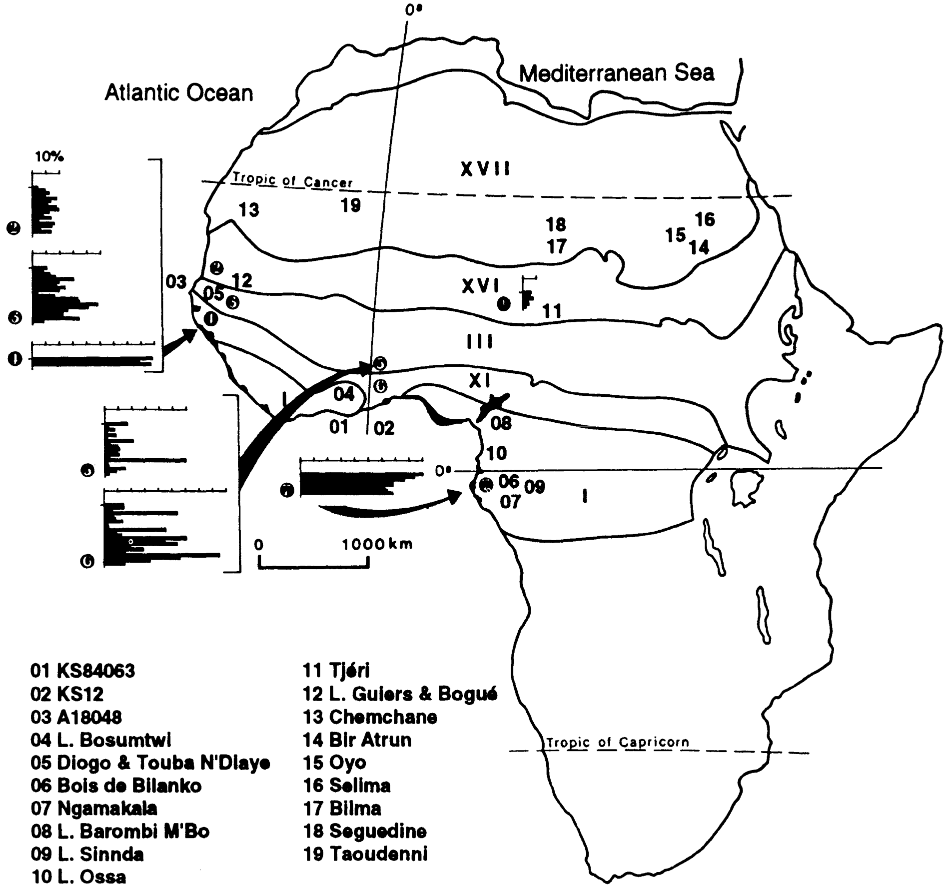


Figure 14.1. Vegetation map of West Africa after White (1983) (see text for details). The modern pollen deposition in the different vegetation zones is illustrated here by the percentages of Arboreal Pollen (calculated against the total pollen grains and spores counted per sample). Numbers in circle refer to studies on modern pollen deposition, whereas the other numbers refer to pollen diagrams cited in the text. Modern pollen deposition clearly reflects the physiognomy of the vegetation in spite of the specificity of the pollen production and distribution leading to extreme under-representation of many tree taxa in pollen diagrams from tropical areas (Ritchie, 1995): Arboreal Pollen percentages are more than 60% in forest (4,7)(Lézine and Hooghiemstra, 1990; Elenga, 1992). They decrease to less than 10% in Sahelian grasslands (1,2) (Maley, 1981; Lézine and Hoghiemstra, 1990). Guineo-Congolian secondary forest and savannas (6)(Lézine and Edorh, 1993) are characterized by high amplitude variations in Arboreal Pollen percentages that reflects locally open or degraded environment. Sudanian dry forests are characterized by Arboreal Pollen percentages centered around 20% (3,5) (Lézine and Edorh, 1993).

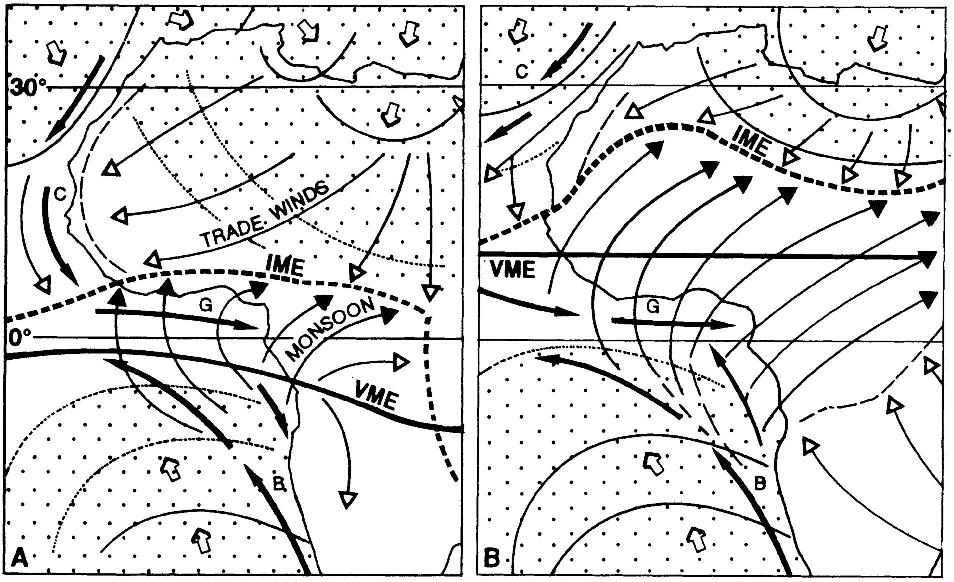


Figure 14.2. Modern features for atmospheric circulation over West Africa in January (A) and in July (B) (after Leroux, 1996). Dotted areas show the penetration of Mobile Polar Highs equatorwards.

feed the anticyclonic air masses which supply the low-layer of the trade wind fluxes towards the Meteorological Equator. These appear as two distinct branches:

1) the maritime trade wind of north-south orientation, which blows permanently along the West African coast, contributes to the southward advection of cool water in the Canaries Current and to local coastal upwelling (Wooster *et al.*, 1976).
2) the continental trade wind of roughly east-west direction, crossing the dry and hot Saharan desert, is mainly responsible for the pollen and dust transport to the ocean.

The Meteorological Equator itself (usually called Intertropical Convergence Zone (ITCZ)) is characterized by a complex structure: inclined in the lower layers and vertical in the middle layers. In the lower layers, the dry continental trade wind that belongs to the African Easterly Jet dynamics, overrides a humid monsoon, the rainfall from which comes in the form of westward moving lines of squalls. In the middle layers of the atmosphere, the vertical structure of the Meteorological Equator offers positive tropical conditions for pluviogenesis with regular and abundant rains. The surface position of the Meteorological Equator moves from 4°N in January to 20–25°N in July. As a result, mean annual precipitation progressively decreases from south to north inducing a zonal distribution of the vegetation (White, 1983; Fig. 14.2). Guineo-Congolian evergreen and semi-deciduous forests lie around the Gulf of Guinea in areas characterized by an average of more than 1,600 mm annual rainfall on both sides of the drier 'Dahomey Gap' (Zone I). They are surrounded to the north by a large extension of secondary forests and grasslands (Zone XI). These humid ecosystems are progressively replaced to the north by Sudanian dry forests and woodlands (1500–500 mm yr–1) (Zone III); Sahelian wooded grasslands (500–100 mm yr–1) (Zone XVI); and Saharan semi-desert and desert (less than 100 mm yr–1) (Zone XVII). In addition, Guinean riverine forests extend near the littoral in azonal positions in the 'Niayes' region of Senegal, up to 16°N latitude (Trochain, 1940).

14.3. Evidence of Past Atmospheric Pattern: Concentration and Influx Values in Marine Pollen Swquences (Fig. 14.3)

Long pollen sequences recovered off West Africa record important changes in strength in the two main air circulations, the continental trade winds and the Atlantic monsoon, prevailing through the last glacial/interglacial cycle. A general increase of trade wind intensity has been demonstrated from 60 kyrs BP with a noticeable acceleration since 40 kyrs (Hooghiemstra, 1989; Lézine and Casanova, 1991) to a maximum reached during the Last Glacial Maximum.

Cores KS84063 (4°28'N–4°11'W; 3,005 m water depth) (Lézine *et al.*, 1994) and KS12 (3°52'N–1°56'W; 2,955 m water depth) (Lézine and Vergnaud-Grazzini, 1993) in the Gulf of Guinea and A180–48 off Senegal (15°19'N–18°06'W; 2450 m water depth) (Lézine *et al.*, 1995) provide detailed evidence for these variations during the last deglaciation and the Holocene. Continent-ocean atmospheric exchanges are deduced from the variations in pollen concentration measured per cubic centimeter of dry sediment by the use of exotic markers *(Alnus)*. These are compared with other terrestrial markers such as dust from arid ecosystems (Lézine *et al.*, 1994) and diatoms from desiccated lakes (Pokras and Mix, 1985).

14.3.1. THE GULF OF GUINEA RECORDS

Core KS84063, with an accumulation rate ranging from 6 to 10 cm per 1,000 years, provides in fine detail the deglacial sequence from *ca.* 15 kyrs to *ca.* 3 kyrs. Pollen concentration averages 1.8 $\times 10^4$ grains/cm^3. Two well-marked maxima (2 and 4 $\times 10^4$ grains/cm^3) are recorded at levels successively dated from the onset of the deglaciation (*ca.* 15 kyrs) then from the 'Younger Dryas' event (*ca.* 10.3 kyrs). A minor peak of 2 $\times 10^4$ cm^3 is then recorded around 7 kyrs. Conversely, early Holocene levels about 8.5 kyrs are characterized by minimum values of pollen concentration (5 $\times 10^3$ grains/cm^3).

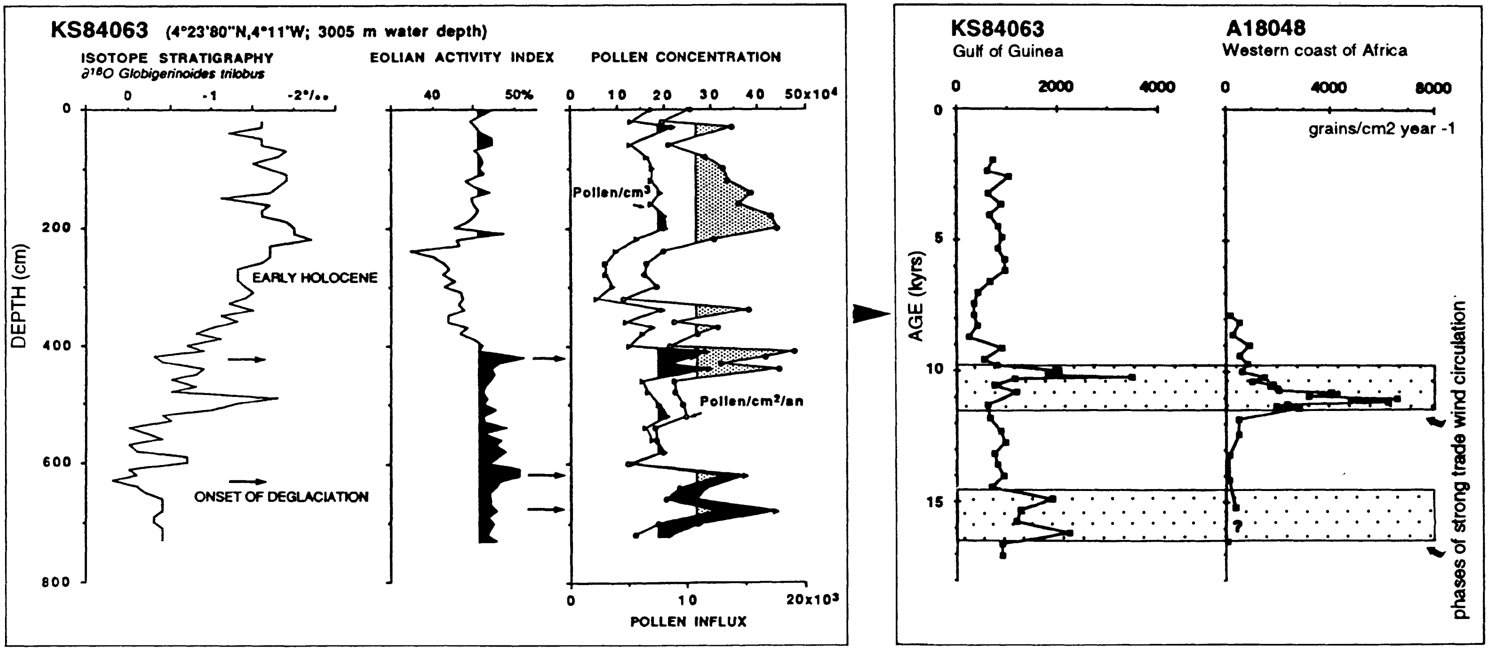


Figure 14.3. Marine pollen records of atmospheric changes over West Africa. On the left, the isotopic stratigraphy for core KS84063, and curves for dust and pollen concentration and influx (Lézine *et al.*, 1994) are plotted versus depth. Dotted/black areas show phases of important continent-ocean exchanges related to increased trade wind circulation. Continent-ocean circulation was particularly strong during the "Younger Dryas" event, near 10.3 kyrs, as shown by the comparison between the curves for KS84063 and A18048 pollen influx reported versus time (on the right) (Lézine *et al.*, 1995).

The two earliest depositional phases of increased pollen input to the ocean are characterized by the occurrence of pollen grains from the Saharan desert: *Artemisia* and *Ephedra*. These taxa are particularly abundant around 15 kyrs, as also recorded in core KS12. Their deposition as far south as 3°N into the Atlantic ocean suggest both the southward displacement or extension of their source zone together with increased continental trade wind fluxes during dry episodes. This inference is corroborated by the high amount of dust at the same levels. Conversely, these pollen grains of Saharan origin are absent from the levels dated from the Holocene, where pollen grains mainly come from the nearest forest ecosystems.

14.3.2. THE SENEGAL SECTOR

A similar trend is observed through core A180–48 recovered off Senegal where the pollen concentration averages 1.3×10^4. Two maxima (4 and 2.5×10^4 grains/ cm^3) are reached at the base of the sequence around 15 kyrs and between 380 and 280 cm around *ca.* 10.5 kyrs. The values of pollen influx (measured per cm^2 per year), which integrate the fluctuations of the accumulation rate through the sedimentary sequences, confirm the importance of the continent-ocean atmospheric exchanges during these two specific episodes of the last deglaciation, particularly during the Younger Dryas event. During this episode, pollen influx reaches up to 6.5×10^3 grains/cm^2 yr–1 in A180–48 and 3.5×10^3 in KS84063.

14.3.3. DISCUSSION

14.3.3.1. *The Glacial-interglacial Transition*

Reinforced eolian activity from a NE direction over the Eastern Atlantic at 15 kyrs then at 10.3 kyrs as recorded in the marine pollen sequences clearly corresponds to severe dryness on the nearby continent leading to the lowering of lake and watertable levels (Gasse *et al.*, 1990; Talbot and Johannessen, 1992; Lamb *et al.*, 1995). It is recorded all along the western coast of the continent (core KS78007; Marret and Turon, 1994) as far north as 37°46'N off Portugal (core SU8118; Lézine and Denèfle, 1997). In the latter core, phases of increased pollen input mainly from steppic environment are characterized by a significant lowering in sea surface temperatures as recorded by the foraminiferal assemblages in the same levels (Bard *et al.*, 1989).

These pollen data suggest that during the dry/cold phases of the last deglaciation, both middle and low latitudes came under a regime dominated by meridional atmospheric exchanges. This agrees well with a previous hypothesis (Leroux, 1993) regarding the role of reinforced anticyclonic arrays over the North Atlantic during glacial times (the Polar Highs) involving air-mass and energy transport from high latitudes to low.

14.3.3.2. *The Holocene*
The Holocene is, to the contrary, characterized by low continent-ocean exchanges, a situation expressive of the dominance of the Atlantic monsoon fluxes in tropical latitudes; and the predominant influence of the westerlies farther north, from about 37°N. Core KS84063 clearly records, however, a short episode of increased pollen input to the ocean around 7 kyrs that might indicate the restoration of dry conditions over West Africa contemporaneous with a widespread though short interval of low lake-levels described in North Africa by Servant (1983) and others.

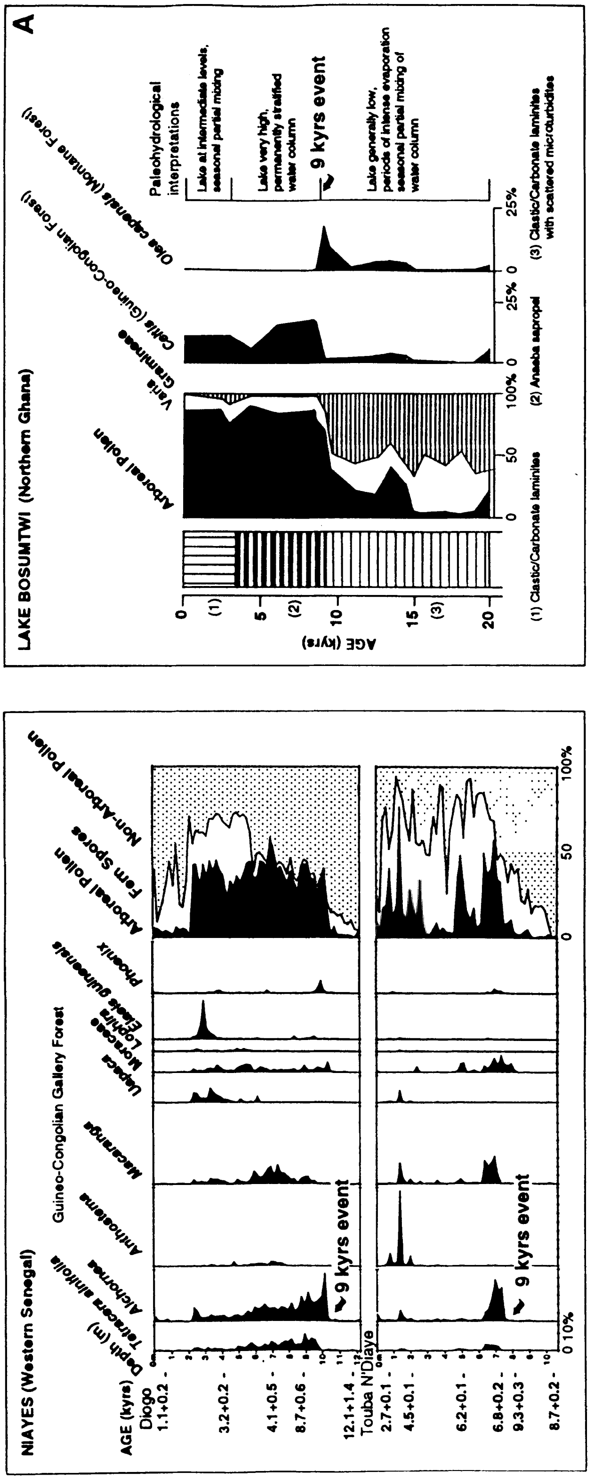
14.4. Evidence of Past Rainfall and Temperature Variations: Vegetation Changes from Continental and Marine Pollen Data

14.4.1. THE FOREST ECOSYSTEM

14.4.1.1. *The Western Domain* (Fig. 14.4a)
Pollen sequences from tropical humid and equatorial Africa west of the 'Dahomey Gap' are scarce. Only one site located at the northern edge of the forest documents the history of the western Guineo-Congolian domain from *ca.* 18 kyrs: Lake Bosumtwi in the mountains of Ghana (6°30'N, 1°25'W, 100 m alt.; Maley, 1989). More to the north, the 'Niaye' region of Senegal (15–16°N, 17°30'–16°30'W, 3–11 m alt.) provides additional elements on the northward forest expansion during the Holocene (Lézine, 1988a).

