PLEISTOCENE CLIMATES

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ABSTRACT

Palaeoclimatology includes data gathering and analysis, as well as synthesis and interpretation. Although speculative evaluations persist, the entire data base for objective synoptic reconstructions in palaeoclimatology has been greatly enhanced during the past several decades. It is now possible to delineate the trend of planetary climates during the Tertiary with a degree of temporal precision and in increasingly quantitative terms. The descriptive model of mid-Tertiary climate so derived differs conspicuously from that of the Holocene interglacial and Pleistocene glacials. Significant and repeated oscillations of climate already marked the Pliocene and assume ecological prominence during the early Pleistocene, with full-scale glacial-interglacial cycles established by 700,000 B.P. The nature of these cycles is discussed, and it is proposed that the term "glacial" be restricted to times of continental glaciation on Europe and North America. The Last Interglacial and Last Glacial, so defined, date from about 125,000 to 75,000 and about 75,000 to 11,000 B.P., respectively. The temporal evolution and spatial patterning of these episodes are described and analyzed, and an explanatory model is presented for the climatic anomalies of the Last Glacial. Finally, the differing wavelengths of first- to sixth-order climatic changes are discussed briefly in terms of potential causative factors.

BASIC PRINCIPLES

Palaeoclimatic research is a cross-disciplinary endeavor that draws its data from a wide array of specialized subdisciplines, whose roots rest in a number of the physical and biological sciences. Ideally such research should proceed in several steps, from specific data analysis and basic evaluation, to thematic or regional collation and synthesis and, ultimately, to higher-level interpretation. Such an approach can be taken either from a past or a contemporary starting point, reflecting primarily on a geological or palaeobiological methodology in the first case, on a meterological vantage point in the second. In either approach, attempts at general interpretation involve implicit or explicit model building that is or should be properly anchored in historical or actualistic experience on the one hand, as well as be compatible with the factual data on the other.

Perhaps the fascination of speculative probings into the unknown can be called upon to excuse the high proportion of irresponsible writings in what may be called palaeoclimatology. If the implications of such a statement seem too harsh, the issue can be rephrased: considering the great amount of often tedious work that goes into the collecting of palaeoclimatic raw data and their basic evaluation, the real effort expended on large-scale interpretations has been relatively small. Such a polarization helps to underscore the patient, ongoing search for clues to the history of the planet that, after some two centuries, has provided little more than qualitative, fragmentary, crudely dated, and often equivocal data on past climates. At its worst, the spate of palaeoclimatic theorization that continues unabated since the 1860s has deluded workers with a chimera of order and, at its best, has stimulated new or redoubled efforts in different subdisciplines or regions. We are still in a data-gathering stage. However, progress in gathering and evaluation of historical and actualistic data has made such strides in the past decade that tangible temporal and spatial patterns are now emerging in the historical sphere, while a far more effective processual understanding of atmospheric dynamics has begun to enhance the possibilities of palaeometeorological research. This increasingly optimistic situation, coupled with the availability of computer technology, has already allowed attempts to simulate Pleistocene circulation anomalies. It now appears to be only a matter of a few years before significant progress can be expected in general palaeoclimatology and, by implication, in deciphering the ultimate causes of climatic change.

In the meanwhile it is imperative for earth scientists and palaeobiologists to continue studies that will produce synoptic pictures of characteristic past climates in space and of changing zonations through time. Only then may we hope for an adequate data base for future palaeometeorological models.

The present paper attempts a synopsis of our current understanding concerning Pleistocene palaeoclimates. It focuses on the zonal patterns that can be inferred from the best-known sectors of the Old World, thus complementing the emphasis of Davis and Wright (both this volume) on the New. It attempts an integration of the terrestrial and oceanic records. And above all it is selective and presented as a personal essay, rather than as a detailed documentation.

As a matter of convenience, contemporary climatic zonation has been traditionally represented cartographically by selected climatic parameters or by arbitrary but synthetic "types." The most elaborate of the parameters available for ready illustration are radiation, temperature, evaporation, precipitation, pressure, and wind belts. These fundamental properties give but minimal expression to the underlying dynamism of climate, specifically to heat and moisture flux, and to the three-dimensional nature of the atmosphere. As a result, such graphic representations, including the standard climatic classifications of W. Köppen and others, should be regarded as little more than heuristic devices to illustrate the basic patterns of climatic organization and distribution on the continents. As is the case with other descriptive models, such world maps or generalizations are useful to convey patterns and associations, but are of limited usefulness in characterizing component parts, such as the climatology of a particular area.