The Lake Bosumtwi record is characterized by the occurrence of montane elements such as Gramineae Pooideae (Talbot *et al.*, 1984) and *Olea* (Maley and Livingstone, 1983) in levels dated from the Last Glacial Maximum and the last Glacial-Interglacial transition up to ca. 9 kyrs. The forest cover was much reduced as recorded by the low amount of Arboreal Pollen and the vegetation was of open, cold-steppic character. During this period, the lake was generally low, rather saline and alkaline, while the general environment was dry, particularly during short events successively centered around 18.5, 14.7, 12 and 10.5 kyrs (Talbot and Johannessen, 1992). The local expansion of montane forest elements indicates that mean temperatures were about 4° lower than today (Maley, 1989). The forest rapidly developed after *ca.* 9 kyrs, and took on its modern aspect with the disappearance of the montane taxa and the development of *Celtis* and other mesophilous elements. The early Holocene period was characterized by increased humidity and by slack winds and/or minimal temperature variations which allowed a notably stable water column to persist in the lake (Talbot and Johannessen, 1992).

Diogo II and Touba N'Diaye pollen diagrams, in the Niaye region register the same abrupt forest development at 9 kyrs in western Senegal between 15 and 16°N. These two pollen diagrams allow one to depict the regional history of the vegetation at the northernmost boundary of the forest domain. That is in spite of the


A
LAKE BOSUMTWI (Northern Ghana)
Arboreal Pollen
Varia
Gramineae
Celtis (Guineo-Congolian Forest)
Olea capensis (Montane Forest)
Paleohydrological interpretations
9 kyrs event
AGE (kyrs)
(1) Clastic/Carbonate laminites
(2) Anaeba sapropel
(3) Clastic/Carbonate laminites with scattered microturbidites
NIAYES (Western Senegal)
Guineo-Congolian Gallery Forest
Depth (m)
Tetracera alnifolia
Alchornea
Anthocleista
Macaranga
Uapaca
Moraceae
Lophira
Elaeis guineensis
Phoenix
Arboreal Pollen
Fern Spores
Non-Arboreal Pollen
Diogo
Touba N'Diaye
9 kyrs event

proximity of the Atlantic ocean and its effect on local hydrology, particularly on the water-table position at the bottom of the interdunes. This is responsible for high amplitude variations in pollen composition between the sites studied.

At the base of the pollen sequences, between *ca.* 12 kyrs and 9 kyrs, pollen spectra are characterized by less than 3% of Arboreal Pollen mainly *Celtis* of Sudanian character; the other forest elements *Alchornea, Ficus* are of scattered occurrence with low frequencies overall. The pollen spectra are dominated by Gramineae. These, together with the occurrence of Sahelian and Saharan steppic taxa testify to dry regional conditions. At 9 kyrs, percentages of Arboreal Pollen taxa increase considerably (up to 57% at Diogo and 77% at Touba N'Diaye) while the taxonomic composition of the pollen spectra includes more than 150 taxa mainly from humid, Sudano-Guinean to Guinean vegetation formations *(Alchornea, Macaranga,* Moraceae, *Elaeis guineensis, Anthostema senegalense...)* recording the expansion of mesophilous gallery forests near the littoral up to 16°N along the Atlantic coast, *i.e.* more than 2 degrees north of their current limit. This expansion was coeval with extensive high lake levels throughout all West Africa (Lézine and Casanova, 1989).

As early as 7.5 kyrs, the most characteristic Guinean forest trees disappeared in relation with drier conditions. Swamp forests of Sudano-Guinean character dominated by *Syzygium* developed in the Niaye area during the middle Holocene, whereas the lowering of the water table caused the reduction of the arboreal cover (Touba N'Diaye). From about 4.5 to 2 kyrs, Guinean species *(Elaeis guineensis, Anthostema senegalense, Uapaca...)* which had disappeared after 7.5 kyrs, went through a second phase of expansion when moisture increased. At 2.5 kyrs, after the last major positive hydrological phase recorded in the Sahelo-Saharan zone (Lézine and Casanova, 1989), a rapid degradation of forests occurred and the regional vegetation assumed its modern semi-arid character.

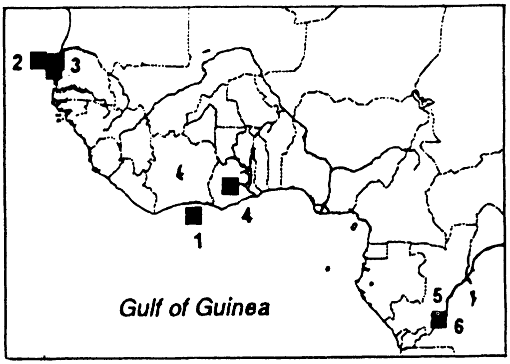
14.4.1.2. *The Equatorial Domain* (Fig. 14.4b)

The pollen record from Bois de Bilanko on the Bakete Plateau in Congo (3°31'S, 15°21'E, 600 m alt.; Elenga and Vincens, 1990) points to an expansion of montane forest elements (*Podocarpus, Olea capensis* and *Ilex mitis*) towards middle altitudes during the last deglaciation up to *ca.* 10.5 kyrs. Their appreciable percentages (*Podocarpus* = 35%; *Olea* and *Ilex* = 5 to 10%) clearly indicate that these cold, montane species developed locally in association with lowland forest elements of Guineo-Congolian affinity (*Syzygium, Macaranga, Canthium,* Combretaceae...). In the same area, the Ngamakala pond pollen sequence (4°4'30"S, 15°23'E, alt. 400 m; Elenga *et al.*, 1994) shows that the forest cover was slightly reduced during the Last Glacial Maximum: the Arboreal Pollen percentages fell to *ca.* 40% in the corresponding

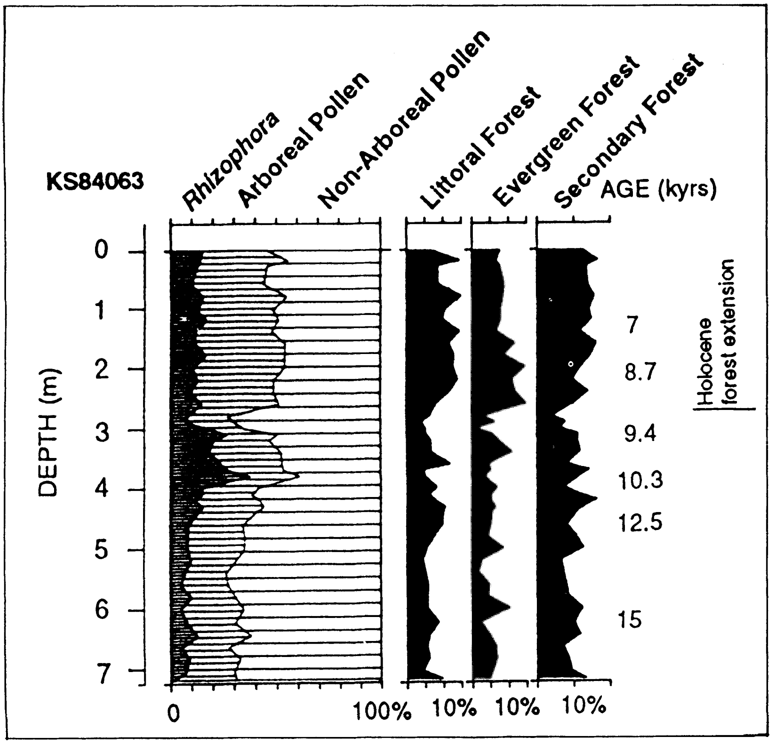
Figure 14.4a. Simplified pollen diagrams for the western Guineo-Congolian forest domain (after Lézine, 1988 for the Niaye region; Maley (1989), Talbot and Johannessen (1992) for Lake Bosumtwi). Both diagrams point to the 9 kyrs event showing the abrupt rise in tree pollen percentages.

levels, whereas they were to reach *ca.* 90% during the Holocene after *ca.* 9 kyrs. In this site however, the taxonomic composition has varied little from *ca.* 24 kyrs BP to the present.

In the Cameroon Highlands, the pollen sequence from Lake Barombi M'bo (4°40'N, 9°24'E, 350 m alt.; Maley, 1989) also records a certain decrease in the representation of Arboreal Pollen at levels dated from the Last Glacial Maximum, the



Location map: (1)KS84063, (2) A18048, (3)Niayes, (4) L. Bosumtwi, (5) Bilanko, (6)Nagamakala



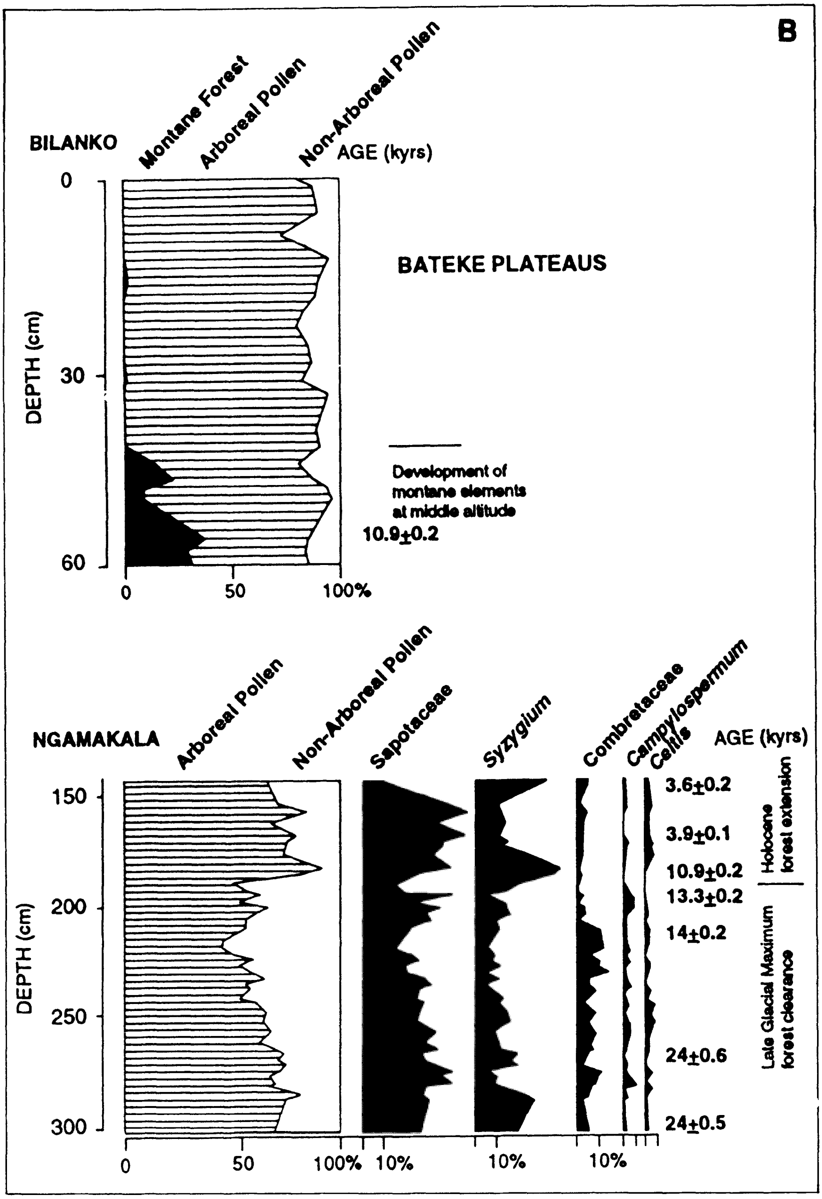


Figure 14.4b. Simplified pollen diagrams for the central Guineo-Congolian forest domain (after Lézine and Le Thomas, 1995; Elenga and Vincens, 1990; Elenga *et al.*, 1994). These diagrams shows the increased importance of rain forest species at the beginning of the Holocene. The Late Glacial Maximum levels are characterized by a slight decrease in Arboreal Pollen percentages and the local development of secondary forest taxa. Near the Equator, the floristic composition of the lowland forest probably did not change considerably, whereas the midland forest was characterized by the development of montane elements.

percentages reached being 25–30%. Then, the forest cover developed from *ca.* 9 kyrs. After which, no major changes are evident in the forest domain during Holocene times. However, the absence of detail concerning the taxonomic composition of the pollen spectra does not permit one to go deeper in discussion concerning the floristic composition of the forest throughout the period studied. A short episode of degradation of the lowland evergreen and semi-deciduous forests is recorded throughout Equatorial Africa around 3 kyrs. Also these ecosystems were locally replaced by open formations dominated by grasses in southern Congo (Lake Sinnda, 3°50'S, 12°48'E, 128 m alt.; Vincens *et al.*, 1994) or by pioneer, heliophilous species in Cameroon (Lake Ossa, 3°45'N, 9°58'E, 8 m alt.; Reynaud-Farrera *et al.*, 1996).

14.4.1.3. *The KS84063 Marine Pollen Record*

Two hundred and forty-two pollen taxa have been determined through core KS84063 allowing to reconstruct the past vegetation history on the nearby continent since 15 kyrs (Lézine and Le Thomas, 1995). This provides one with a regional view of vegetation modifications out of the direct influence of local hydrological or geomorphological settings. The pollen sequence establish first and foremost the permanency of the arboreal cover along the Gulf of Guinea during the whole period studied, as demonstrated by the low amplitude of Arboreal Pollen variation throughout the core, no more than 20%. Between 15 kyrs and 9 kyrs, the domination of heliophilous taxa (belonging mostly to the family Euphorbiaceae) suggests an opening up of the Guineo-Congolian forest in response to regional dryness. The rain forest (*Saccoglotis*, Moraceae, *Parkia, Lophira, Irvingia*...) developed widely between *ca.* 9 kyrs and 8 kyrs. Then, during the middle- to late Holocene period, strong modifications occurred in the forest composition probably due to decreasing rainfall. More open rain and secondary forests coexisted in a swampy environment.

14.4.2. THE SAHELIAN DRY ECOSYSTEMS (Lézine, 1989)

The Tjéri section (Fig. 14.5) together with additional outcrops in the Lake Chad depression (13°44'N, 16°30'E; Maley, 1981) provides information on vegetation changes in the Lake Chad watershed since 13 kyrs. Lake Chad lies in a nowadays dry, Sahelian steppic environment. But its drainage basin covers some 2.5 million km^2, including areas of Sudanian and Sudano-Guinean vegetations to the South and the foothills of the Tibesti massif to the northeast. According to Maley (1981), the regional vegetation of the lake was of grassland semi-desert type *ca.* 12.5 kyrs while wind transport from northern areas (the Tibesti mountain, the Saharan and the Mediterranean zones) was important. Then after 9.2 kyrs, fluctuations through the Tjéri sequence were essentially between three eco-groups of taxa: Sahelian, Sudanian and Sudano-Guinean. Higher percentages of local Sahelian taxa are recorded during phases of higher lake level at 9–8.5, 6.4–4.45 and 3.5–3 kyrs (Servant-Vildary, 1977). The second of these lacustrine phases corresponded to the maximum extent of the paleolake; and may have occurred during a phase of extension of the Sahelian vegetation zone toward the north (7.5–5.1 kyrs). Maley (1981) suggests that the main part of the Sudanian and Sudano-Guinean pollen grains of southern origin transported to the lake then was linked to increased fluvial transport, in relation to local rainfall in the corresponding vegetation zones, and so did not reflect significant modification in the vegetational environment of Lake Chad. The peak of Sudano-Guinean taxa around 7.1 kyrs is however interpreted as reflecting a certain advance of Sudanian elements toward the north. The fluviatile

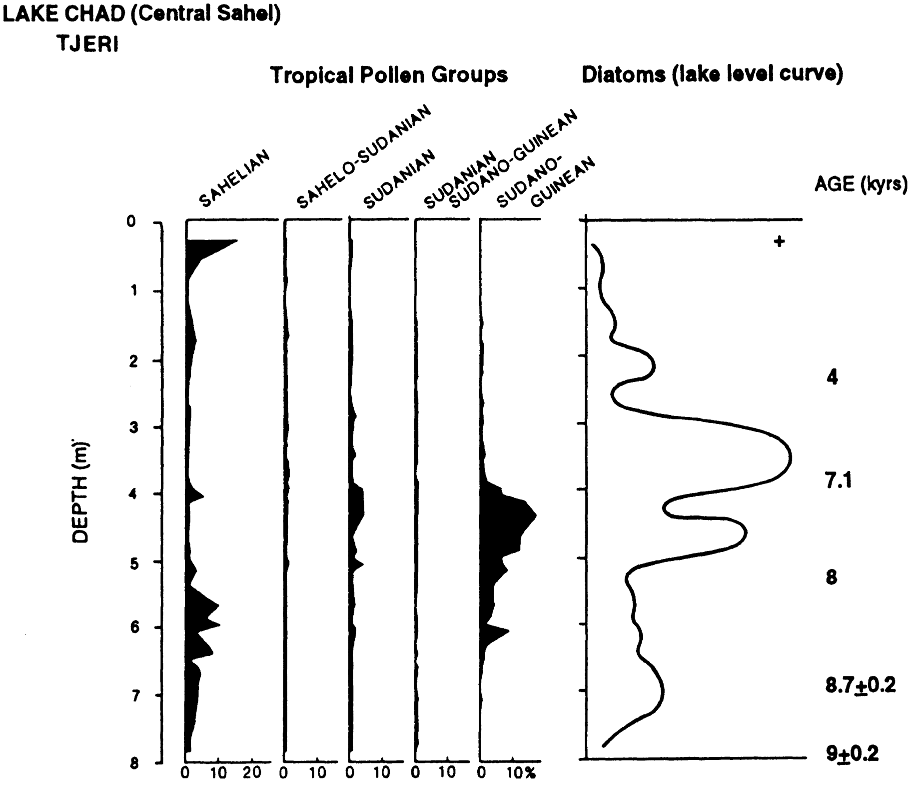


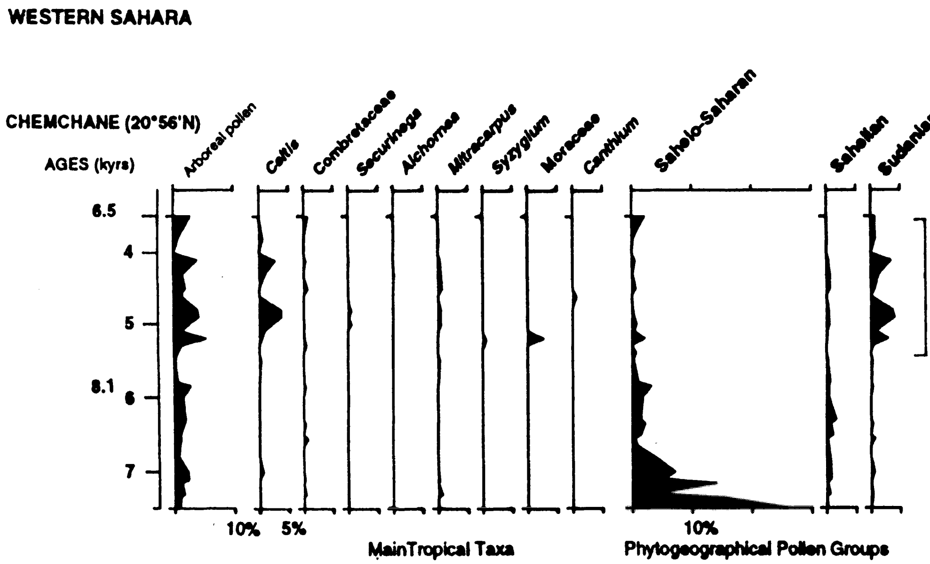
Figure 14.5. Simplified pollen diagram for Lake Chad (central Sahel) (from Maley, 1981). Pollen data are compared to diatom ones interpreted in terms of lake level fluctuations (after Servant-Vildary, 1977).

contribution from the Sudanian and Sudano-Guinean vegetation zones to the lake decreased steadily after 4 kyrs and ceased altogether about 500 AD.

In Lake Guiers (16°07'N, 15°55'W, 1 m alt.; Lézine, 1988b) and along the Senegal River banks at Bogué (16°35'N, 14°17'W; Michel and Assémien, 1970), pollen data confirm the increase in elements of southern origin by fluvial transport, related both to a northern extension of the corresponding vegetation up to the shores of the lake and along the Senegal River as gallery forests as well as to increased fluvial activity, from the base of the sequences studied (*ca.* 6 kyrs) to *ca.* 2 kyrs. About 2 kyrs, the vegetation took on its modern aspect and the fluvial inputs strongly decreased or ceased (Monteillet *et al.*, 1981; Chamley and Diester-Hass, 1982).

14.4.3. THE SAHARAN DESERT (Fig. 14.6)

Pollen diagrams from the Tropical Sahara document climatic phases favorable to lake extensions, notably during the early to middle Holocene. The strong eolian deflation during dry episodes (and particularly after *ca.* 4 kyrs) makes it impossible to record the complete history of past environments in this area. At Chemchane in Mauritania (20°56'N, 12°13'W; Lézine *et al.*, 1990; Lézine, 1993), the pollen diagram shows the progressive rise of fresh waters in the depression from about 11–12 kyrs: a halophitic, sebkha-type vegetation mainly composed of Chenopodiaceae-Amaranthaceae was progressively replaced by aquatics (Cyperaceae and *Typha*), the maximum extension of which corresponded to an important lacustrine epi-



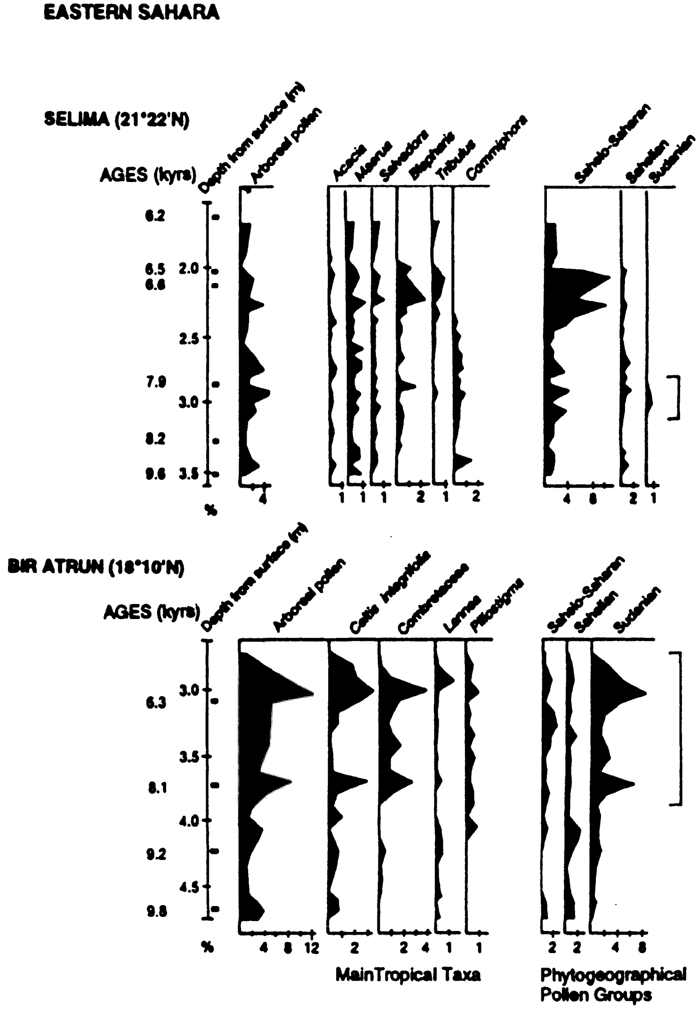


Figure 14.6. Simplified pollen diagrams for the modern Sahara. The diagrams show curves for Arboreal Pollen, the main tropical taxa and the phytogeographical groups of tropical taxa reported versus depth. The latter groups include Arboreal and Non-Arboreal taxa. From West to East, the same phytogeographical pattern is recorded, characterized by the development of Sudanian taxa indicated by] on the diagrams.

sode dated between *ca.* 8.3 and 6.5 kyrs from stromatolites surrounding the lake. During this interval, the regional vegetation of Chemchane was Sudano-Sahelian in character, dominated by a continuous graminoid cover with plants such as *Celtis integrifolia*-type, *Lannea, Rhus, Securinega virosa, Alchornea*... currently found 400 to 500 km to the South.

Similar vegetation changes during the early Holocene lacustrine episode are also to be observed in the eastern Sudan from pollen analyses at three sites located in regions that today are hyperarid deserts: Bir Atrun (18°10'N, 26°39'E; Ritchie, 1984; Ritchie and Haynes, 1987), Oyo (19°16'N, 26°11'E; Ritchie *et al.*, 1985; Ritchie, 1994) and Selima (21°22.2'N, 29°18.6'E; Ritchie and Haynes, 1987; Haynes *et al.*, 1989). These yielded continuous sequences for the period *ca.* 9.5 to 4.5 kyrs. At Bir Atrun and Oyo, Sudanian *(Celtis, Combretaceae, Lannea, Piliostigma...)* and Sahelian elements *(Balanites, Acacia, Commiphora, Grewia...)* were particularly well developed up to *ca.* 6 kyrs, suggesting an ambience of wooded savanna similar to that found today in the Ennedi, Darfur and throughout the Sahelo-Sudanian vegetation zone of Sudan. After 6 kyrs, the increased importance of Saharan taxa might indicate a partial replacement of savanna by semi-desert scrub due to climatic deterioration. At Selima, the Sudanian elements were, within the tropical group of taxa, much less important. Their maximum values, never in excess of one percent, are reached during a humid period centered *ca.* 8 kyrs.

The presence of Sahelian taxa (such as *Balanites* and *Mitracarpus*) and of more humid types (such as *Celtis, Securinega* and even *Hymenocardia* and *Elaeis*) around 8.6 kyrs at Bilma, Niger (18°41'N, 12°55'E, Schulz *et al.*, 1990), between 8.2 and 6.5 kyrs at Seguedine, Niger (20°20'N, 12°50'E) and before 5.65 kyrs at Taoudenni, Mali (22°30'N, 4°W; Schulz, 1987) accords with the preceding diagrams. These pollen data demonstrate a general extension of wooded formations of Sudanian character as far as the Tropic of Cancer in the western Sahara (Chemchane, Taoudenni) and 19°N in the eastern Sahara (Oyo, Selima) during the early Holocene. Similarly, desert grassland and scrubland communities probably extended up to 25°N in the central Sahara (Schulz, 1987). Pollen data from the southern fringe of the modern Sahara are in agreement with other paleobotanical investigations such as plant macroremains studies carried out in central (Fachi, Niger; 18°18'N, 11°23'E) and eastern Sahara (Wadi Shaw; *ca.* 20°30'N) (Neumann, 1991) showing the presence in well-dated charcoal layers of numerous Sudanian tree species. The inference is that, during the early Holocene humid maximum the absolute desert was probably eliminated (in Street-Perrott *et al.*, 1990).

14.4.3. ANALYSIS

14.4.3.1. *The Question of the Last Glacial Maximum Forest Refugia* (Fig. 14.7)

During the Last Glacial Maximum, the northern boundary of the Equatorial forest domain probably retreated southward as indicated in the Bosumtwi pollen diagram. This corroborates previous botanical analyses according to which the lowland rain and semi-deciduous forest formations of tropical Africa have experienced large variations in geographical extent, with marked shrinkage during climatically adverse periods (Aubreville, 1949). However, other pollen data from Equatorial Africa do not reveal any such alternation. Arboreal Pollen percentages are maintained throughout the last glacial/deglacial, while floral composition of spectra regarding the lowland evergreen and semi-deciduous forests do not change considerably.

So these data do not provide, up to now, clear evidence for an opening of the forest or the survival of forest species only in restricted areas, the 'forest refugia', during the last glacial as deduced *e.g.* by Maley (1989) from botanical and zoological studies. However, more pollen diagrams are required to give a satisfactory response to this question. According to Ritchie (1995), our current knowledge of the vegetation response to climate changes in Equatorial Africa and particularly of for-

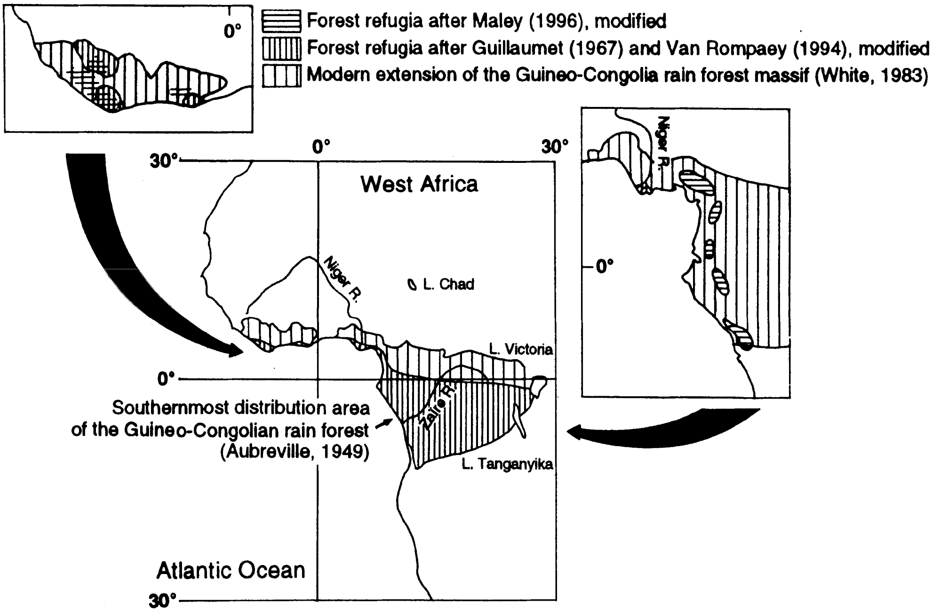


Figure 14.7. Map showing the southernmost migration of the Guineo-Congolian forest during glacial times after Aubreville (1949). Details are shown for the location proposed for Late Glacial Maximum 'forest refugia' in the western and central forest domain from botanical and zoological studies of species endemism. Up to now, the 'forest refugia' theory is not confirmed by the available pollen data which are still too scarce. More detailed investigations, based on high resolution pollen sequences, are needed to explain the origin of endemism variations in the forest.

est dynamics is severely limited by the scarcity of available data. Additional difficulties come from the specific character of pollen in local ecosystems; the large number of tree species overall; the low pollen productivity of species characterized by predominantly zoophilous pollinating systems; and the small number of pollen grains that can be determined at a specific level. All of this tends to lead to extreme under-representation or exclusion of many tree taxa in pollen diagrams.

Recent studies carried out in the Amazon forest massif by Colinvaux *et al.* (1996) underline the above considerations and indicate a need to reconsider the theory of 'glacial forest refugia' in the Equatorial regions. Using pollen data from Lake Pata, characterized by a large amount of arboreal pollen from lowland rain forest through the whole glacial/deglacial time (from 42 kyrs to the Holocene), the authors conclude that the reduction in precipitation was not sufficient to fragment the rain forest. They add that 'alternatives to the refuge hypothesis are required to explain species endemism... The primary environmental forcing of the Amazon system in glacial time was a drop in temperature, a direct consequence of which was an increase in diversity as rain forest communities accommodated more cool-adapted taxa as well as their present array'.

14.4.3.2. *The Temperature Changes at the Glacial-Deglacial Transition*

The present-day distribution of *Podocarpus* populations is restricted to the highlands of Cameroon, Nigeria and Sao Tomé between 1200 and 2500m in altitude (White, 1983). The extension of this element together with other montane species (such as *Ilex mitis, Olea capensis...*) toward middle altitudes in the Equatorial forest domain before the Holocene has been interpreted as response to a drop in mean annual temperature by 4 to 6°C compared to modern values, applying a mean temperature gradient of 0.6°C/100 m to the vegetation margins (Maley and Elenga, 1993; Elenga *et al.*, 1994). In addition, the regional conditions were probably not as dry as previously suggested, with a cloud cover being similar to that nowadays found at higher altitudes. However, the entry of cold, montane species into the Equatorial forest domain of Africa did not reach altitudes lower than 700 m, these taxa being absent in the surroundings of the Ngamakala pond.

The 4 to 6°C Late Glacial Maximum cooling estimated in the West African Equatorial midlands agree well with that recorded elsewhere in the tropics (*e.g.* Colinvaux *et al.*, 1996). It confirms the sensitivity of the tropical mid- and lowlands to the temperature changes that affected the Earth during the late Pleistocene glaciation (Broecker, 1995).

14.4.3.3. *The Rainfall Variations*

The lack of quantitative paleoclimatic parameters based on modern analogs such as those applied to East Africa (Bonnefille *et al.*, 1990; Vincens *et al.*, 1993) make it difficult to reconstruct precisely rainfall variations during the last deglaciation in West Tropical and Equatorial African lowlands. At present, the only paleo-precipitation reconstructions attempted from pollen records have been from the southern margin of the Saharan desert. There comparison between fossil pollen assemblages and modern ones and related plant distribution has been used to infer that the 400 mm isohyet was displaced north of its present-day position to at least 19°N in the Eastern Sudan and 21°N in Mauretania during the Early Holocene. After 6 kyrs, a certain climatic deterioration was recorded and the mean annual rainfall probably decreased from 400 to 300 mm at Oyo, Sudan. Finally, intensified aridification after 4.5 kyrs produced the modern desert vegetation associated, on this site, with a mean annual rainfall of only 5 mm.

Southwards, in the modern Sahel, comparison between pollen and limnological data (Lézine and Casanova, 1989) allows one to propose the following chronology of paleohydrological events:

The development of humid ecosystems (such as rain forests near the Gulf of Guinea; mesophilous forest in the Niaye area; Sudanian savannas in the desert) between 9 and 8 kyrs BP was coeval with wide lake extensions. This was likewise contemporaneous with a maximum intensity of Atlantic monsoon fluxes and related greater rainfall over North Africa (COHMAP, 1988). New quantitative estimates for this, taking into account the surface albedo changes in the inland delta of the Niger (Street-Perrott *et al.*, 1990), have proposed an increase in precipitation of 150–320 mm pa, *i.e.* slightly lower than those derived from pollen data in the Southern Sahara. Changes in the composition of the vegetation (with increased importance of Sudanian taxa) and changes in paleohydrological features (which take on a mosaic-like character) make it possible to date the beginning of well-marked seasonality in precipitation *ca.* 7.5 kyrs.

The middle and late Holocene were then characterized by a minor phase of increased rainfall responsible for the reestablishment of Guinean species near the littoral and high water-table levels between 4.5 and 2 kyrs. After this date, only minor fluctuations of lake levels are recorded (Maley, 1981). Related rainfall variations were not sufficient to modify the regional vegetation of the Sahel that therefore assumed its modern aspect at 2 kyrs.

The Equatorial rain forest ecosystem seems to have been remarkably stable during the middle and late Holocene. However, the short event of forest clearance recorded at 3 kyrs has been related to a relative dryness. The vegetation response in this case was probably amplified by human activity.

14.5. Concluding Remarks

The review of pollen data from Tropical West Africa presented here illustrates the complementarity of land and marine investigations in the interpretation of paleoclimate change. The 'Younger Dryas' event is a good example for this approach. Marine sediments record an abrupt increase of wind-transported pollen from dry ecosystems at this time all along the West African coast and even northward off Portugal, thus reflecting a strong anticyclonic circulation. However, only one continental pollen diagram might have recorded the effect of the related intense dryness at this time on the tropical ecosystems. At Bosumtwi, the first post-glacial phase of forest advance recorded around 13 kyrs was suddenly interrupted between 12.5 and 9.5 kyrs, probably in relation to dry conditions.

In addition, this review points to the need of additional continental pollen studies (1) to ascertain more exactly the vegetation dynamics within the forested areas (*e.g.* to discuss the theory of glacial forest refugia) as well as in the Sahel and the Saharan desert (*e.g.* to study the migration rates of humid ecosystems to the north during post-glacial times and the low amplitude variations of the vegetation cover through the Holocene) and (2) to go deeper in the data-model comparisons for simulations using the surface conditions such as albedo, soil moisture or roughness. Whether present-day analogs can be found for the last glacial ecosystems in the Tropics remains problematic.

Acknowledgements

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References

Aubréville, A. (1949) *Contribution à la paléohistoire des forêts de l'Afrique tropicale.* Société d'Edition de Géographie maritime et coloniale, Paris.

Bard, E., Fairbanks, R., Arnold, M., Maurice, P., Duprat, J., Moyes, J., and Duplessy, J.C. (1989) Sea-level estimates during the last deglaciation based on d^{18}O and accelerator mass spectrometry ^{14}C ages measured in *Globigerina bulloides. Quaternary Research* 31, 381–391.

Bernac, P. (1988) Evolution de la végétation et du climat dans l'ouest Cameroun entre 25,000 et 11,000 ans BP. *Institut Français de Pondichéry, Travaux de la Section Scientifique et Technique* 25, 91–103.

Bonnefille, R., Roeland, J.C. and Guiot, J. (1990) Temperature and rainfall estimates for the past 40,000 years in equatorial Africa. *Nature* 346, 347–349.

Broecker, W.S. (1995) Cooling the tropics. *Nature* 376, 212–213.

Chamley, H., Diester-Hass, L. (1982) Effets du déplacement de l'Embouchure du fleuve Sénégal au Quaternaire supérieur sur la sédimentation de la marge ouest-africaine. *Comptes Rendus de l'Académie des Sciences,* Paris, 2, 295, 673–678.

COHMAP (1988) Climatic changes of the past 18,000 years: observations and model simulations. *Science* 241, 1043–1052.

Colinvaux, P.A., De Oliveira, P.E., Moreno, J.E., Miller, M.C., and Bush, M.B. (1996) A long pollen record from lowland Amazonia: Forest and cooling in glacial times. *Science* **274**, 85–88.

Elenga, H. (1992) Végétation et Climat du Congo depuis 24,000 ans BP. Analyse palynologique de séquences sédimentaires du pays Bakété et du littoral. Thesis, University Aix-Marseille.

Elenga, H., Schwartz, D. and Vincens, A. (1994) Pollen evidence of late Quaternary vegetation and inferred climate changes in Congo. *Palaeogeography, Palaeoclimatology, Palaeoecology* **109**, 345–356.

Elenga, H. and Vincens, A. (1990) Paléoenvironnements quaternaires récents des plateaux Batéké (Congo): Étude palynologique des dépôts de la dépression du bois de Bilanko. *In* Lanfranchi R., and Schwartz, D. (eds.), *Paysages Quaternaires de l'Afrique Centrale Atlantique.* ORSTOM, Paris, pp. 271–281.

Gasse, F., Téhet, R., Durand, A., Gibert, E., and Fontes, J.C. (1990) The arid-humid transition in the Sahara and the Sahel during the last deglaciation. *Nature* **346**, 141–146.

Haynes, C.V., Eyles, C.H., Pavlish, L.A., Ritchie, J.C., and Rybak, M. (1989) Holocene palaeoecology of the eastern Sahara: Selima oasis. *Quaternary Science Reviews* **8**, 86–99.

Hooghiemstra, H. (1989) Variations of the NW African Trade Wind regime during the last 140,000 years: Changes in pollen flux evidenced by marine sediment records. *In* M. Leinen and M. Sarnthein (eds.), *Paleoclimatology and Paleometeorology: Modern and Past Patterns of Global Atmospheric Transport,* Kluwer Academic Publishers, Dordrecht, pp. 733–770.

Lamb, H.F., Gasse, F., Benkaddour, A., El Hamouti, N., Van Der Kaars, S., Perkins, W.T., Pearce, N.J., and Roberts, C.N. (1995) Relation between century-scale Holocene arid intervals in tropical and temperate zones. *Nature* **373**, 134–137.

Leroux, M. (1993) The mobil Polar High: A new concept which explains the actual mechanisms of the meridian airmass and energy exchanges, and the global propagation of palaeoclimatic changes. *Global and Planetary Changes* **7**, 69–93.