Figure 1b attempts a descriptive generalization of Holocene, *i.e.,* interglacial, climate for a composite, idealized landmass. Following the prototype of G. T. Trewartha (1968), and ultimately Köppen, it shows the zonation of the arid and semiarid climates, of the frost-free tropics, and of those key gradations of the temperate and arctic zones that implicitly express the length and intensity of the vegetative period. The high degree of generalization of this model and its emphasis on simple parameters of temperature, relative moisture, and growing season make it particularly suitable to apply to typical categories of palaeoclimatic evidence. At the same time, its very simplicity is advantageous when attempting similar descriptive models for Pleistocene glacial climates on the one hand (Fig. 1c), and Tertiary non-glacial climates on the other (Fig. 1a).

The underlying rationale for planetary distributions of climate is to be sought in solar radiation, in the shape and rotation of the planet, and in the geometry of its orbit, particularly the angle of the ecliptic. Within this context, zonation is more specifically related to (a) radiation received, as controlled or modified by latitude, altitude, and albedo; (b) atmospheric water; (c) pressure and wind belts; and (d) the distribution of land and water. The details are subsumed into the dynamics of the general circulation of the atmosphere, a powerful set of forces put in motion by the indispensable latitudinal heat exchange and moisture flux. This heat engine has been in operation in the past as much as it is today. Yet the total spectrum of solar radiation, the orbital parameters of Earth's axis and orbit, the shape and arrangement of the land masses, atmospheric water, as well as albedo, are each variable, whether independently or dependently. Therefore it comes as no surprise that the heat engine is capable and culpable of a great variety of perturbatious, with a wide range of amplitudes and wavelengths (Flohn, 1973b).

Although the intricate functional details of the energy system are as yet imperfectly understood, the rudimentary concept of the heat balance remains the most fundamental mechanism to consider in any palaeoclimatic evaluation. Thus the contemporary (interglacial) circulation model finds symptomatic expression in latitudinal temperature gradients that reflect strongly on polar ice packs and ice sheets; at the same time, these gradients show marked asymmetry between summer and winter hemispheres. The

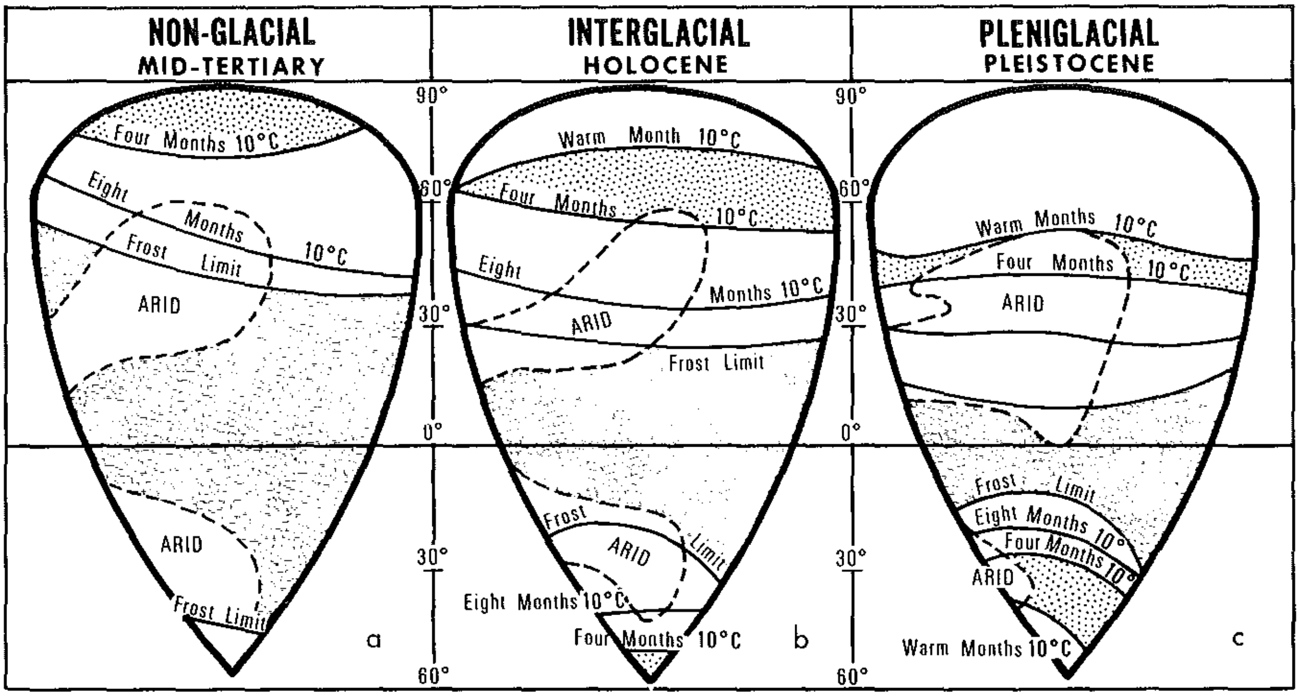

FIGURE 1. Descriptive models of (a) non-glacial, (b) interglacial, and (c) plenigacial climatic zonation. The holocene pattern is modified after Trewartha (1968).

net heat loss of ice-mantled polar regions is primarily responsible for the intensity of contemporary zonal gradients and for the contrast of summer and winter circulation patterns. Potential deviations from this thermally controlled model continue to offer the most productive palaeometeorological insights into glacial as well as non-glacial circulations.

TERTIARY CLIMATES

Geological and palaeobiological studies during the last century have served increasingly to demonstrate that the Tertiary was characterized by remarkably warm climates in high latitutdes. In particular, since the 1920s, it has been possible to debunk a variety of lower or middle latitude "tillites," while at the same time, closer studies of the fossil floras and faunas provided semiquantitative estimates for the positive thermal deviations involved (Schwarzbach, 1963).

Specifically, the evidence now suggests (a) remarkably warm conditions for the early to mid-Tertiary, particularly the Eocene, and (b) a general cooling trend in middle and high latitudes during the later Tertiary, as a prelude to the Pleistocene. Initially a warm-temperature Arcto-Tertiary forest, with *Sequoia,* pine, fir, spruce, willow, birch, and elm, appears to have dominated the planetary belt 50° to 70° N, while a subtropical evergreen forest, with the *Nipa* palm and numerous tropical oaks, is indicated between 25° and 50° N, with coral reefs at the same latitudes (Axelrod, 1960; Frakes and Kemp, 1972). Molluscan assemblages from the west coast of North America suggest Eocene winter water temperatures as much as 15° C warmer than today; floras from the western United States infer mean temperatures 20° C, from western Europe 12° C, with southern Chile 7° C warmer than now (Fig. 2). Semiarid floras are hinted at along the western littorals of lower middle latitudes (Axelrod, 1960), and increasing continentality, with drier conditions, is evident in the lee of mountain ranges developing during the Alpine Orogeny. Temperatures trended downward in higher latitudes beginning in early Oligocene times, increasingly so by the late Miocene (Fig. 2).

This picture of warm Tertiary climates in middle latitudes is complemented by an increasingly substantial body of evidence that the polar regions were devoid of ice sheets and permanent pack or shelf ice

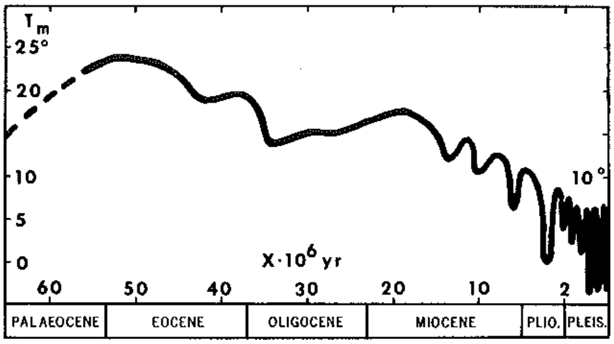

FIGURE 2. Schematic representation of Cenozoic, mid-latitude temperature trends. Data based on Schwarzbach (1963, 1968), Nairn (1964), Frenzel (1968), Hamilton (1968), Frakes and Kemp (1972), Mercer (1973), and others. Time scale after van Couvering (1972). Note scale distortion for Pliocene and Pleistocene.

until the late Miocene or Pliocene. At times there were restricted mountain glaciers of temperate type, for example, in Antarctica during the mid-Eocene and early Oligocene and again since the late Miocene, and in Alaska during the late Miocene. It was not until the Pliocene that the southern oceans had cooled significantly and ice sheets formed on Antarctica, a continental plate that had hovered in polar latitudes throughout the Tertiary (Mercer, 1973). Large-scale glaciations of Iceland, the Andean piedmont in Argentina, and the Sierra Nevada of California are variously dated between 3.5 million and 3.0 million years (Mercer, 1973), implying an initial phase of drastic worldwide cooling, comparable to that of one of the Pleistocene glacials.