Lézine, A.-M. (1988a) Les variations de la couverture forestière mésophile d'Afrique occidentale au cours de l'Holocène. *Comptes rendus de l'Académie des Sciences,* Paris, **307**, 2, 439–445.

——. (1988b) New pollen data from the Sahel (Lake Guiers, Senegal). *Review of Palaeobotany and Palynology* **55**, 141–154.

——. (1989) Late Quaternary vegetation and climate of the Sahel. *Quaternary Research* **32**, 317–334.

——. (1993) Chemchane, histoire d'une sebkha. *Sécheresse* **4**, 25–30

Lézine, A.M. and Casanova, J. (1989) Pollen and hydrological evidence for the interpretation of past climates in Tropical West africa during the Holocene. *Quaternary Science Reviews* **8**, 45–55.

——. (1991) Correlated oceanic and continental records demonstrate the past climate and hydrology of North Africa (0–140 kyr). *Geology* **19**, 307–310.

Lézine, A.M., Casanova, J. and Hillaire-Marcel, C. (1990) Across an early Holocene humid phase in western Sahara, pollen and isotope stratigraphy. *Geology* **18**, 264–267.

Lézine, A.-M. and Denefle, M. (1997) Enhanced anticyclonic circulation in the eastern north Atlantic during cold intervals of the last deglaciation inferred from deep-sea pollen records. *Geology* **25**(2), 119–122.

Lézine, A.M. and Edorh, T. (1991) Modern pollen deposition in West African Sudanian environments. *Review of Palaeobotany and Palynology* **67**, 41–58.

Lézine, A.M. and Hooghiemstra, H. (1990) Land-Sea comparisons during the last glacial-interglacial transition: pollen records from West Tropical Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* **79**, 313–331.

Lézine, A.M. and Le Thomas, A. (1995) Histoire du massif forestier ivoirien au cours de la dernière déglaciation. *In* 2° *Symposium de Palynologie Africaine,* Tervuren (Belgique) *Publ. Occas. CIFEG,* Orléans 5, 31, 73–85.

Lézine, A.M., Tastet, J.P. and Leroux, M. (1994) Evidence of atmospheric paleocirculation over the Gulf of Guinea since the Last Glacial Maximum. *Quaternary Research* **41**, 390–395.

Lézine, A.M., Turon, J.L. and Buchet, G. (1995) Pollen analyses off Senegal: Evolution of the coastal palaeoenvironment during the last deglaciation. *Journal of Quaternary Science* 10(2), 95–105.

Lézine, A.M. and Vergnaud-Grazzini, C. (1993) Evidence of forest extension in West Africa since 22,000 BP: A pollen record from the Eastern Tropical Atlantic. *Quaternary Science Reviews* 12, 203–210.

Maley, J. (1989) Late Quaternary climatic changes in the African rain forest: forest refugia and the major role of sea surface temperature variations. *In* M. Leinen and M. Sarnthein (eds.), *Paleoclimatology and Paleometeorology: modern and past patterns of global atmospheric transport*, Kluwer Academic Publishers, Dordrecht, pp. 585–616.

——. (1972) La sédimentation pollinique actuelle dans la zone du lac Tchad (Afrique centrale). *Pollen et Spores* 14(3), 263–307.

——. (1981) Etudes palynologiques dans le bassin du Tchad et paléoclimatologie de l'Afrique nord-tropicale de 30,000 ans à l'époque actuelle. ORSTOM, Paris, *Travaux et Documents* 129.

Maley, J. and Elenga, H. (1993) Le rôle des nuages dans l'évolution des paléoenvironnements montagnards de l'Afrique tropicale. *Veille climatique satellitaire* 46, 51–63.

Maley, J. and Livingstone, D.A. (1983) Extension d'un élément montagnard dans le sud du Ghana (Afrique de l'ouest) au Pléistocène supérieur et à l'Holocène inférieur: premières données polliniques. *Comptes rendus de l'Académie des Sciences,* Paris, 2, 296, 1287–1292.

Marret, F. and Turon, J.L. (1994) Pollen and dinocyst results for the last 18,000 years off North West Africa: a sea-continent correlation. *Marine Geology* 118, 107–117.

Michel, P. and Assemien, P. (1970) Etudes sédimentologique et palynologique des sondages de Bogié (basse vallée du Sénégal) et leur interprétation morphoclimatique. *Revue de Géomorphologie Dynamique* 19(3), 97–113.

Monteillet, J., Faure, H., Pirazzoli, P.A., and Ravise, A. (1981) L'invasion marine du Ferlo (Sénégal) à l'Holocène supérieur (1,900 BP). *Paleoecology of Africa* 13, 205–215.

Neumann, K. (1991) In search for the green Sahara: Palynology and botanical macro-remains. *Paleoecology of Africa* 22, 203–212.

Pokras, E.M. and Mix, A.C. (1985) Eolian evidences for spatial variability of Late Quaternary climates in Tropical Africa. *Quaternary Research* 24, 137–149.

Reynaud-Farrera, I. (1995) Histoire des paléoenvironnements forestiers du sud-Cameroun à partir d'analyses palynologiques et statistiques de dépôts holocènes et actuels. Thesis, University Montpellier.

Reynaud-Farrera, I., Maley, J. and Wirrmann, D. (1996) Végétation et climat dans les forêts du Sud-Ouest Cameroun depuis 4,770 ans BP: analyse pollinique des sédiments du lac Ossa. *Comptes rendus de l'Académie des Sciences,* Paris 322, IIa, 749–755.

Ritchie, J.C. (1984) A Holocene pollen record from Bir Atrun, Northwest Sudan. *Pollen et Spores* 24, 4, 391–410.

——. (1987) Modern pollen spectra from Dakhley oasis, Western Egyptian desert. *Grana* 25, 177–182.

——. (1994) Holocene pollen spectra from Oyo, northwestern Sudan: Problems of interpretation in a hyperarid environment. *The Holocene* 4, 1, 9–15.

——. (1995) Tansley Review n°83: Current trends in studies of long-term plant community dynamics. *New Phytologist* 130, 469–494.

Ritchie, J.C., Eyles, C.H., and Haynes, C.V. (1985) Sediment and pollen evidence for an early to mid-Holocene humid period in the eastern Sahara. *Nature* 314, 352–355.

Ritchie, J.C. and Haynes, C.V. (1987) Holocene vegetation zonation in the eastern Sahara. *Nature* 330, 645–647

Schulz, E. (1987) Die Holozäne vegetation der zentralen Sahara (N-Mali, N-Niger, SW Libyen). *Palaeoecology of Africa* 18, 143–159.

Schulz, E., Joseph, A., Baumhauer, R., Schultze, E., and Sponholz, B. (1990) Upper Pleistocene and Holocene history of the Bilma region (Kawar, NE Niger). *Proceedings of the 15th Colloquium of African Geology*, 281–284.

Servant, M. (1983) Séquences continentales et variations climatiques: Évolution du bassin tchadien au Cénozoïque supérieur. ORSTOM, Paris, *Travaux et Documents* **159**.

Servant-Vildary, S. (1977) Étude des diatomées et paléolimnologie du bassin tchadien au Cénozoïque supérieur. ORSTOM, Paris, *Travaux et Documents* **84**.

Street-Perrott, F.A., Mitchell, J.F.B., Marchand, D.S., and Brunner, J.S. (1990) Milankovitch and albedo forcing of the tropical monsoons: a comparison of geological evidence and numerical simulations for 9,000 y BP. *Transactions of the Royal Society of Edinburg: Earth Sciences* **81**, 407–427.

Talbot, M.R. and Johannessen, T. (1992) A high resolution paleoclimatic record for the last 27,500 years in Tropical West Africa from the carbon and nitrogen composition of lacustrine organic matter. *Earth and Planetary Science Letters* **110**, 23–37.

Talbot, M.R., Livingstone, D.A., Palmer, P.G., Maley, J., Melack, J.M., Delibrias, G., and Gulliksen, S. (1984) Preliminary results from sediment core from Lake Bosumtwi, Ghana. *Palaeoecology of Africa* 16, 173–192.

Trochain, J. (1940) *Contribution à l'étude la végétation du Sénégal.* Larose, Paris.

Vincens, A., Buchet, G., Elenga, H., Fournier, M., Martin, L., De Namur, C., Schwarz, D., Servant, M., and Wirmann, D. (1994) Changement majeur de la végétation du lac Sinnda (vallée du Niari, sud-Congo) consécutif à l'assèchement climatique holocène supérieur: apport de la palynologie. *Comptes rendus de l'Académie des Sciences,* Paris 318, II, 1521–1526.

Vincens, A., Chalie, F., Bonnefille, R., Guiot, J., and Tiercelin, J.J. (1993) Pollen derived rainfall and temperature estimates from Lake Tanganyika and their implication for late Pleistocene water levels. *Quaternary Research* **4**, 343–350.

White, F. (1983) *The Vegetation of Africa.* UNESCO, Paris.

Wooster, W.S., Bakun, A., Mclain, D.R. (1976) The seasonal upwelling cycle along the eastern boundary of the North Atlantic. *Journal of Marine Research* **34**, 313–341.

Chapter 15

Sahara Environmental Changes during the Quaternary and their Possible Effect on Carbon Storage

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Today the Sahara comprises an area of more than eight million square kilometers of the desertic geobiome (ecosystem). Vegetation is sparse, rare or absent. Lithosoils are dominant between large sand dunes areas (Erg) and organic carbon is low to absent in the soils. The total vegetation biomass (phytomass) and soil organic matter probably does not exceed 8–15 gigatonnes (10^{12} kg) of carbon. A mean annual rainfall of 100 mm or 150 mm is usually taken as the edge of the Sahara, but wide areas receive less than 5 mm annual rainfall.

15.1. Pleistocene Climatic and Environmental Oscillations

The Saharan environment has undergone a succession of changes during the Quaternary. The main modifications occur as rapid changes between two opposite tendencies which can basically be termed 'wet' and 'dry'. During 'wet' intervals on the coast, Atlantic ocean water extended up to 100 km inland as large gulfs on the flat lowlands, producing a very irregular coastline well recorded by shell sand/limestone deposits. Such transgressions occurred several times, and are dated from: isotope stage 7, stage 5, stage 3 (usually below the present sea level) and during the Mid-Holocene (also called the Nouakshottien). These marine beds may have overlain the 'Continental terminal' rocks (sandstones, sandy clays, iron duricrust) in the Tertiary (probably Oligocene to Pliocene). They contain material reworked from kaolinitic weathering on land at a time when the climate of the Sahara was tropical humid.

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During each wet interval (with high transgressing sea levels, corresponding to interstadials in the temperate zone), inland sand dunes tended to be inactive, flattened, covered with Sudanian to Sahelian vegetation, and altered by soil development. The soils, morphology, and associated rich prehistoric sites suggest several generations of sand dune reactivation, separated by wet periods: the latter are Neolithic with artifacts during the Holocene; Aterian (*i.e.* middle Palaeolithic) during stage 3 and possibly 5d; and Acheulean also related to wadi terracing. Evidence of the recent wet maxima are soft, locally laminated diatomites and lacustrine limestone, dated mainly 6–9 ka and 20–30 ka. Older limestone are associated with Acheulean industries; and some are dated from the isotopic stage 5 and 7 (Petit-Maire *et al.*, 1982).

During wet episodes, the Sahara was a landscape over which vegetation, fauna and soils were widely expanding, even in its most remote parts, and humans left their mark everywhere. During these periods five to ten times (and possibly more) carbon than at present was stored in the total biomass.

The paleohydrologic conditions that prevailed during the wet phases in the Sahara are characterized by a high groundwater level. In the former dune areas, high porosity favored lateral renewal of the water and permitted fresh water fauna to survive several thousand years in some lakes. But increasing evaporation may alter the water balance, transforming lakes into sebkha-type depressions while increasing the salt concentration in both residual surface water and ground water. It is estimated that less than one percent of the lacustrine deposits have been preserved in inter-dune depressions or sebkhas, this because of strong deflation and erosion. Most of the lacustrine deposits in the Sahara have been blown over the Atlantic ocean and even as far as the Caribbean.

During the very dry phases, the strong winds produced two distinct landscape features, depending on the local eolian sand budget. Pre-existing soils were stripped away along wide deflation corridors leaving stony pavements (reg, regolith) on which the former drainage pattern is clearly visible as wadis and their terraces. The stones and associated prehistoric tools show different patinas which locally are very dark in the older Acheulean. The other feature basically consists of wide stripes of sand in various dune forms aligned with the dominant wind. During lowered sea level (corresponding to the glacial maximum in high latitudes) the sand stripes extended onto the continental shelf. The sand was thus introduced to the marine environment, directly by wind transportation and coastal redistribution but also indirectly by turbidities which moved it into the submarine canyons. Acheulean tools under the older dunes are well preserved, and tend to have no patina, especially in the Mauritanian Sahara. It appears that the sand stripes are being repeatedly reshaped from an older stock of sand.

Approximate dates for the arid periods are indicated by marine and lacustrine deposits and by soils and artifacts above and below the dune systems. The last great and general sand dune reactivation began with the Last Glacial Maximum. But it probably lasted until 14–12 ka BP (locally designated Ogolian A). Other periods (Ogolien B, Ogolien C, *etc.*) are believed to correspond to isotopic stages 4, 6 and probably older (8, 10?). There were several important reactivations during the Holocene; and another during the last twenty years, this with a strong anthropogenic component. With such complex reworkings of the sand, dating of the dunes must take into consideration the 'fluid' behavior of the system on millennial time scales. A similar difficulty is posed for the dating of underground water.

During the dry periods the ground water level might have fallen even lower than at present when it may often be several hundred meters below the general surface. Moreover, during times of lowered sea levels, the freshwater table at the coast adjusted its slope and drained more towards the sea. Probably this produced an oasis mechanism whereby fresh water arose as springs around depressions no longer affected by sea water pressure. In each case, the surrounds were probably a refugium for flora, animals and human beings (for instance, during the Aterian at stages 3/4/5a). Inland, in depressions when lakes dried up, similar springs appeared, juxtaposing fresh water and salt that had accumulated through evaporation. During these intervening arid phases, life will have been concentrated in but a few points and the total biomass much reduced from its present levels.

15.2. The Holocene Fluctuations

The lacustrine deposits sometimes preserved in even the most arid sectors of the Sahara, as it is today, testify that summer monsoon rains regularly reached north of the Tropic of Cancer around 8,500 radiocarbon years BP. Throughout this 'Climatic Optimum', meridional temperature gradients were diminished, while the intensity of the anticyclonic regime was markedly weakened by a decrease in meridional transport. Rainfall in summer, spring and autumn was enhanced by the increased power of the Atlantic monsoon flow. There was also (though especially around 7,500 years ago) an increase in the rains in winter, the season of least evaporation.

Evidently then, the climatic and hydrological parameters of the Saharan region were very different from those of the present. The lakes which existed throughout the Sahara were fed by rivers, runoff and subterranean water tables. These lakes persisted year-round, and perhaps lasted for centuries or millennia without ever drying up. They were fed—according to their latitude—from either the north or the south.

However, the regions that have been best studied in this respect are close to large fluvial basins such as the Senegal, the Niger, the Chari-Longone, and the Nile. Such lakes may largely respond to fluctuations in climate happening somewhere far upstream in the river catchment areas, well outside the area of immediate interest. It is important to distinguish between those lakes that were fed by rivers and those fed only by ground water. It is the signal from the latter that best reflects the local situation as set within the broader African framework.

Among the Holocene lacustrine basins that were principally influenced by neighboring rivers are the Chemchane and Tichitt, yet to be studied in Mauritania. Other large Holocene lake basins were fed only by relatively local rivers and rainfall, for example the Azrag or Idjill. These have yet to be studied in the same detail. The existence of large locally-fed late Pleistocene/Holocene lakes seems to indicate that, at various times between 16,000 and 8,000 years ago, the monsoon isohyets were displaced a good 800 km relative to their position during the glacial maximum. For instance, amongst the 376 radiocarbon dates in the database for Mauritania (Vernet, 1992), 40 concern the environment during the period between 9,260 and 6,000 years ago. Those which relate to lake sediments, plant fragments, peats and bone indicate conditions decidedly more humid than at present. Six dates indicating human habitation are also found for virtually the same period (9,120–6,020 years BP), and these confirm the presence of water and of lakes.

These data indicate broadly the amplitude of climatic and hydrogeological changes but not the origin of the rains nor the exact duration of humid phases. Only a detailed study of laminated deposits in several lacustrine basins will enable us to track latitudinally the evolution of Holocene seasonality. Key areas along the large Sahelian rivers of western Sahara have not yet been the object of systematic research. Such study could contribute to definitive measurement of the variations the African monsoon showed during the Holocene.

15.3. Conclusion

A feature of the changing Quaternary environment is that the events occur on all time scales (2–3 y; 20–30 y; a few centuries; 2–3, 20 and 100 ka) depending on the specific biologic, hydrologic, oceanic, climatic, and astronomic dynamics. There are no long sustained 'stable' conditions. This constantly fluctuating behavior of the environment has appeared bounded until now by the two extremes of wetness and aridity. Recovery of the living environment takes longer than its destruction and, because the cycles or quasi-cycles are not symmetric, all these changes are cumulatively trending toward desertification on the million-year time scale. For this reason, the terrestrial carbon that was stored in the Sahara during the early Quaternary was probably 10 times larger than at present.

References

Petit-Maire, N. *et al.* (1982) *Le Shati: Lac Pléistocène du Fezzan (Libye).* Centre National de la Recherche Scientifique, Paris.

Vernet, R. (1992) *Le Sahara et le Sahel Paleoenvironnements et Occupation Humaine à la Fin du Pleistocene et à l'Holocene. Inventaire des Datations ^{14}C.* CMA, Meudon.

Chapter 16

Climatic Change during the Pleistocene/ Holocene Transition in Upland Western Maharashtra, Western India

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16.1. Introduction

Upland Western Maharashtra is drained by the Krishna and Godavari rivers and their tributaries. These rivers have their source in the Western Ghats and flow into the Bay of Bengal. The region is semi-arid (800–400 mm rainfall) due to its location on the leeward side of the Western Ghats. The major rivers therefore derive much of their discharge from the high rainfall source region (up to 6,000 mm), and bring water to the semi-arid zone downstream. The landscape is dominated by flat denudational surfaces at 1,200, 1,000, 800 and 700 meters above sea level (Kale and Rajaguru, 1988) of pre-Quaternary age developed on 60 myr old Deccan Trap basalt. The rivers lack well-developed floodplains and alluvium is confined to a narrow belt less than 2 km wide. The non-alluvial part of the landscape is basalt bedrock, covered with varying thicknesses of weathered bedrock, locally called 'murrum' on which a soil has developed. This weathered mantle developed during the Tertiary, when the climate was comparatively humid. The thickness of the alluvium is generally less than 20 m. This alluvium includes several fills ranging in age from the Early Middle Pleistocene to the Late Holocene. The Holocene deposits are non-calcareous silts and sand and generally form an alluvial terrace 5–6 m high inset into an older Pleistocene alluvial fill terrace 10 to 15 m high. The Pleistocene alluvium is calcareous and dominated by sandy silt with lenses of gravels and fissured clays. Colluvium covers some footslopes.

The rivers of Upland Western Maharashtra are almost unaffected by either tectonic or sea level changes during the Quaternary. The rivers preserve evidence for

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several phases of aggradation and incision, which were responses to Quaternary climatic change. Rajaguru and Kale (1985), and Kale and Rajaguru (1987) have shown that most of the streams were aggrading during the last glacial period (21–12 kyr BP) and cutting/eroding through the alluvial fills during the post glacial period (10–4 kyr BP). Some evidence for aggradation is seen around 12 kyr BP. They showed that summer monsoons were weak during the Terminal Pleistocene and were strong during the Early-Mid Holocene (particularly between 7–4 kyr BP). The Late Pleistocene aridity seen in Western Maharashtra was severe. The most dramatic evidence of the severity is the filling of gorges seen on the Mandwe River near Chilewadi, and on the Mula River near Bote. The Mandwe evidence is particularly significant as it is within the high rainfall catchment zone.