The non-glacial (acryogenic) climate of the early and Middle Tertiary has been described as the "normal climate of geological time." The generalization retains its validity, complemented as it now is by an array of qualitative and semiquantitative indicators for a clear latitudinal zonation of climate (Fig. 1a). This zonation paralleled the modern latitudinal grid, in view of the insignificant meridional "drift" of continental plates during the Cenozoic. At the same time, there were significant departures, both in terms of meridional temperature gradients and degree of continentality of the existing landmasses. The limited amount of snow and ice within the polar circles substantially reduces albedo and winter cooling, and the Antarctic albedo may have been cut by half (Flohn, 1973a). Therefore it is possible that the meridional gradient of the winter hemisphere between 0° and 90° latitude might have been comparable to that between 0° and 40° today. In effect, a (modern) summer-type circulation must have applied to the Tertiary winter hemispheres. Winter airmass contrasts would have been small, and westerly depressions shallow and slow-moving, as well as restricted to a relatively weak thermal and pressure cline at somewhat higher latitudes. If the meridional temperature gradient of the middle troposphere were reduced by 50%, it is probable that the latitude of the subtropical anticyclones would shift 10° poleward (Flohn, 1969, Fig. 6), displacing the core of the westerlies to the polar circles. Consequently, the primary palaeometeorological departures must be visualized for the winter hemispheres, although summers would also have been substantially warmer in middle and higher latitudes, with minimal short-term thermal variability.

The non-glacial climatic model developed here for the Tertiary (Fig. 1a) illustrates the anomalous nature of both glacial *and* interglacial climates in Pleistocene times. Additionally it provides an indispensable zero point from which to evaluate and interpret the "climatic deterioration" of the late Miocene and Pliocene, as well as the unusual amplitude and wavelength of the violent and rapid oscillations of Pleistocene climate.

THE EARLY AND MIDDLE PLEISTOCENE

As the increasingly positive pulsations of the glaciers indicate, the Pliocene was in every sense a prelude to the Pleistocene Ice Age. In addition to establishing a permanent ice sheet on Antarctica and, possibly, on Greenland, the Pliocene also saw a spasmodic cooling of middle latitudes. This is best exemplified in the incomplete pollen diagrams of the Netherlands Pliocene, and in the successive decimation of tropical, evergreen-broadleaved trees from the accompanying macrobotanical record (van der Hammen *et al.*, 1971). It is precisely this unstable, "transitional" character of the Pleistocene that makes a satisfactory definition of the Plio-Pleistocene boundary an increasingly difficult and arbitrary proposition. A decision of the 1948 International Geological Congress to relate this boundary to the point of dramatic cooling in the marine Calabrian sequence of Italy has been of little help, since external correlations have only been possible on rare occasions via micropalaeontological arguments (*e.g.*, Poag, 1972), while the number of late Cenozoic cold-climate oscillations has proven to be hopelessly complex. The proposed Calabrian Plio-Pleistocene boundary can be applied as a useful marker only if it is tied in with the potassium-argon-dated palaeomagnetic stratigraphy. Therefore it becomes an expediential chronostratigraphic marker, perhaps fortuitously linked to a minor geomagnetic reversal, such as the Olduvai event of about 1.8 million years ago.

Whatever definition of the Plio-Pleistocene boundary is ultimately adopted as a matter of convenience, very few chronostratigraphic sequences that carry a significant palaeomagnetic record can be accurately calibrated with one another at this time. Oxygen-isotopic evaluation of deep-sea cores has not been completed in sufficient detail for the early Pleistocene, and the "standard" Netherlands sequence shows palaemagnetic gaps or inconsistencies that require reinvestigation (Brunnacker, 1975). Perhaps the longest, and most complete, continental record presently available for the early and middle Pleistocene has been studied —including detailed magnetostratigraphy—in the relatively dry lowlands of Czechoslovakia and eastern Austria (Kukla, 1975). Here eolian loess, colluvium, soil horizons, and ecologically sensitive snail assemblages prove the existence—within the last 700,000 years—of eight complete cycles of dry, cold-climate (glacial) loess, alternating with biochemical weathering as inferred from (interglacial) paleosols. Including the Krems profile of Lower Austria, it further appears that a minimum of eight additional loess cycles span the preceding million years of the early Pleistocene back to the Olduvai geomagnetic event. This picture of complexity matches that of the deep-sea record and serves to illustrate the degree to which other sequences, such as those glacial episodes based on continental tills, are hopelessly incomplete and therefore unsuitable for Pleistocene subdivision (Butzer, 1974).