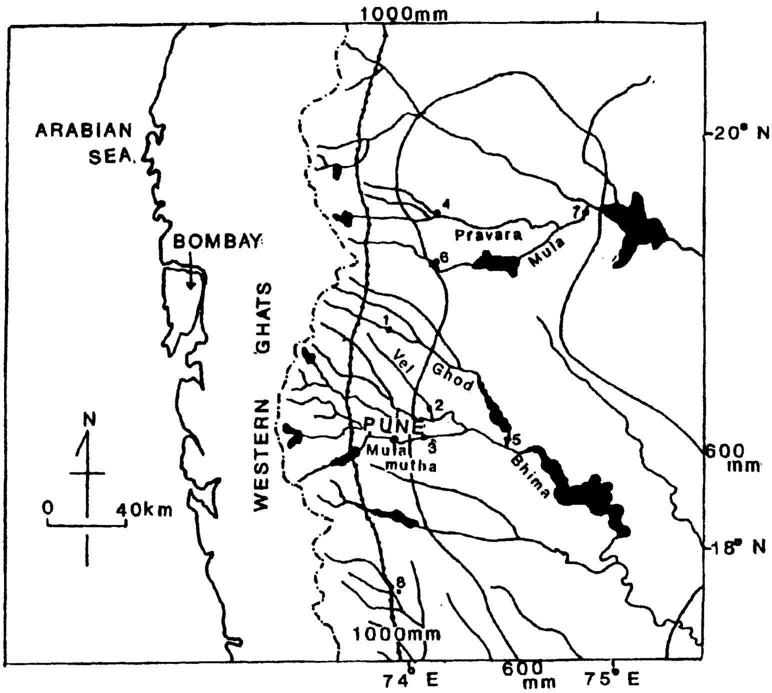
Mishra and Rajaguru (1993) have shown that the Quaternary record consists of short episodes of aggradation or erosion punctuating long periods of inactivity. Sadakata *et al.* (1993) showed that available radiocarbon dates from the region clustered into pre and post Last Glacial Maximum phases of gravel aggradation. This led us to infer that the rivers of the region have responded sensitively to global climatic changes. In this paper we focus on the sites where we have obtained absolute dates (radiocarbon, uncalibrated) between 14 and 7 kyr BP. Eleven sites have been dated within this timespan and the dates are given below in the order of oldest to youngest. The dates are uncalibrated using the radiocarbon half life of 5,730 years.

16.2. Variability of Fluvial Response to Climatic Change

While all the fluvial systems of Upland Western Maharashtra appear, for the reasons outlined above, sensitive to climatic change, it is quite likely that different parts of the fluvial system respond differently to changes in climate. Our ongoing research is focused on understanding this variability. The climatic gradient in the region is one of high rainfall in the windward Western Ghat zone with decreasing rainfall as one travels to the east, into the rainshadow zone. Out of the rainshadow zone, the rainfall tends to increase again. As the western Ghats are oriented almost North/South, the longitudinal position is a good approximate indicator of climatic differences and similarities. The rivers that originate in the high rainfall Western Ghats respond to climate differently to those that originate in the rainshadow zone to the east. Besides this, the size of the drainage basin, relief and surface cover are other factors creating variability in the fluvial response to Quaternary climatic change. In our current research we are contrasting the climatic record in the Western region, with that in the Eastern region. We find that while aggradation was ubiquitous during the Late Pleistocene, in the region West of approx. 75°E. long.; the region to the East of 75°E. shows evidence of Holocene aggradation. Data are currently more abundant for the Western zone and only the sites of Shaksal Pimpri, Ranjegaon and Akoni belong to the Eastern zone. All of them date from the Early

Table 16.1. Radiocarbon dates from the Pleistocene/Holocene transition in Upland Western Maharashtra

Site name	River	Material	Lab no.	Date	Context
Sangamner	Pravara	Shell	PRL 470	14840±350	gravel
Chandoli	Ghod	Shell	BS 1227	13510± 200	gravel
Nevasa	Pravara	Shell	BS 576	13220± 190	gravel
Sashtewadi	Mulamutha	Shell	BS 1226	13020± 190	gravel
Inamgaon	Ghod	Shell	BS 146	12040 ±150	gravel
Gargaon	Mula	Wood	TF 1111	10310 ±155	silt
Asla	Krishna	Shell	TF 1173	10035 ±150	silt
Talegaon	Vel	Shell	BS 1228	9420± 90	gravel
Ranjegaon	Sindphana	Shell	BS 1256	7800 ±130	silt
Shaksal-pimpri	Sindphana	Shell	BS 1259	7800 ±100	silt
Akoni	Nandi	Shell	BS 417	7150 ±80	gravel



1 Chandoli
2 Talegaon
3 Sashtewadi
4 Sangamner
5 Inamgaon
6 Gargoan
7 Nevasa
8 Asia
Lake

Figure 16.1. Upland Western Maharashtra.

Holocene. As Pleistocene archaeological sites are preserved mainly when there is an opportunity for burial, the two zones preferentially preserve archaeological sites of different periods.

16.3. Archaeological Evidence for the Period between 14–7 kyr

Most of the archaeological evidence for this period comes from the dated gravels themselves. Microliths are associated with all the dated samples except Asla and Gargaon, where the sample for dating comes from a silt deposit. The sites are therefore important for their geochronological and palaeoenvironmental information, but do not tell us much about human behavior during this period.

From an archaeological point of view, the most interesting site is Shaksalpimpri. Here, microliths, animal bones and some potsherds along with the datable shells come from the cultivated surface of a field. The shells were dated to 7.8 kyr BP. The association of the bones, microliths and shells therefore may be acceptable, although the material was found in a disturbed context. However, the earliest dates for pottery in this region go back only to 3.5 kyr BP. Therefore the association of the pottery with the shells, bones and microliths requires further documentation. The pottery is however entirely handmade, with mainly open basin forms. It is also similar to pottery reported by Sali (1996) from Patne, which he places stratigraphically between the Mesolithic and Chalcolithic. The association is therefore *prima facie* reasonable, although, considering its significance further work is required to confirm it. The association of the bones with the date of 7.8 kyr BP is supported by the fluorine/phosphate (100F/P205) ratio of the bones. The element fluorine occurs in most groundwater and replaces hydroxyl (OH) ions from hydroxyapatite in the buried bones. Fluorine gets accumulated in bones as the time passes. Fluorine/phosphate ratio is a better indicator of the relative age than fluorine content alone, since the ratio is independent of the density of bone which may change differently in varying soil environments. The theoretical maximum ratio is 8.92. The ratio increases with time. The method was first used by Oakely (1955) and helped in proving that the Piltdown man was a fraud. In India Joshi and Kshirsagar (1986) showed that this method was useful in relative dating. The saturation value of around 8 has been observed in fossils from the Late Middle Pleistocene period. The bones from Shakshal Pimpri gave a value of 3.94, which matches quite closely with 3.06t from Ranjegaon, which has the same C14 date.

The importance of this site is that both the bone and shells show evidence of human modification. Shakshal Pimpri therefore was the site of a human settlement during the early Holocene. The occupation was on the surface of the Pleistocene terrace and the occupational debris was preserved by a cover of flood silt soon after the abandonment of the site. The matching date from Ranjegaon, on the same river, in a similar geomorphological context, implies that they were affected by floods in the same period. There is less cultural material at Ranjegaon, and although shells and microliths were found, only a single bone fragment was, and no pottery.

16.4. Description of Dated Sites

16.4.1. SANGAMNER (*Indian Archaeology:* A review 78/79:105)

At Sangamner, a gravel 20 cm thick interbedded with clays was dated. This gravel is fine, sandy and shows cross beds. It is sandwiched within clays at 7 m above the modern Pravara level. The gravel, being a component of a floodplain facies is indicative that at 14.8 kyr BP the Pravara River was still in an aggradational mode.

16.4.2. CHANDOLI (this paper)

The dated gravel at Chandoli is a one meter thick sandy, pebbly gravel that has cut into older alluvium (see Fig 16.2) It occurs at 10 m above the present bed level of the Ghod River and is exposed over an area of 100 m x 20 m. It is not covered by any later sediment. The bivalve shells (which were used for dating) are whole single valves, abraded by transport in the gravel. The gastropod shells, however, are unabraded. Compact basalt and abraded calcrete nodules derived from the older alluvium are predominant in the gravel lithology. This gravel also contains a microlithic industry. The tools are also slightly abraded. The abrasion of the tools and shells is important in linking them with the phase of gravel deposition.

Two features are important in interpreting this gravel deposition in terms of fluvial processes. The first is the presence of a well-developed soil on the older alluvium that has been cut into by the gravel. This implies that before the deposition of the gravel, the river was not in an aggradational mode; it must have been at least slightly incised into the older alluvial deposits. The second feature is the complete isolation of the gravel. It is not part of a floodplain with both overbank and channel facies. It appears to have been deposited by a flood that was strong enough to deposit the gravel in an overbank situation, while eroding part of the older alluvium. The channel of the Ghod River just upstream of the gravel flows through a stretch of bedrock with a constricted channel. The flood waters flowing through the constricted bedrock stretch would have had an increased stream power allowing the lifting of gravel into suspension. This gravel therefore records a period of increased floods around 13.5 kyr.

16.4.3. NEVASA (Kale and Rajaguru, 1988)

The gravel at Nevasa is about one meter thick and cuts into Late Pleistocene silty deposits. It is similar to the gravel at Chandoli and Sashtewadi in its unconformable relationship to the older alluvium. At Nevasa however there is no soil preserved on Pleistocene alluvium at the site were the gravel is exposed, although further away from the river, it is present. This shows that erosion of the Pleistocene alluvium had begun before the deposition of the gravel. The gravel also shows relatively good sorting and rounding.

16.4.4. SASHTEWADI

The dated gravel at Sashtewadi is a one m thick sandy pebbly gravel containing abraded bivalve shells and a few microliths. This gravel cuts into the older alluvium at Sashtewadi (Fig 16.2). It is first seen cutting into a reddish silt at 4 m above the Mula bed and then rises to about 7 m above the river bed where it cuts into the soil developed on the older alluvium. The condition of the shells in this gravel are single, unbroken abraded bivalve shells like those seen in the Chandoli gravel. The gravel is also texturally and lithologically similar to the gravel at Chandoli. It is sandy—pebbly with compact basalt and calcrete nodule pebbles.

The two features of the isolated gravel cutting into older alluvium on which a soil had developed are present at Sashtewadi as at Chandoli. The local geomorphic setting is also similar to that at Chandoli in that the river runs through a stretch of minor rapids developed in the bedrock channel at this locality. This could there-

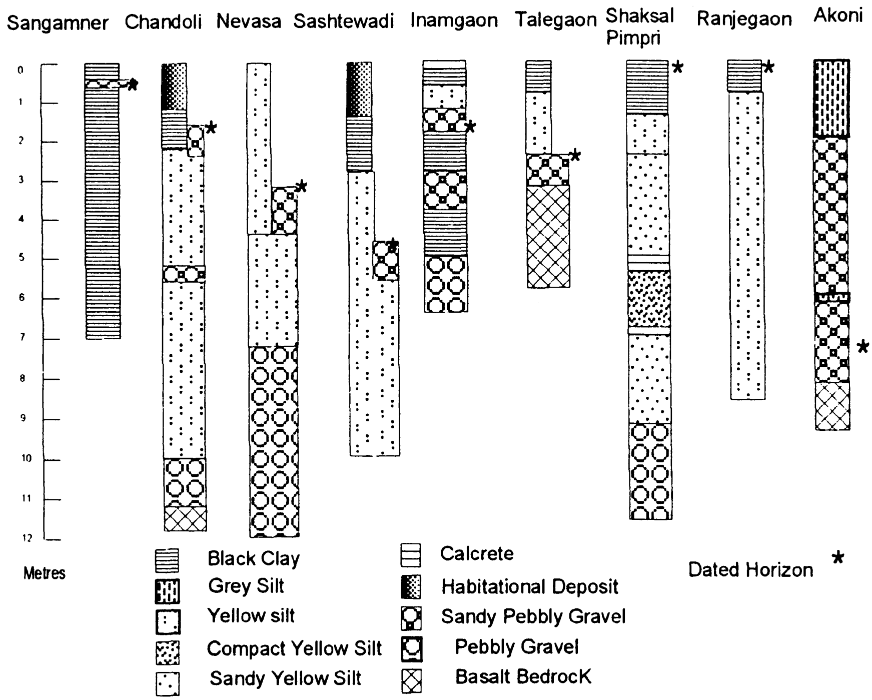


Figure 16.2.

fore be a factor in this site recording floods. The dates from the two sites differ by about 500 years, although the standard deviations of the dates just overlap. The presence of abraded bivalve shells in both the gravels implies that the gravel deposition was not one event. The shells would have to both grow in the gravel and then be transported in it. The gastropod shells however show no abrasion. The gastropod shells are much more fragile than the bivalves. The gastropods present in the gravel were not transported and therefore belong to the final phase. The gravels at Chandoli, Nevasa and Sashtewadi therefore are important evidence for increased discharges in the rivers during 13 to 14 kyr BP.

16.4.5. INAMGAON (Rajaguru *et al.*, 1979)

The Inamgaon gravel matches the Sangamner gravel in its character. The Inamgaon gravel is also a thin bed of cross bedded sandy gravel interlayering within the clays (Rajaguru *et al.*, 1979). A recent visit to the site showed that this gravel is part of a 5 m thick fill inset into the older alluvial fill 10–15 m thick.

16.4.6. GARGAON (*Indian Archaeology:* A review 1972/73:)

The wood from Gargaon is from a brown silt that abuts the Pleistocene alluvium. It is reported (Rajaguru and Kale, 1988) to be 12 m above the river level.

16.4.7. ASLA (Rajaguru and Kale, 1985)

The shells at Asla do not come from a gravel, but were found within silty alluvium only a few meters above the modern Krishna river. This data is important in showing that by 10 kyr BP, and after the minor aggradation of 12 kyr BP the rivers have reached almost the modern level.

16.4.8. TALEGAON (this paper)

The shells at Talegaon were collected from a gravel that is exposed in a field. The positioning of this gravel in relationship to the alluvium is being further studied.

16.4.9. SHAKSAL PIMPRI (this paper)

Shells were collected from the surface of a cultivated field on a 10 m Pleistocene terrace at Shaksal Pimpri. These shells show old breaks, due to human activity. Animal bones, microliths and some potsherds were found in the same context. The shells were dated to 7.8 kyr BP. While the artifacts could have survived on the terrace surface for the 7.8 kyr, it is unlikely that bones and shells would have (especially in the unabraded and unweathered condition they were found in), unless they were covered by some sediment. Therefore we think that the evidence for human activity was preserved by a thin cover of flood silt, shortly after the abandonment of the site. This therefore dates not only the human activity but also the flood which preserved it.

16.4.10 RANJEGAON (this paper)

The shells at Ranjegaon occurred in an identical situation to those at Shaksal Pimpri, except that the material was less abundant and no potsherds were found.

16.4.11 AKONI (Kale and Rajaguru, 1984)

Microliths and shells were found in a gravel exposed in a well on the Nandi river, a small tributary of the Bhima. Thus this shows the aggradation of one of the low order streams during the Early Holocene in the more easterly zone.

16.5. Pleistocene/Holocene Transition in Western Upland Maharashtra

The limited data available suggest that the fluvial deposits of Western Upland preserve a sensitive and complex record of Quaternary climate change. The rivers not only adjusted to changing climate, but different components of the fluvial systems varied in their response. While this does complicate interpretation, it also means that different components of the system were sensitive to different climatic variables. Depositional records absent in one region are present in others. With a more complete record, a richer archaeological and palaeoenvironmental interpretation becomes possible. Based on present data, we can divide the Pleistocene/Holocene transitional period into the following stages:

16.5.1. LATE PLEISTOCENE AGGRADATION

The period between 26–14 kyr was one of aggradation. Most of the dates from this period come from gravels interlayering with floodplain deposits. These gravel are poorly sorted, with a large component of reworked calcrete nodules. Although it is the gravels that have been dated, they represent only a minor component of the terrace fill, which is predominantly silty. This is the major phase of Late Pleistocene aridity.

16.5.2. INITIATION OF THE INCISIONAL PHASE BETWEEN 13–14 KYR BP

Three sites have been dated to this phase—Nevasa, Chandoli and Sashtewadi. The dated units are all gravels which have an unconformable relation to the Late Pleistocene alluvium. The lithology and rounding of the gravels also shows better rounding and more mature lithology compared to the gravels of 26–14 kyr. We have interpreted these gravels as having been deposited during the initiation of the incision into the Late Pleistocene alluvium. These gravels may be restricted to favorable stretches of the rivers.

This evidence goes well with the finding that the Himalayas were free of ice below 3,100 m by 15 kyr (Singh and Agrawal, 1976). The first major phase of deglaciation was very rapid throughout the world, with near interglacial conditions being established before the climatic reversal during Younger Dryas times.

16.5.3. RAPID/COMPLEX CHANGE IN CLIMATE BETWEEN 12–10 KYR BP

For the time period between 10 and 12 kyr BP the climatic evidence is varied, suggesting either rapid climatic change or different responses between the dated sites or both. Therefore we find that on the Inamgaon floodplain aggradation was going on at 12.0 kyr and at Gargaon silty aggradation at 10.3 kyr while at Asla, the river was close to the modern bed level at 10.0 kyr. The records of climate for this time period from other parts of the world do indeed show this to be a period of rapid change in climate. More data is needed to understand it better.

16.5.4. EARLY HOLOCENE FLOODS

The covering of the microlithic sites at Shakshal Pimpri and Ranjegaon have been interpreted as evidence for large floods during the early Holocene.

16.5.5. EARLY HOLOCENE AGGRADATION

The gravel at Akoni shows the aggradation of one of the minor tributaries with a drainage basin entirely within the semi-arid zone. Though such a tributary might become defunct during the periods of arid climate, the aggradation indicates a response to a discharge capable of transporting the accumulated sediment supply. The Nandi River was therefore aggrading, during the time when the larger rivers, with discharge from the Western Ghats, were eroding the Pleistocene floodplains.

16.6. Human Response to Environmental Change during the Pleistocene/Holocene Transition

The data on the human response to changing environments during the Pleistocene Holocene transition is scanty in comparison to the evidence for environmental change. The most important point is that man was present in the region during all the phases of climatic change. Artifacts are present in all the dated gravels. Moreover, their density in the gravels is quite high. The artifacts are usually unabraded or slightly abraded. This would imply they were discarded in the gravel, rather than washed in from a different location.

The stone industry during the Pleistocene/ Holocene transition was microlithic. Small chalcedony blades were backed and inserted into handles to make tools. The main types of backed blades are lunates and points.

Therefore Upland Western Maharashtra continued to be a congenial home to the Terminal Pleistocene hunter gatherers, as the allochthonous rivers were perennial water sources even in the rain shadow zone. Badam (1979) has shown that the region supported a rich herbivorous fauna even during the Late Pleistocene. The rapid fluctuations in climate must have provided a challenge to the Late Pleistocene hunter-gatherers, but much more archaeological data are required before the human response can be properly known.

16.7. Conclusion

Late Pleistocene hunter-gatherers have coped with the very drastic climatic fluctuations during the Pleistocene Holocene transition although we can imagine that the adjustment might have been difficult. In contrast, modern human populations have increased their vulnerability to environmental changes in spite of a more sophisticated technology due to the increased demand on resources. Understanding past climatic changes and planning for future ones therefore is of prime importance.

The major purpose of this particular paper has been to present the new data from Western Maharashtra that shows not only that the Upland rivers responded to Late Pleistocene aridity by aggradation, but that minor episodes of aggradation and erosion are also present recording the response of the fluvial systems to short episodes of climatic change during the Pleistocene Holocene transition. The gravels from Nevasa, Chandoli, Sashtewadi and Talegaon, representing overbank gravel deposition in channel constriction localities during floods, have been recognized for the first time. The dating of an incisional phase before 13.5 kyr has also been suggested for the first time based on the dates from Chandoli and Sashtewadi and Nevasa. A minor aggradation is seen from the evidence from Inamgaon and Gargaon. The rivers have reached to the present entrenched levels by around 10 kyr.