Although the details of the early Pleistocene remain vague, it is now apparent that middle and higher latitudes were subjected to repeated and substantial oscillations of climate between the Olduvai magnetic event of 1.8 million years and the Matuyama-Brunhes magnetic reversal of 0.7 million years: (a) The central European loess cycles verify at least eight periods of dry, cold climate (Kukla, 1975). (b) The alluvial terraces of the Rhine River record at least five phases of higher competence aggradation, with ice-rafted debris (indicative of at least severe winter freezing of the Rhine and its affluents) first appearing in a cold cycle beginning shortly after 900,000 years, and syngenetic ice-wedge casts (indicative of permafrost and temperatures 17° C colder than today) first present in deposits dating about 700,000 years (Brunnacker, 1975). (c) The palaeobotanical record of the Netherlands shows three phases of open, tundra-like vegetation and several additional episodes of less severe cold, with boreal woodland, each leading to further eradication of Tertiary floral elements (van der Hammen *et al.,* 1971). (d) The complex record of the Massif Central, France, includes five episodes of periglacial activity (severe soil frost) (Bout, 1975), as well as repeated vegetation change (van der Hammen *et al.,* 1971). (e) The oxygen-isotopic record of the deep seas (reflecting changing ocean isotopic composition, related preeminently to the growth and dissipation of northern hemisphere ice sheets) shows cyclic oscillations of "glacial-interglacial" amplitude extending back almost to 800,000 years (Shackleton, 1975).

In effect, an uncertain but substantial number of cold-warm cycles can be recognized during the early Pleistocene, although the severity of the cold intervals did not match that of glacial phases during the middle and later phases of the Pleistocene. Each of the cold hemicycles created a subarctic or boreal vegetation in western Europe, while providing a cold, steppic environment suitable for loess accumulation in east-central Europe. The deep-sea record indicates extensive northern hemisphere glaciation for the later of these cold phases. In general, the available evidence can be interpreted with recourse to repeated cold impulses, characterized by a progressive intensification of climatic stress.

Turning to the middle and late Pleistocene, the European interglacial pollen horizons and the Rhine alluvial terraces indicate a minimum of six, and the Czechoslovakian loess, a total of eight cold-warm cycles since the Brunhes-Matuyama geomagnetic reversal of 700,000 years. Because eight comparable cycles are also indicated in both the Pacific and Atlantic deep-sea cores, this must be seen as evidence for eight glacial intervals of hemispheric or global significance during the past 700,000 years. At least five of the glacials since 500,000 B.P. were sufficiently severe to produce permafrost conditions in mid-latitude Europe. Long pollen records, also indicative of repeated, major vegetation shifts in the Mediterranean Basin (Macedonia) and in tropical South America (Colombia) (van der Hammen *et al.,* 1971), are available also and, despite certain problems of resolution and inferred correlations, verify the geomorphologic evidence for significant environmental changes in lower latitudes (Butzer, 1974). Yet, a word of caution must be interjected—the climatic cycles of higher latitudes are presently of little or no value in analyzing the mid-Pleistocene records of tropical continents.

The deep-sea record, as supported by various other lines of evidence, identifies ten, full-scale, glacial-interglacial cycles, as opposed to the four, "classic" glaciations, during the last 800,000 years. To what extent were these oscillations truly "cyclic"? The amplitude of variation between maximum glaciation and maximum deglaciation differed in detail, but remained basically comparable through time, suggesting some fundamental constraints to the planetary glacial-interglacial pendulum. However, there was no strict periodicity, and the wave-length of the superimposed fluctuations expressed by Shackleton's (1975) stage units is by no means constant. Those "cold" stages of significant glacier expansion, that can be identified prior to the late Pleistocene, had a variable duration of 18,000 to 67,000 years, compared with 23,000 to 73,000 years for the "warm," essentially deglacial stages—yet with only 9000 to 35,000 years' duration in terms of standard interglacial pollen horizons. Each glacial and interglacial hemicycle was different in its detail, so that one cannot speak of repetitive climatic events. Finally, correlation of Shackleton's multiple deep-sea units with existing, continental time-stratigraphic schemes is next to impossible, although cautious comparisons with detailed lithostratigraphic sequences, defined with some measure of radiometric control, appear to be possible eventually.

THE LATE PLEISTOCENE

The nature of late Pleistocene climatic trends and oscillations is presently known in considerable detail, thanks largely to the study of semicontinuous, long profiles of marine or terrestrial origin. Until the 1940s glacials and interglacials were crudely defined by successive till sheets and correlated alluvial terraces, alternating with lake and bog deposits or interglacial paleosols. The first refinements of loess stratigraphies at this point allowed for better linking of tills and alluvial units, and an increasingly elaborate subdivision, based primarily on paleosols of interglacial or interstadial type. By and large, interglacials remained episodes of non-deposition or downcutting and weathering. Palynological resolution of localized interglacial deposits yielded the only longer temporal sequences to juxtapose with seemingly much more protracted intervals of glacial accumulation.

During the decade 1955 to 1965, the advent of detailed C^{14} dating opened a host of controversies about the actual palaeoclimatic "events" of the past 75,000 to 100,000 years. At the same time, study of deep-sea cores began to reveal the inadequacies of even the best till, loess, and alluvial sequences in recording the actual train of events, because of their incompleteness or their temporal distortion. However, time and again, interpretation of the deep-sea cores proved equally premature, as problems of evaluation and dating remained as serious as in the realm of terrestrial records. Fortunately a few palaeomagnetic markers now served to calibrate these long records more objectively, and cautious palaeoecological evaluations can be extracted from a broad range of specialized investigations. The resulting data, with a temporal confidence limit of ± 1000 at 10,000 years and ± 5000 at 75,000 years, can now be broadly compared—internally as well as with the major trends inferred from glaciological and terrestrial stratigraphies.

Turning to the 130,000-year time span of the late Pleistocene and Holocene, several approaches can be singled out and briefly summarized.

1. Oxygen isotopic determinations of planktonic (and benthonic) foraminifera show fluctuations resulting almost entirely from changes in ocean isotopic composition, which reflect on growth and wastage of northern hemisphere ice sheets (Shackleton, 1975). The resulting curves of global ice volume show non-glacial conditions about 128,000 to 75,000 years ("last interglacial") and again since about 13,000 years ("postglacial"); partial deglaciation characterized an intermediate interval about 64,000 to 32,000 B.P. ("interstadial" or "interpleniglacial"). Maximum deglaciation was achieved approximately 125,000 B.P. and again about 5000 B.P. The minor fluctuations superimposed on this general curve are relatively small, suggesting a considerable inertia of the key glaciers.

2. Quantitative palaeoenvironmental analyses of foraminiferal assemblages reveal more rapid and complex fluctuations, which were related primarily to summer and winter temperatures, as well as to the salinity of ocean surface waters (Imbrie *et al.*, 1973). In detail, one such critical curve shows warm conditions in the North Atlantic (52°35' N) about 127,000 to 109,000 B.P., with fluctuating but cooler conditions about 109,000 to 73,000 B.P., and maximum cold about 73,000 to 11,000 B.P., interrupted by three warmer intervals that are centered some time near 59,000, 48,000, and 31,000 B.P. (Sancetta *et al.*, 1973). However, deglacial warming was

not general and synchronous, and micro-correlations of deep-sea faunal horizons through an ash marker horizon show that polar waters retreated only gradually from the North Atlantic (Ruddiman and McIntyre, 1973). This glacial termination shifts from about 13,500 B.P. off the western coast of Britain, to 6500 B.P. off Greenland, demonstrating a 7000-year span of deglacial warming, comparable to that of the dated recession of the continental glaciers and to the well-known European vegetation succession. Overall, the faunal data suggest that the swing of the climatic pendulum was considerably more violent and rapid than was the response of the northern hemisphere glaciers. At the same time, average climatic anomalies of broad latitudinal zones were not identical, with higher latitudes cooling first and warming last, and being affected by substantially greater deviations than were the lower latitudes.