The rivers of Upland Western Maharashtra are thus beginning to yield a more detailed palaeoclimatic record. At present the inferences have been made on relatively little data. We expect further work will give a more complete picture of the impact of past climatic changes on the fluvial systems and past humans living in Upland Western Maharashtra.

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References

Badam, G.L. (1979) *The Pleistocene Fauna of India.* Deccan College, Pune.

Joshi, R.V. and Kshirsagar, A. (1986) *Chemical Analysis of Archaeological Bones from India: Fluorine and Fossilation Process.* Deccan College, Pune.

Kale, V.S. and Rajaguru, S.N. (1988) Morphology and denudation chronology of the coastal and upland river basins of western Deccan Trappean landscape (India): A collation. *Zeistchrift fur Geomorphologie* 32, 311–327.

——. (1987) Late Quaternary alluvial history of the northwestern Deccan upland region. *Nature* 325, 612–614.

——. (1984) Some observations on the Holocene geomorphic history of the Nandi River, Maharashtra. *Bulletin of the Deccan College Research Institute* 43, 111–113.

Mishra, S. and Rajaguru, S.N. (1993) Quaternary deposits at Bhedaghat, near Jabalpur, Madhya Pradesh. *Man and Environment* 18(2), 7–12.

Oakley, K.P. (1955) Analytical methods of dating bones. *Advancement of Science,* 11, 3–8.

Rajaguru, S.N. and Kale, V.S. (1985) Changes in fluvial regime of western Maharashtra upland rivers during Late Quaternary. *Journal of the Geological Society of India* 26, 16–27.

Rajaguru, S.N., Pappu, R.S. and Deotare, B.C. (1979) The terminal Pleistocene microlithic industry of Inamgaon. *Bulletin of the Deccan College Research Institute* 39, 150–159.

Sadakata, N., Maemoku, H., Rajaguru, S.N., Mishra S., and Fujiwara, K. (1995) Late Quaternary environmental change in the Pravara Basin, northwestern Deccan Upland, India. *Proceedings of the International Symposium on Palaeoenvironmental Change in Tropical-Subtropical Monsoon Asia.* Special Publication no. 24 of the Research Center for Regional Geography, Hiroshima University, pp. 43–56.

Sali, S.A. (1995) Neolithic of Maharashtra. *In* C. Margabandhu and K.S. Ramachandran (eds.), *Spectrum of Indian Culture,* Prof. S.B. Deo Felicitation Volume 1, pp. 44–45.

Singh, G. and Agrawal, D.P. (1976) Radiocarbon evidence for deglaciation in North-Western Himalayas, India. *Nature* 260, 232.

Chapter 17

Aspects of Climate Variability and the UNESCO International Hydrological Programme

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17.1. Introduction

This chapter is intended to provide an indication of how the contents of this book fit within the framework of the *UNESCO International Hydrological Programme* and associated partners (*e.g.* WMO, IGBP-BAHC, GEWEX) under the umbrella of WCP-Water (World Climate Programme-Water). To achieve this objective, this short contribution will select some of the cross-cutting issues which link some of findings from the paleoclimatic studies, with the problems of using outputs from atmospheric General Circulation Models (GCMs) in hydrology and water resources planning. Reference will also be made to contributions from the combined efforts of the micrometeorological-hydrology communities towards improving GCMs which forms a very active sub-project of the IHP. This latter aim requires my going into more detail so that the non-specialist has a better appreciation of the current difficulties in coupling hydrological models with GCMs. Subsequently the value in developing climatic variability scenarios based on paleoclimatic and historical evidence should therefore become apparent.

Theme 1 (Global hydrological and geochemical processes) of the current Fifth Phase of the International Hydrological Programme (IHP-V, 1996–2001) is divided into four sub-projects, *viz.:*

Project 1.1 Application of methods of hydrological analysis using regional data sets (Flow Regimes from International Experimental and Network Data Sets/FRIEND)

A. S. Issar and N. Brown (eds.), Water, Environment and Society in Times of Climatic Change, 335–355.

Project 1.2	Development calibration of coupled hydroecological/ atmospheric models
Project 1.3	Hydrological interpretation of global change predictions
Project 1.4	Strategies for water resource assessment and management under conditions of anthropogenic global climate change.

This monograph is a contribution toward project 1.4. Essentially, the IHP-V is pursuing the issue of climate variability from four perspectives:

- an assessment of paleoclimatic and historical climate variability, and the possible impacts on water resources-society-civilization (Project 1.4);
- the use of outputs from GCM scenarios to assess global warming on the hydrology and water resources (Projects 1.3 and 1.4);
- a focus on hydrological processes at different scales to help improve GCMs (Project 1.2);
- the evaluation of long-term hydrological data sets for the detection of climate variability (Project 1.1).

This account will highlight on some of the issues and contributions of these other projects in the context of this book.

17.2. Selected Issues from the Paleoclimatic Setting

Much publicity has been given to the impact of anthropogenic influences on global climate through the enhancement of greenhouse gases and aerosols through the First IPCC *(Intergovernmental Panel on Climate Change)* Assessment published in 1990 (IPCC, 1990). The recent Second IPCC Assessment has provided a comprehensive up-date on the scientific analysis and the water resources implications, including hydrology and socio-economic impacts (Houghton *et al.*, 1996; Watson *et al.*, 1996).

A salient message of these proceedings, however, is that until we have a much more comprehensive understanding of the complex combination of *natural forcing mechanisms* which influence *natural climate variability* (Street-Perrott, 1994), even on time scales ranging from decades to century, then it will remain difficult to filter out the more publicized anthropogenic effects from those connected with natural variability (Bonell, 1998). The paleoclimate changes described present daunting challenges to global climate modelers even if confined to the Holocene. Obviously, it is very desirable to be able to simulate by modeling past climate change. For one thing, uncertainty in future projections can thereby be minimized. As Oldfield (1997:2) remarked "... unless models can achieve an adequate level of realism in simulating past climate conditions, their performance in predicting future conditions will remain less certain than decision-makers would like."

These challenges are all the more formidable when climate variability in previous inter-glacial periods is considered. For example, the Eemien interglacial period (*c.* 135 kyr—115 kyr BP) is an obvious target for reconstruction because the global climate appears to be warmer than the Holocene, thereby providing an analog for future climate evolution. Results from the 1990–1992 Greenland Ice-core Project (GRIP, 1993) present disturbing evidence that the 'relative stability' of climate in the current Holocene interglacial may be an exception. Drawing on a variety of data (stable isotopic ratios; chemical and physical properties, and greenhouse gas concentrations in bubbles entrapped in the ice cores), the GRIP project revealed dramatic shifts in temperature and greenhouse gases (GRIP, 1993; Johnson *et al.*, 1997). In his summary of this project, White (1993) remarked the "Holocene climate appears to have been one, and only one, state, whereas the new results show that the Eemien had three. The middle matches our own Holocene climate. A significantly colder state and a significantly warmer state (> 2°C higher than at present) existed in the Eemien." What is particularly disconcerting is the way that these principal phases, acute yet radical transitions in climate occurred with "... mode switches ... completed in as little as 1–2 decades ... became latched for anything between 70 yr and 5 kyr" (GRIP, 1993:207). Such findings raise urgent questions as to why such dramatic oscillations have not occurred in the Holocene? Why has the Holocene exhibited such extreme climatic stability, maybe very much the exception rather than the rule for an interglacial? Could the current anthropogenic intervention induce Eemien-type instability into the Holocene? (GRIP, 1993; Dansgaard *et al.*, 1993).

The role of the ocean circulation and the nature of its coupling with the atmospheric circulation is thought to be critical to the above findings. For example, a major source of inter-annual variability in the atmospheric circulation is the North Atlantic Oscillation (defined in Lamb & Peppler, 1987[1]; Hurrell, 1995). The persistent above-average winter temperatures across Europe since the 1980s can be attributed to the NAO being 'latched' into one extreme phase (Hurrell, 1995). Sutton & Allen (1997), as summarized by McCartney (1997) indicate concerning the NAO that coupling between ocean and atmosphere rather than the atmosphere *alone* is the driving mechanism. These writers describe how the winter atmospheric circulation has a 'memory' of sea surface (temperature (SST) anomalies in the Atlantic Ocean (which persist for several years) and how such anomalies feedback to the atmosphere. SST heat anomalies tend to be isolated beneath a seasonally heated layer in summers and then exposed anew each winter by convective overturning of the upper oceanic water: this in interaction with the atmospheric turbulence associated with these stormy latitudes. The effects of this oceanic 'memory' of past conditions and in turn, of the atmosphere 'seeing' the SST anomalies from previous winters (through the winter exposure anew of oceanic thermal anomalies), provides a basis for the continued duration for extreme phases of the NAO. Moreover, the ocean-atmosphere coupling associated with the NAO is not exclusively a mid-latitude phenomenon unconnected with tropical Atlantic SST and related climate variability phenomena. Sutton & Allen (1997) present evidence of linkages between the subtropical and tropical North Atlantic SSTs which climatologists had previously tended to ignore (Sutton & Allen, 1997).

It is within this framework of ocean-atmosphere coupling that the NAO, or related processes, are thought to have been a contributory cause of the decadal-frequency climatic oscillations noted in the Eemien interglacial (GRIP, 1993). Sudden shifts in the oceanic circulation such as 'the Atlantic conveyor belt' (Broecker, 1995), which at present transports ocean heat via the Gulf Stream Drift north-east towards Europe, is an integral part of the above-mentioned ocean-atmosphere coupling. Adkins *et al.* (1997), having investigated of the chemical composition of oceanic sediments from the Bermuda Rise, suggest the termination of the Eemien interglacial can be attributed in part to abrupt changes in the deep water component (return flow) of the Atlantic conveyor belt circulation. A quantum reduction of water transfer in the latter might take only 400 years to complete. Thereafter the climate system did not revert, and ice growth took place. In his review of this work, Lehman (1997), remarked that the present inter-glacial nearly came to an abrupt end during the Little Ice Age that started in the late Middle Ages. Had the semi-permanent snowfields coalesced over the high plateaux of Labrador and Baffin Island, the albedo feedback (reflectance of solar energy) from the snow and ice coupled with decreasing summer insolation (on the decline locally for the past 9,000 years because of changes in weather patterns) "might have triumphed over interglacial warmth" (Lehman, 1997:119). If these events had coincided with a slight weakening of the North Atlantic oceanic conveyor belt, then the termination of the Holocene could have been conceivable. Enhanced Arctic ice melting, induced by 'greenhouse gas' forcing, has the potential to affect deep water formation (through water density effects) within the conveyor-circulation (Hurrell, 1995; Lehman, 1997). Consequently, the expressed concerns of Lehman remain applicable to future climate.

Elsewhere the recent simulation of the Last Glacial Maximum, LGM (21,000 years ago) by Ganopolski *et al.* (1998) (reviewed by Stocker, 1998) using a simplified, coupled ocean-atmosphere climate model demonstrated the crucial role of the hydrological cycle linked with its effect on the North Atlantic conveyor belt. With a reduction in precipitation, and an even more significant decrease in evaporation, which arose from the expanding ice cover in the North Atlantic (from 75°N to 55°N), the resulting surface freshwater anomaly caused the area of deep-water formation to move about 20° south. It is pertinent to note that the modeling work of Ganopolski *et al.* (1998) goes some way towards addressing Oldfield's (1997) earlier remarks. Prior to simulating the LGM, Ganopolski *et al.* model was first verified by being able to simulate modern climate at a global scale. The subsequent simulated changes within the LGM are broadly in agreement with evidence from various paleoclimatic-paleoceanographic data bases (Stocker, 1998)

17.3. The Application of Atmospheric General Circulation Models (GCMS) Linked with Climate Variability (and Change)

With the increasing sophistication of computer technology, these developments have enabled the simulation of global climate sensitivity to imposed forcing, such as increased carbon dioxide concentrations, forest conversion and land degradation, through the use of *atmospheric General Circulation Models,* known as GCMs. These models are the primary tools for providing a full three-dimensional representation of the atmosphere. At the basic level, the following boundary conditions have to be satisfied, *viz.*, (i) input of solar radiation at the top of the atmosphere; (ii) orography; (iii) land-sea distribution; (iv) the albedo of bare land; (v) surface roughness (vi) vegetation characteristics, and (vii) surface-subsurface moisture. Oceanic conditions such as Sea Surface Temperature (SSTs) are either among the prescribed boundary conditions or else simulated using a separate *Oceanic General Circulation Model (OGCM)* (Gates *et al.*, 1996; Loaiciga *et al.*, 1996). Consequently it is important to recognize that, although the acronym GCM is commonly understood to refer to an atmospheric General Circulation Model (Jager & Ferguson, 1991:573), this could be more appropriately termed an AGCM with prescribed SSTs (*e.g.* Mitchell, 1991:60). The latter term is, in fact, used in selected sources (*e.g.* Kalma and Calder, 1994; Gates *et al.*, 1996; Trenberth, 1997).

The most commonly used GCMs are outlined by, for example, Cess *et al.* (1990) and Gates *et al.* (1996). These models operate with large grid scales, their horizontal resolutions usually being in the range of 2.5° to 10° for both latitude and longitude. They incorporate the laws of conservation of mass momentum and energy for an air parcel; and include such key atmospheric processes as radiative transfer, cloud formation and precipitation together with boundary layer and surface physics. The vertical air columns over each grid are divided into layers to represent the thermodynamically graded nature of the atmosphere (Giorgi & Mearns, 1991; Loaiciga *et al.*, 1996).

Several aspects of GCMs need to be appreciated before their outputs can be related to the hydrological cycle. Originally these models were intended for the simulation of the average, synoptic-scale (say, 10° latitude and longitude), atmospheric circulation for specific forcing conditions in the short term. With such relatively coarse resolution, the capturing of processes related to 'local' forcings (at the drainage basin scale of a few kilometers) are inevitably lost which immediately leads to difficulties in their application to hydrological questions. The same can even apply, indeed, to the simulation of regional climates (Giorgi & Mearns, 1991; Shiklomanov, 1998). In other words, GCMs are most sensitive to large-scale forcings derived from Earth orbital characteristics, changing solar emissions, radiatively active gases, the oceanic-continental distribution and major topographical features as affecting the surface exchange of radiation; momentum; sensible heat and water vapor.

Significantly, SST anomalies and their relationship to such global features as the El Nino-Southern Oscillation (ENSO), the Sahelian periodic drought and North Atlantic Oscillation (NAO) are still not properly incorporated by GCMs because of the continuing need for a better understanding of the processes (and feedbacks) with such phenomena (Philander, 1990; Stone *et al.*, 1996; Hurrel, 1995; Bonell, 1998). Nonetheless, Gates *et al.* (1996:253–258) outline recent progress in simulating such interannual variability as that with ENSO in the tropics. The synchronous coupling of AGCM-OGCM would be a step forward. However, no comprehensive steps on these lines have yet been taken at the appropriate spatial or time scale. The oceanic circulation time scale is often measured in 10^2 to 10^3 years as compared with a day or less for the atmosphere, a contrast which still tends to inhibit effective AGCM-OGCM coupling (Loaiciga *et al.*, 1996). Therefore atmospheric GCMs do not fully incorporate the oceanic effects on climate (Lins *et al.*, 1997).

Elsewhere the greatest uncertainty in GCMs concerns feedback mechanisms with ice, water vapor, vegetation, sea water and, above all, radiative feedback from clouds (Giorgi & Mearns, 1991; Cess *et al.*, 1996). Most important is the role of cloud radiative feedback mechanisms and their role in precipitation (Lins *et al.*, 1997). Commonly, a 'flux adjustment' (energy, water and momentum) is applied to correct for these feedback processes. In this way, the differences between modeling based on SSTs and the continental climatic fields are kept to a minimum. In which connection, Lins *et al.* (1997) have remarked that the practice of using 'flux adjustments' (colloquially known as 'tuning' the models) was entirely artificial. However such measures have proved no longer necessary in recent trials with the GCM at the National Center for Atmospheric Research (NCAR) (Trenberth, 1997).

As regards hydrological applications, it is apparent that GCMs would be more efficacious if operated at smaller spatial resolution (between a few kilometers and 100 km). Two factors at present preclude this at global level. The computing CPU (Central Processing Unit) requirements are too large; and the parameterizations of the atmospheric and surface physics at the higher resolution are not comprehensively adequate.

17.3.1. GCMS APPLIED TO VOLCANIC ERUPTIONS

Some attention has been given in the preceding chapters to volcanic eruptions as a prime cause of climate variability in the Holocene. The eruption of Mt. Pinatubo on 15 June 1991 presented a unique opportunity to study the impact on global climate. The injection of some 30 million tons of sulphur dioxide aerosol into the stratosphere had caused an increase in the global albedo from 0.236 to 0.250 by August 1991. The transient, aerosol shroud resulted in a net cooling of 4.3 Wm^{-2} between 40°N and 40°S; and 8 Wm^{-2} between 5°S and 5°N (McCormick *et al.*, 1995). After correction for ENSO effects, global cooling of between 0.4°C and 0.5°C occurred. Despite the aforementioned problems with GCMs, such temperature changes were in line with the upper range of predictions from selected climate models (Pearce, 1993; McCormick *et al.,* 1995). Thus the simulation of future volcanic impacts on global climate is a feasible proposition.

Kress (1997) suggested, however, that the impacts of volcanic activity on global climate was primarily controlled by the aerosol chemistry, that is, the quantities of sulphur emitted. The conventional view was that volcanic dust causes cooling by blocking solar radiation. However, the residence time in the atmosphere has since been shown to be short. By contrast, "... volcanic sulphur emissions react to produce sulphuric acid, aerosols which have much longer residence times" (Carroll, 1997:543). These sulphuric aerosols are more efficient in backscattering and absorbing incident solar radiation. Therefore atmospheric cooling results (Kress, 1997).

The above atmospheric chemistry consideration is one explanation why the Mt. Pinatubo eruption was so effective on global climate, even though the resulting aerosol mass loading was smaller than Tamboro (1815) and Krakatau (1883) (McCormick *et al.*, 1995). It is thus evident that the quantity of volcanic dust ejected skywards is a less critical indicator for projecting climatic impacts.

17.4. GCM Scenarios and their Applications in Hydrology and Water Resources for the Projections of Global Warming

Through the process of the IPCC, projections of climate change (*e.g.* in precipitation and temperature) for scenarios with increased atmospheric concentrations of greenhouse gases (positive feedback) and aerosols (some of which counter global warming as a negative feedback) (Houghton *et al.*, 1996; Mitchell *et al.*, 1995; Andreae, 1996) are incorporated by *a selected group* of hydrologists into hydrological models (Arnell *et al.*, 1996) for water resources management assessment (Kaczmarek *et al.*, 1996). More recently, van Dam (1998, ed.) provided an overall assessment of this approach based on the IHP-IV project H-2.1, *Study of the relationship between climate change (and climate variability) and hydrological regimes affecting water balance components.*

As GCMs operate at far coarser spatial scales than hydrological models, it is necessary in this approach to downscale GCM results to the catchment scale. Several techniques are available to achieve this objective, as outlined by Arnell *et al.* (1996), *viz.*:

- the use of GCM-simulated changes in precipitation, temperature and evaporation for direct interpolation at the catchment scale;
- estimation of catchment-scale weather from large-scale climatic features, such as weather types and mean sea-level pressure fields;
- the use of stochastic (probability theory) methods for the generation of point-scale weather variation at a given point;

the use of limited-area models embedded (nested) within GCMs to simulate regional climate at a higher spatial resolution. This method will be discussed in more detail below.

17.4.1. THE DIFFICULTIES USING GCM OUTPUTS FOR HYDROLOGICAL APPLICATIONS, AND SELECTED RESPONSES FROM THE HYDROLOGICAL COMMUNITY

Broadly speaking, three kinds of difficulty are encountered when utilizing the outputs from scenarios: namely, the predictive unreliability of the present generation of GCM; the associated spatial resolution of GCMs being too coarse for drainage basin studies (Arnell, 1995); and the need to improve process-based hydrological models across different scales (Bonell, 1998). Such comments are even more pertinent when it is recognized that different GCMs often provide variable and even contradictory results, *especially for precipitation*, for the same region (Shiklomanov, 1998). Nonetheless depending on the scenarios which specify the expected rates of increases in greenhouse gas and aerosols concentration, there is a general expectation of global warming ranging from 1.5 to 4.5°C (Houghton *et al.*, 1996). *These projections are assuming that the comparatively stable climatic regime of the Holocene is otherwise maintained (which on the basis of earlier contributions in this book could be challenged).*

Any such critique of using GCM scenarios in hydrology should take into account the fact that GCMs "were not originally intended for the purpose of climate-change assessment" (Rind *et al.*, 1992:121). They were designed to simulate the large-scale atmospheric circulation in the short term (Loaiciga *et al.*, 1996). Moreover, "... there is a very serious mismatch of scale between the scenarios of climate change (for example, developed from GCMs with 500 km resolution) and the scale of operation of many climate impact models" (Giorgi & Mearns, 1991:193) and the latter includes so-called physically-based hydrology models (Leavesley, 1994; Bonell, 1998). Some comments by Trenberth (1997:133) are particularly pertinent here. He observed that "... it is easy to criticize a model. But the challenge to any proponent of say, a missing process is to insert into any one of these models (*i.e.* GCMs) and demonstrate that it makes a quantitative difference to the outcome. *The burden of proof that a model result is not valid should be on the critic, not the modeler*" (italics inserted here by the current author).

Despite the above challenge the hydrological community has only been selectively pro-active (Leavesley, 1994; Nemec, 1994; Bonell, 1998). This is in spite of there having been recent reviews of the use of hydrologic models to conceptualize and investigate relationships between climate change and water resources (Leavesley, 1994; Schultz *et al.*, 1995). The latter also includes the previously mentioned initiative from the IHP-IV project H-2.1, as presented by Leavesley (1998). *An additional group* of hydrologists (in fact, much influenced by micrometeorologists) have responded to Trenberth's challenge by concentrating on the exchange of latent heat and water mass (evaporation and transpiration) *in the vertical plane* alone, this in order to improve estimates of *total evaporation* (Ward and Robinson, 1990:79). Total evaporation is one of the parameters that has not been well represented in GCMs, it being highly sensitive to soil moisture, surface albedo and several vegetation characteristics (*e.g.* aerodynamic and stomatal roughness, turbulent and diffusion coefficients). To simulate the vertical exchange of water and energy, soil-vegetation-atmosphere transfer schemes (known as SVATS) are used for nesting within selected GCMs.

However, these SVAT schemes (models) are one-dimensional only; and simply study the exchange of energy, momentum and water flux between the surface and the overlying atmosphere in the vertical plane. They are later incorporated into GCMs by treating each grid square independently for SVAT modeling (Thomas & Henderson-Sellers, 1991). Since the mid-1980s several intensive investigations have been conducted in the field to validate SVATS. These have involved a concentrated effort at evaluating the regional climatic consequences of imposed forest-removal in the Amazon basin because the vegetation feedbacks there are poorly understood (Henderson-Sellers *et al.*, 1995a; Lean *et al.*, 1996). Land use conversion modifies the surface albedo, soil moisture and vegetation feedbacks which, in turn, affect the energy balance and water vapor exchange. As "... evaporation is a critical mechanism by which feedbacks are induced following surface (global) warming" (Loaiciga, 1996:100), considerable attention to improving SVAT schemes is warranted as part of the broader objective of land surface parameterization (Entekhabi & Eagleson, 1989; see review by Kalma & Calder, 1994). The experimental strategy conventionally followed is a variant of the HAPEX *(Hydrological Atmospheric Pilot Experiment)* strategy whereby vertical exchanges of water and energy fluxes (including the use of micrometeorological towers) are measured over a 100 km x 100 km area, at different spatial as well as time scales. The recent HAPEX-Sahel experiment in Niger provides a good example (Gouterbe *et al.*, 1994), the results of which were recently reported in a Special Issue of the *Journal of Hydrology* (vol. 188–189/1–4, 1997).

Before highlighting some of the key issues connected with land surface parameterization for GCMs, it is necessary to consider why the hydrological community at large has not been pro-active in climate change investigations. Experimental hydrology has traditionally ranged from the small experimental basin $\leq 10\ km^2$) down to the plot or hillslope scale. Even with the development of the more complex, 'process-based' distributed-parameter hydrological models, the focus has remained on the same range (Bonell, 1998). Several authors (*e.g.* Beven, 1989; Beven *et al.*, 1995; Bathurst & O'Connell, 1992) have provided a critique (including Leavesley, 1994; 1998) of the problems associated with the current generation of more complex hydrological models. These include the inappropriate use of algorithms developed at the microscale ($< 10\ km^2$) or even laboratory scale level (*e.g.* infiltration equations) for modeling applications at larger scales (Beven, 1989). Consequently, despite the existence of topographic-wetness models (*e.g.* Beven *et al.*, 1995), the lateral redistribution of both surface and subsurface water poses considerable problems for the provision of estimates of soil moisture on, for example, the GCM grid scale. Further complications arise where bedrock 'topography' (rather than surface topography) controls the lateral redistribution of subsurface moisture (McDonnell *et al.*, 1996), and where deep groundwater becomes a major source for forest transpiration during prolonged drought.

Algorithms were originally based on the simple, so-called 'bucket schemes' (Manabe, 1969; reviewed by Avissar & Verstraete, 1990:44; Loaiciga *et al.*, 1996:106) which consider the water balance of a soil 'block' (connected with the upper soil layer) with an allocated saturation water capacity for use in GCMs. The bucket model approach has subsequently been replaced by the more advanced SVAT schemes which represent more faithfully the vegetation and soil characteristics, as part of the estimation of vertical transfers of energy, moisture and momentum. Nonetheless SVATs are still not able to account for the availability of deeper water for transpiration. Yet this is a critical factor in relation to deep-rooted closed forests (Nepstad *et al.*, 1994) such as those found in the Amazon basin and elsewhere in the humid tropics, often in association with deeply weathered regoliths (Bonell, 1998).

Several issues arise from the above comments. There is a need to include a drainage basin perspective within HAPEX-type experiments (Bonell with Balek, 1993; Nemec, 1994) so as to parameterize better the horizontal redistribution of both surface and subsurface water across all scales up to the GCM level. Then there is the vexed issue of 'scale' and 'scaling' (see review by Bloschel & Sivapalan, 1995) for the *aggregation* (and *upscaling*) of field measurements and *disaggregation* (downscaling) of GCM outputs. As noted by Michaud & Shuttleworth (1997), 'aggregation' generally refers to the spatial averaging of some surface variable, *e.g.* albedo, fraction of vegetation cover, soil hydraulic properties. "Aggregation is a more limited enterprise than 'scaling' because scaling seeks to find a basis for relating (mathematically) a phenomenon at one scale to an analogous phenomenon at other scales" (Michaud & Shuttleworth, 1997:177).

A related problem is the use of measurements in GCMs taken at smaller scales. Such 'disaggregation' particularly applies to the characterization of soil hydraulic properties in order to estimate soil moisture and lateral subsurface water transfer, especially during storms at the hillslope or drainage basin range of scales (see review in Bonell, 1998). A complication is that the spatial heterogeneity inherent in soil hydraulic characteristics makes it hard to secure representative estimates. The latter is further aggravated by the widespread use of field methodologies originating from the laboratory scale (identified with the discipline of soil physics) for the *in situ* determinations of soil hydraulic properties (Bonell, 1998). Consequently there is continued uncertainty in developing for hydrological models algorithms which 'capture' more precisely the physical processes of water transfer even within microscale catchments ($\leq$ 10 km^2). This has caused experimental hydrology (*e.g.* hillslope hydrology) to continue taking for the most part a centripetal perspective by focusing research at the microscale; and so disregard the calls of Dooge (1986) to develop research initiatives at the mesoscale (*ca.* 10–10,000 km^2) or higher. The latter is an essential step in the development of new algorithms in order to provide the appropriate hydrological inputs into GCMs.

The principal exceptions to the above comments concern initiatives in the United States linked with the GEWEX *(Global Energy and Water Cycle Experiment).* Take, for example, the *First ISLSCP Field Experiment,* FIFE (ISLSCP being the International Satellite Climatology Program) which was conducted at a tall-grass prairie site in Kansas in the late 1980s. Wood and his co-workers (see a comprehensive review in Wood, 1995) used the drainage basin approach to pioneer the use of various techniques to upscale the fluxes of water and energy in the land-atmosphere system (Famiglietti & Wood, 1994a,b). A drainage basin perspective is similarly being followed in the current GCIP campaign in the Mississippi drainage basin (GCIP means *GEWEX Continental-Scale International Project*).

17.4.2. ASPECTS OF MACROSCALE HYDROLOGICAL MODELING LINKED WITH GCMS

It has been emphasized throughout that outputs at the GCM scale are incongruent with the requirements of hydrology at the smaller drainage basin scale. As Loaiciga *et al.* (1996:133) remarked, "GCMs cannot 'see' the smaller scale river basin because of their coarse grid resolution. *Subgrid hydrologic models* are needed to resolve the larger scale GCM predictions and predict smaller scale hydrologic phenomena." Various nesting schemes are being pioneered to provide from GCM outputs hydrological estimates which are more useful at the river basin scale. Such schemes are known as *Limited Area Meteorological models* (LAMs); and Giorgi & Mearns (1991) demonstrate how their utilization improves regional climate simulation in the western United States. LAMs take into account *mesoscale forcings* such as topography, by increasing model resolution over the region of interest through either nesting (embedding) a LAM within a lower resolution global model or else employing variable resolution grids (Giorgi & Mearns, 1991). Other writers refer to the above nesting schemes as Macroscale Hydrological Models (MHM, Vorosmarty *et al.*, 1993) or sub-grid parameterizations (Entekhabi & Eagleson, 1989; Wood *et al.*, 1992). The linkage between scale and resolution in models for simulating the hydrologic dimension in climate forcing was originally summated by Vorosmarty *et al.* (1993), and subsequently adapted by Loaiciga *et al.* (1996). Table 17.1 provides example of possible variants of the nesting or multiple scale approach.

Table 17.1. Scales and resolutions of models for hydrologic simulation of climate change forcings (adapted by Loaiciga *et al.*, 1996 from Vorosmarty *et al.*, 1993)

Scale	Model	Boundary	Resolution	
			Atmosphere	Hydrology
Continental	Linked GCM-MHM	Interactive in GCM	≥ 100 km (subhourly)	10–50 km (weekly to monthly)
Meso (Regional)	Linked meso-MHM	Prescribed or interactive in nested GCM	10–50 km (minutes)	1–10 km (daily)
Local	Hillslope-small catchment	Topographically determined	Prescribed point forcings (subdaily)	≥ 1 km (subdaily)

17.5. Selected Issues in Land Surface Parameterization of Hydrological Processes for Coupling with GCMS, Linked with Subgrid Hydrological Process Descriptions

GCMs have conventionally been based on spatially-locatable land surface parameters, this principally because of data availability. For example, Henderson-Sellers *et al.* (1995a:67) lists the estimated values of twenty-four ecotype and soil factors intended for inputting into a GCM in order to evaluate the impact of imposed deforestation on the Amazon basin regional climate.

A major challenge is to investigate the sensitivity of climate simulations to the spatial variability of these surface parameters. The SVAT schemes of Sellers *et al.* (1986) (the *Simple Biosphere Model*) and the process model of Warrilow *et al.* (1986) incorporate in some measure spatial variability within a model grid element. So, too, does the statistical-dynamic model of Entekhbi & Eagleson (1989) which is distinctive in addressing the spatial variability of soil moisture contents and their impact on run-off generation. All three models address the key issue of the spatial variability of precipitation within GCM grids. It has, in fact, been shown elsewhere (Jain *et al.*, 1992) that the performance of the more 'physically-based' hydrological models is most sensitive to precipitation estimation, a point often forgotten in the quest for better representation of other hydrological parameters such as soil hydraulic properties (*e.g.* field-saturated hydraulic conductivity or permeability) (Bonell, 1998).

Elsewhere, as Thomas & Henderson-Sellers (1991:904) remarked, "... one of the rationales for incorporating spatial variability of precipitation within GCM grid cells is to allow a more realistic generation of runoff. The GCM grid-average precipitation is generally of low intensity (a consequence of the spatial averaging of the point process that it represents) and thus runoff tends to be underestimated."

Through the use of various techniques, as described in Thomas & Henderson-Sellers (1991), an attempt was made to incorporate the spatial and temporal variability of rainfall, and then model the two types of runoff generation pathways at the surface, *viz.*, infiltration-excess (Hortonian) overland flow and saturation (saturation-excess) overland flow (see Bonell, 1993, 1998 for discussion of these storm pathways). Thomas & Henderson-Sellers (1991) developed adequate simulations for both annual and monthly periods when addressing the use of the infiltration-excess runoff generation mechanism *only.* However, for time scales of one day or less, modeling the surface hydrology "is more significantly a function of the spatial variability of catchment characteristics" (Thomas & Henderson-Sellers, 1991: 909). These include the infiltration capacity, field saturated hydraulic conductivity and the geomorphology. Consequently, more work is required on the explicit treatment of saturation-excess overland flow and the spatial characterization of runoff-generating mechanisms in general (Wood *et al.*, 1992). In that regard, Wood *et al.* (1992) describe a simple, three-parameter model termed the *variable infiltration capacity (VIC) model* for use in GCM grid cells.

The foregoing discussion has indicated that there is currently a concentrated effort towards land surface parameterization and a greater field testing and validation of SVAT schemes and methods for representation of sub-grid variability. For example, through a joint WMO/GEWEX-GCIP initiative a project has been launched to evaluate comprehensively several SVAT schemes linked with GCMs. Known as the *Project for Intercomparison of Land-surface Parameterization Schemes* (PILPS), part of its task is to make model intercomparisons coupled to at least one mesoscale model and one GCM (Henderson-Sellers *et al.*, 1993). Preliminary results presented by Henderson-Sellers *et al.* (1995b) indicate that a key factor behind model instability *can be the problem of soil moisture parameterization.*

Then again, some of the conclusions from the 'Tuscon Aggregation Workshop' in March 1994, as summarized by Michaud & Shuttleworth (1997), deserve serious consideration. This workshop focused on how to represent within GCMs various parameters at the grid-scale. Kabat *et al.* (1997) made a critical distinction between *effective* and *aggregated* soil parameters. They defined *effective parameters* as the area-average values or distributions over a domain with a *single, distinct textural soil type.* Such parameters can be estimated either by scaling or by the more recent pioneering use of the inverse modeling of unsaturated flow (see Kabat *et al.*, 1997 for detailed description of these techniques). In contrast, *aggregated soil parameters* represent GCM grid-domains with several textural soil types. Indeed, when one is concerned with the overall spatial averaging (within a GCM grid) of any heterogeneous surface variable (*e.g.* albedo) then this approach is appropriate. In practice, it is difficult to separate the process of *aggregation* from that of *scaling* in order to take account of sub-grid scale variability. As noted earlier, however, scaling is a more involved procedure which seeks to detect whether a particular parameter is *scale* *invariant* and then to express this mathematically. There is still considerable debate beyond the scope of this Chapter on the broad issue of scaling in hydrological modeling (see Kalma & Sivapalan, 1995). Aggregation, however, is more limited an enterprise. It can be simply the average (*e.g.* arithmetic or logarithmic) of a parameter over a heterogeneous grid cell and depends much on the size of the domain over which the averaging is carried out.

As regards the aggregation of parameters connected with vegetation, Michaud & Shuttleworth (1997:179) noted "... there has been significant and satisfactory progress in developing aggregation techniques" in flat or moderate topography (slopes ≤ 20%). This applies to such related surface attributes as albedo, aerodynamic properties, surface resistance and roughness length, all of which are connected with the evaporation process. Significantly, energy fluxes can be estimated to within 10% from the fluxes obtained from higher-resolution simulations. The most problematic areas still are (1) the aggregation of soil hydraulic properties, (2) lateral flow by near-surface water and groundwater flow, and (3) the effect of distinct lateral changes in vegetation height (Michaud & Shuttleworth, 1997). Wood (1997) emphasized that a failure to incorporate the small-scale variability of soil moisture has adverse effects on the estimation of areal-average evaporation within coarse-scale grids. In common with the findings of PILPS (Henderson-Sellers *et al.*, 1995b), the appropriate representation of soil moisture as well as lateral surface and subsurface flow (*cf.* Thomas & Henderson-Sellers, 1992) within GCM grids continues to pose a research challenge. In this context, the concluding comments of Kabat *et al.* (1997) are worth citing, "... the aggregated (soil) parameters, although predicting the evaporation fluxes well, failed to predict the water balance terms, such as downward percolation flux or runoff, that are associated in a more direct way with the soil water flow processes. This is a serious drawback which could eventually hamper the efforts focused on improvement of the hydrological cycle, and overall parameterization of soil hydrological processes in mesoscale atmospheric models and in GCMs" (p. 393).

Whilst not directly connected with the above remarks, it is pertinent that using the Hadley Centre model (the climate version of the U.K. Meteorological Office Unified Forecast/Climate GCM), Lean *et al.* (1996) detected the sensitivity of area-average evaporation to a change in the maximum infiltration capacity of the soil (and its effect on the runoff process) in the context of climate impacts arising from imposed deforestation and partial deforestation of the Amazon basin. The above modeling by Lean *et al.* (1996) incorporated revised land-surface parameterization from the ABRACOS (*Anglo-Brazilian Climate Observation Study*, 1990–1994) (Gash *et al.*, 1996). As part of that endeavor, Culf *et al.* (1995) detected 'a surprising result' that forest vegetation albedo showed seasonal variations with soil moisture which did not register a constant value, contrary to what had previously been assumed in GCMs.

17.6. The Importance of Lateral Surface Hydrology and Subsurface Hydrology in Connection with the Land Surface Parameterization of GCMS and the Links with the UNESCO IHP

A persistent theme has been the problematic representation of soil moisture and of lateral, surface and subsurface (including post-rainfall) flow within SVAT schemes. Henderson-Sellers *et al.* (1995a) indicated that modeling connected with GCMs freely utilize parameterizations at the microscale in selected SVAT schemes for application to coarse resolution GCM models. Moreover over this last decade (since the development of HAPEX-type field experiments) the prime focus has been on the parameterization in the vertical dimension of land-atmosphere interactions of energy, water balance and momentum. The horizontal redistribution of surface and subsurface water fluxes, using a traditional experimental catchment approach, has been largely neglected (Nemec, 1994; Bonell, 1998). It should be recalled that the International Hydrological Decade, IHD (1965–1974), the predecessor of the IHP, generated a global network of representative and experimental drainage basins. Capitalizing on the IHD experience, the IHP can appropriately take a lead in using a drainage basin approach to help resolve some of the issues raised above in connection with the appropriate parameterization of soil moisture. To achieve this objective, the UNESCO IHP (IHP-V Project 1.2) has been working very closely since 1992 with the IGBP-BAHC *(International Geosphere Biosphere Programme core project Biospheric Aspects of the Hydrological Cycle)*, this on the basis of a formal collaborative agreement between the two agencies. A principal aim has been the development of a hydrology-hydrochemistry plan using a nested drainage basin approach (Vorosmarty *et al.*, 1998; shown in Bonell, 1998), as an element within the planned large-scale field experiment for the Amazon basin known as LBA, the Large Scale Biosphere-Atmosphere Experiment in Amazonia) (Concise Experimental Plan, LBA, 1996; Integrated Science Plan, LBA, 1998). The LBA is scheduled to commence in 1998.

17.7. Overview and Future Actions of the UNESCO IHP

The preceding survey has indicated that the coupling of atmospheric and hydrological models within the GCM framework, still presents intractable difficulties in terms of achieving the desired accuracy of hydrological outputs for water resources planning. The scale problem, in particular, has been persistently emphasized as a major impediment to the successful coupling of atmospheric-hydrological models. As Lins *et al.* (1997) remarked, however, that scale discordance may be overstated because other factors—such as the mentioned feedback mechanisms (*e.g.* cloud, ice, water vapor, vegetation, ocean)—limit more the accuracy of climate model simulations. They go on to state "indeed, the paramount problem for anyone interested in using climate model output, whether for (water) resource planning, evaluation, or management, is the accuracy of climate simulation in space and time" (Lins *et al.*, 1997:64). Such comments imply that this is mostly a meteorological problem and that contributions from the hydrological community are limited (see conclusions and recommendations from IHP-IV project H-2.1 in van Dam, 1998). The mentioned improvement in SVAT schemes (linked with the surface hydrology) is an important departure from this attitude. Nonetheless GCMs are the only available tool for modeling climatic evolution, and the hydrological community have a responsibility to direct future research towards the better understanding of water transfer (in both the vertical and the lateral planes) across scales appropriate to addressing regional-scale hydrologic variability as linked with climate change. The UNESCO IHP (through the IHP-IV project H-1.1 and currently IHP-V project 1.2) is closely collaborating with the IGBP-BAHC towards these objectives through an evaluation of current experimental data and analytical methodologies used in large-scale field experiments such as the LBA. Paralleling this initiative are the current preparations for the *Second International Conference on Climate and Water* (Espoo, Finland) in August 1998. Through the umbrella of WCP-Water (World Climate Programme-Water), the UNESCO IHP [IHP-V projects 1.3 (*Hydrological interpretation of global change predictions*) and 1.4 (*Strategies for water resource assessment and management under conditions of anthropogenic global climate change*)] is concentrating its scientific networks on making a major contribution towards this conference. The diverse range of issues raised in this chapter will be comprehensively addressed at that meeting.

Three salient points emerge from this overview. It is apparent that considerably more time will be required before GCMs (coupled with hydrological models) can achieve the accuracy required for water planners to address the climate change issue. Secondly, a comparison between the Holocene and previous interglacials, such as the Eemian, has emphasized that the current epoch is unusual in terms of providing, on the whole, a comparatively steady and dependable climate. As White (1993:186) pertinently remarked, "... we humans have built a remarkable socioeconomic system during perhaps the only time when it could be built, when climate was stable enough to let us develop the agricultural infrastructure required to maintain an advanced society. We don't know why we have been so blessed, but even without human intervention, the climate system is capable of stunning variability." *Global climate is thus much more prone to secular change (and to short-term variability) than perhaps we have appreciated in the past.* Anthropogenically-induced forcing could provide a catalyst for destabilizing the Holocene which is the basis for the IPCC Assessments (Houghton *et al.*, 1995).

The studies in this book do, however, warn us that even though the Holocene has been comparatively stable globally, significant changes in regional climate have still occurred *without any anthropogenic causes*. The ramifications of future climate shifts on water resources coupled with a continued rise in global population, one must contemplate with alarm. In that context, a principal message of this book is that policy-makers should make greater use of analogs of Holocene climate variability within the context of water resources planning. Such steps should be taken now rather than wait for further improvement in GCMs. Current planning of the sixth phase of the IHP (2002–2007) is taking into consideration the preceding strategy.

Notes

1. The NAO Index is defined as the difference between normalized mean winter (December–February) surface-pressure anomaly for Ponta Delgadas (Azores) and that for Akureyri (Iceland).

References

Adkins, J.F., Boyle, E.A., Keigwin, L. and Cortijo, E. (1997) Variability of the North Atlantic thermohaline circulation during the last interglacial period. *Nature* **390**, 154–156.

Andreae, M.O. (1996) Raising the dust in the greenhouse. *Nature* **380**, 389–390.

Arnell, N.W. (1995) Scenarios for hydrological climate change impact studies. *In* H. Oliver and S.A. Oliver (eds.), *The Role of Water and the Hydrological Cycle in Global Change,* Springer, Berlin, pp. 389–407.

Arnell, N., Bates, B., Lang, H., Magnuson, J.J., and Mulholland, P., with Lead Authors (Fisher, S., Liu, C., McKnight, D., Starosolszky O., Taylor M.) and other contributing authors (1996). Hydrology and freshwater ecology. *In* T. Watson, C.M. Zinyowera, R.H. Moss, and D.J. Dokken (eds.), *Climate Change 1995. Impacts, Adaptations and Mitigation of Climate Change: Scientific-Technical Analyses*, Cambridge University Press, pp. 325–363.

Avissar, R. and Verstraete, M.M. (1990) The representation of continental surface processes in atmospheric models. *Reviews of Geophysics* **28**(1), 35–52.

Beven, K.J. (1989) Changing ideas in hydrology—The case of physically-based models. *J. Hydrol.* **105**, 157–172.

Beven, K.J., Lamb, R., Quinn, P., Romanowicz, R. and Freer, J. (1995) TOPMODEL. *In* V.P. Singh (ed.), *Computer Models of Watershed Hydrology,* Water Resources Publications, pp. 627–668.

Bloschl, G. and Sivapalan, M. (1995) Scale issues in hydrological modeling: A review. *Hydrol. Process.* **9**, 251–290.

Bonell, M. (1993) Progress in the understanding of runoff generation in forests. *J. Hydrol.* **150**, 217–275.

——. (1998) Possible impacts of climate variability and change on tropical forest hydrology. *Climate Change* (in press).

Bonell, M. with Balek, J. (1993) Recent scientific developments and research needs in hydrological processes of the humid tropics. *In* M. Bonell, M.M. Hufschmidt and J.S. Gladwell (eds.), *Hydrology and Water Management in the Humid Tropics—Hydrological Research Issues and Strategies for Water Management,* UNESCO-Cambridge University Press, Cambridge, pp. 167–206.

Broecker, W.S. (1995) Chaotic climate-global temperatures have been known to change substantially in only a decade or two. Could another jump be in the offing? *Scientific American* **273**, 62–68.

Carroll, M.R. (1997) Volcanic sulphur in the balance. *Nature* **389**, 543–544.

Cess, R.D., Potter, G.L., Blanchet, J.P., Boer, G.J., Del Genio, A.D., Deque, M., Dymnikov, V., Galin, V., Gates, W.L., Ghan, S.J., Kiehl, J.T., Lacis, A.A., Le Treut, H., Li, Z.-X., Liang, X.-Z., McAvaney, B.J., Meleshko, V.P., Mitchell, J.F.B., Morcrette, J-J., Randall, D.A., Rikus, L.J., Roeckner, E., Royer, J.F., Schlese, U., Sheinin, D.A., Slingo, A., Sokolov, A.P., Taylor, K.E., Washington, W.M., Wetherald, R.T., Yagai, I., and Zhang, M-H. (1990) Intercomparison and interpretation of cloud-climate feedback processes in nineteen atmospheric general circulation models. *J. Geophys. Res.* **95**, 16601–16615.

Cess, R.D., Zhang, M-H., Ingram, W.J., Potter, G.L., Alekseev, V., Barker, H.W., Cohen-Solal, E., Colman, R.A., Dazlich, D.A., Del Genio, A.D., Dix, M.R., Dymnikov, V., Esch, M., Fowler, L.D., Frasezr, J.R., Galin, V., Gates, W.L., Hack, J.J., Kiehl, J.T., Le Treut, H., Lo, K.-W., McAvaney, B.J., Meleshko, V.P., Morcrette, J-J., Randall, D.A., Roeckner, E., Royer, J.F., Schlesinger, M.E., Sporyshev, P.V., Timbal, B., Volodin, E.M., Taylor, K.E., Wang, W., and Wetherald, R.T. (1996) Cloud feedback in atmospheric general circulation models: an update. *J. of Geophysical Research* **101(D8)**, 12791–12794.

Culf, A.D., Fisch, G. and Hodnett, M.G. (1995) The albedo of Amazonian forest and ranch land. *J. Climate* **8**, 1544–1554.

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.S., Hammer, C.U., Hvidberg, C.S., Steffensen, J.P., Sveinbjornsdottir, A.E., Jouzel, J., and Bond, G., 1993. Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* **364**, 218–220.

Dooge, J.C. (1986) Looking for hydrologic laws. *Wat. Resour. Res.* **22(9)**, 46S-58S.

Entekhabi, D. and Eagleson, P.S. (1989) Land-surface hydrology parameterisation for atmospheric general circulation models including subgrid scale spatial variability. *J. Climate* **2**, 816–831.

Famiglietti, J.S. and Wood, E.F. (1994a) Multiscale modeling of spatially variable water and energy balance processes. *Wat. Resour. Res.* **30**, 3061–3078.

——. (1994b) Application of multiscale water and energy balance models on a tallgrass prairie. *Wat. Resour. Res.* **30**, 3079–3093.

Ganopolski, A., Rahmstorf, S., Petoukhov, V., and Claussen, M. (1998) Simulation of modern and glacial climates with a coupled global model of intermediate complexity. *Nature* **391**, 351–356.

Gash, J.H.C., Nobre, C.A., Roberts, J.M., and Victoria, R.L. (1996) *Amazonian Deforestation Climate.* Wiley, Chichester.

Gates, W.L., Henderson-Sellers, A., Boer, G.J., Folland, C.K., Kitoh, A., McAvaney, B.J., Semazzi, F., Smith, N., Weaver, A.J., Zeng, Q-C., and others (1996) Climate Models—Evaluation. *In* J.T. Houghton, L.G. Meira Filho, B.A. Callander, N. Harris, A. Kattenberg, and K. Maskell (eds.), *Climate Change 1995,* Ch. 5.

Giorgi, F. and Mearns, L.O. (1991) Approaches to the simulation of regional climate change: A review. *Rev. Geophysics* **29**, 191–216.

Goutorbe, J-P., Lebel, T., Tinga, A., Bessemoulin, P., Brouwer, J., Dolman, A.J., Engman, E.T., Gash, J.H.C., Hoepffner, M., Kabat, P., Kerr, Y.H., Monteny, B., Prince, S., Said, F., Sellers, P., and Wallace, J.S. (1994) HAPEX-Sahel: A large-scale study of land-atmosphere interactions in the semi-arid tropics. *Ann. Geophysicae* **12**, 53–64.

GRIP (Greenland Ice-core Project) Members (1993) Climate instability during the last interglacial period recorded in the GRIP ice core. *Nature* **364**, 203–207.

Henderson-Sellers, A., Yang, Z-L. and Dickinson, R.E. (1993) The project for intercomparison of land-surface parameterization schemes. *Bull. Am. Met. Soc.* **74(7)**, 1335–1349.

Henderson-Sellers, A., Pitman, A.J., Love, P.K., Irannejad, and Chen, T.H. (1995) The project for intercomparison of land-surface parameterization schemes (PILPS): Phases 2 and 3. *Bull. Am. Met. Soc.* **76(4)**, 489–503.

Henderson-Sellers, A, Mcguffie, K. and Durbidge T.B. 1995, Modeling the hydrological response to large scale land use change, *In* R.A. Feddes (ed.), *Space and Time Scale Variability and Interdependencies in Hydrological Processes,* Cambridge, pp. 63–88.

Houghton, J.T., Meira Filho, L.G., Callander, B.A., Harris, N., Kattenberg, A., and Maskell, K. (1996) *Climate Change 1995. The Science of Climate Change.* Cambridge University Press, U.K.

Hurrell, J.W. (1995) Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation. *Science* **269**, 676–679.

IPCC (1990) J.T. Houghton, G.J. Jenkins and J.J. Ephraums, eds. *Climate Change: The IPCC Scientific Assessment.* Cambridge.

Jager, J. and Ferguson, H.L. (1991) *Climate Change: Science, Impacts and Policy. Proc. Second World Conference,* Cambridge.

Johnsen, S.J., Clausen, H.B., Dansgaard, W., Gundestrup, N.S., Hammer, C.U., Andersen, U., Anderson, K.K., Hvidberg, C.S., Dahl-Jensen D., Steffensen, J.P., Shoji, H., Sveinbjornsdottir, A.E., White, J., Jouzel, J., and Fisher, D. (1997) The $\delta^{18}O$ record along the Greenland Ice Core Project deep ice core and the problem of possible Eemian climatic instability. *J. of Geophysical Research* **102(C12)**, 26397–26410.

Kalma, J.D. and Calder, I.R. (1994) Land Surface Processes in Large-Scale Hydrology, World Meteorological Organisation. Operational Hydrology Report No. 40.

Kabat, P., Hutjes, R.W.A. and Feddes, R.A. (1997) The scaling characteristics of soil parameters: From plot scale heterogeneity to subgrid parameterization. *J. Hydrology* **190**, 363–396.

Kaczmarek, Z. with Arnell, N.W. and Stakhiv, E.Z., Hanaki, K., Mailu, G.M., Somlyody, L., Strzepek, K., and other contributing authors (1996) Water resources management. *In* T. Watson, C.M. Zinyowera, R.H. Moss, and D.J. Dokken (eds.), *Climate Change 1995. Impacts, Adaptations and Mitigation of Climate Change: Scientific-Technical Analyses,* Cambridge U.P., pp. 325–363.

Kress, V. (1997) Magma mixing as a source for Pinatubo sulphur. *Nature* **389**, 591–593.

Lamb, P.J. and Pepper, R.A. (1987) North Atlantic oscillation: Concept and an application. *Bull. Amer. Met. Soc.*, IL, 1218–1225.

Lean, J., Bunton, C.B., Nobre, C.A., and Rowntree, P.R. (1996) The simulated impact of Amazonian deforestation on climate using measured ABRACOS vegetation characteristics. *In* J.H.C. Gash, C.A. Nobre, J.M. Roberts, and R.L. Victoria (eds.), *Amazonian Deforestation and Climate,* Wiley, Chichester, pp. 549–576.

Leavesley, G.H. (1994) Modeling the effects of climate change on water resources—A review. *Climate Change* **28**, 157–177.

——. (1998) Overview of models for the evaluation of the impacts of climate change on hydrology. *In* J.C. van Dam (ed.), *Impact of Climate Change and Climate Variability on Hydrologic Regimes,* Cambridge U. P.—UNESCO International Hydrology Series (in press).

Lehman, S. (1997) Sudden end of an interglacial. *Nature* **390**, 117–119.

Lins, H.F., Wolock, D.M. and J., Mccabe G.J. (1997) Scale and modeling issues in water resources planning. *Climatic Change* **37**, 63–88.

Loaiciga, H.A., Valdes, J.B., Vogel, R., Garvey, J., and Schwartz H. (1996) Global warming and the hydrologic cycle. *J. Hydrology* **174**, 83–127.

Manabe, S. (1969) Climate and the ocean circulation: I, The atmospheric circulation and the hydrology of the Earth's surface. *Mon. Wea. Rev.* **7**, 739–774.

McCartney, M. (1997) Is the ocean at the helm? *Nature* **388**, 521–522.

McCormick, P.M., Thomason, L.W., and Trepte C.R. (1995) Atmospheric effects of the Mt. Pinatubo eruption. *Nature* **373**, 399–404.

McDonnell, J.J., Freer, J., Hooper, R., Kendall, C., Burns, D., Beven, K., and Peters, J. (1996) New method developed for studying flow on hillslopes. *EOS, Transactions, American Geophysical Union* 77(47), 465–472.

Michaud, J.D. and Shuttleworth, J.W. (1997) Executive summary of the Tucson aggregation workshop. *J. of Hydrology* **190**, 176–181.

Mitchell, J.F.B., Johns, T.C., Gregory, J.M., and Tett, S.F.B. (1995) Climate response to increasing levels of greenhouse gases and sulphate aerosols. *Nature* **376**, 501–504.

Nemec, J. (1994) Climate variability, hydrology and water resources: do we communicate in the field? (HAPEX, FIFE, GEWEX, GCIP and what next?). *Hydrol. Sci. J.* **39**, 193–197.

Nepstad, D.C., De Carvalho, C.R., Davidson, E.A., Jipp, P.H., Lefebvre, P.A., Negreiros, G.H., Da Silva, E.D., Stone, T.A., Trumbore, S.E., and Vieira, S. (1994) The role of deep roots in the hydrological and carbon cycles of Amazonian forests and pastures. *Nature* **372**, 666–669.

Oldfield, F. (1997) Forward to the past—An update on PAGES. *IGBP Newsletter* **31**, 2–3.

Pearce, F. (1993) Pinatubo points to vulnerable climate. *New Scientist* 19 June 1993.

Philander, S.G.H. (1990) *El Niño, La Lina and the Southern Oscillation.* Academic Press, New York.

Rind, D., Rosenzweig, C. and Goldberg, R. (1992) Modeling the hydrological cycle in assessments of climate change. *Nature* **358**, 119–122.

Schultz, G.A., Hornbogen, M., Viterbo, P., and Noilhan, J. (1995) *Coupling Large-Scale Hydrological and Atmospheric Models,* International Association of Hydrological Sciences (IAHS), Special Publication No. 3, IAHS Press, Wallingford, 96 pp.

Sellers, P.J. (1987) Modeling effects of vegetation on climate. *In The Geophysiology of Amazonia—Vegetation and Climate Interactions,* Wiley/UNU, pp. 297–344.

Shiklomanov, I.A. (1998) Climate change hydrology and water resources the work of the IPCC, 1988–1994. *In* J.C. van Dam (ed.), *Impacts of Climate Change and Climate Variability on Hydrological Regimes.*

Stocker, T.F. (1998) A glimpse of the glacial. *Nature* **39**, 338–339.

Stone, R.C., Hammer, G.L. and Marcussen, T. (1996) Prediction of global rainfall probabilities using phases of The Southern Oscillation Index. *Nature* **384**, 252–255.

Street-Perrott, F.A. (1994) Drowned trees record dry spells. *Nature* **369**, 518.

Sutton, R.T. and Allen, M.R. (1997) Decadal predictability of North Atlantic sea surface temperature and climate. *Nature* **388**, 563–567.

Thomas, G. and Henderson-Sellers, A. (1991) An evaluation of proposed representations of subgrid hydrologic processes in climate models. *J. of Climate* **4**, 898–910.

Van Dam, J.C. (1998) *Impact of Climate Change and Climate Variability on Hydrologic Regimes. Cambridge University Press—UNESCO International Hydrology Series* (in press).

Vorosmarty, C.J., Gutowsky, W.J., Person, M., Chen, T.C., and Chase, D. (1993) Linked atmosphere-hydrology models at the macroscale. *In Macroscale Modeling of the Atmosphere* (Proc. of the Yokohama Symp.) IASH Publ. No. **214**, IASH, Wallingford, U.K., pp. 3–27.

Vorosmarty, C.J., Bonell, M. and Becker, A. (1998) Hynest. *In* T. Dunne and A. Becker (eds.), *Opportunities for Hydrologic Research in the Humid Tropics with Specific Reference to the Amazon Basin,* UNESCO IHP (Paris)/IGBP-BAHC (Potsdam) (in press). (Developed from *The Hydrological Component of the Amazon Experiment LBA—Draft Science Plan,* based on the IGBP(BAHC)-UNESCO IHP Workshop, Piracicaba, Brazil, 30 November–3 December 1994.)

Ward, R.C. and Robinson, M. (1990) *Principles of Hydrology.* McGraw-Hill, Maidenhead, U.K.

Watson, T., Zinyowera, C.M., Moss, R.H., and Dokken, D.J. (1995) *Climate Change, Impacts, Adaptations and Mitigation of Climate Change: Scientific-Technical Analyses.* Cambridge University Press.

White, J.W.C. (1993) Don't touch that dial. *Nature* **364**, 186.

Wood, E.F. (1995) Heterogeneity and scaling land-atmosphere water and energy fluxes in climate systems. *In* R.A. Feddes, (ed.), *Space and Time Scale Variability and Interdependencies in Hydrological Processes,* UNESCO-Cambridge University Press, Cambridge, pp. 3–19.

——. (1997) Effects of soil moisture aggregation on surface evaporative fluxes. *J. of Hydrology* **190**, 397–412.

Wood, E.F., Sivalapan, M. and Beven, K. 1990. Similarity and scale in catchment storm response. *Rev. Geophysics* **28**, 1–18.

Wood, E.F., Lettenmaier, D.P. and Zartarian, V.G. (1992) A land-surface hydrology parameterization with subgrid variability for general circulation models. *J. of Geophysical Research* **97**(D3), 2717–2728.

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