3. The deep-sea record is complemented by staple isotope investigation of the Greenland ice sheet, where periodic deviations of O^{18} are mainly determined by the temperature of formation of the precipitation that feeds the glacier (Johnsen *et al.*, 1972). The resulting time-calibrated curve gives an intriguing record of atmospheric temperature change over the northern North Atlantic that in general resembles the deep-sea oxygen isotopic record. This curve, which is paralleled by data from the Antarctic ice sheet, can be translated very approximately into a representation of changing temperature gradients between the sub-arctic and tropical North Atlantic—*if* it is assumed that air temperatures over the equator remained essentially unchanged (Fig. 3). The general

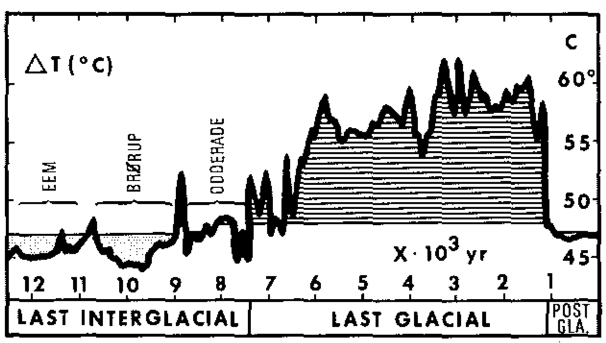

FIGURE 3. Temperature gradients over the North Atlantic Basin since 125,000 B.P., assuming only limited air temperature deviations at the equator. Data based on Johnsen *et al.* (1972). (Corrected time scale, but distortion probable prior to 75,000 B.P.)

circulation pattern of the North Atlantic about 125,000 to 90,000 B.P. was evidently much like that of the present. Then it underwent violent oscillations until about 63,000 B.P., and subsequently maintained a glacial pattern until nearly 12,000 B.P., with intermediate-level, high-latitude warmups centered at approximately 56,000 and 38,000 B.P. (Fig. 3).

These three lines of evidence provide the most authoritative data presently available for subdividing the late Pleistocene. Obvious subdivisions are the Last Interglacial, about 127,000 to 73,000 B.P.; the Last Glacial, about 73,000 to 11,000 B.P.; and the Postglacial or Holocene, since 11,000 B.P. Correlation of these units with continental terminologies is more problematical.

In particular, the Last Interglacial in Europe is widely designated as the Eemian, named after a marine transgression in the Low Countries, but more commonly identified from characteristic pollen diagrams. However, varvites associated with the Eemian pollen succession yield a total duration of only 9000 to 11,000 years (Turner, 1975) for the associated span of warm to temperate woodland vegetation. This indicates that the Eemian is coincident with only the warmest part of the non-glacial, deep-sea isotopic stage 5 (namely, 5e). At the same time, the palynologically defined interstadial horizons Amersfoort, Brörup, and Odderade—which follow the Early Würm glacial of van der Hammen *et al.* (1971) and were once thought to be younger than 75,000 B.P.—now appear to date about 115,000 to 75,000 B.P. Sancetta *et al.* (1973) have taken this situation to define the Early Würm chronologically between 109,000 and 73,000 B.P., arguing that this period was already decidedly cooler (in mid-latitudes) than the present.

The Last Interglacial in the Mediterranean Basin, and by extension along other lower-latitude coasts, traditionally has been defined with reference to certain sea-level stages and their associated faunas. As currently understood, some of these horizons (including such terms as Eutyrrhenian and Neotyrrhenian, or Monastirian I and II, or alternatively, II and III) are considerably older (180,000–220,000 B.P.), whereas the youngest date from three stages between 125,000 and 75,000 B.P. (Butzer, 1975a, b). A comparable tripartite subdivision of similar age (Barbados I, II, III) is being more and more widely accepted in other regions. These relatively high sea levels represent a "warm" Mediterranean sea and predate the regressional eolianites (and related loess-like colluvial sediments) widely attributed to the onset of the Würm glacial-eustatic regression (Butzer, 1975a; Brunnacker and Ložek, 1969).

Thus, logically, the glacial boundary would be drawn just after 75,000 B.P. on the basis of the littoral stratigraphies. In palynological terms, the high sea levels of 125,000 to 75,000 B.P. coincided approximately with a long episode of warm-temperate environment (the combined Pangaion, Drama, and Elevtheroupolis stages of van der Hammen *et al.,* 1971), interrupted by only two brief dry-cool intervals, at sites in southern Spain and Greece (discussion, Butzer, 1975a, with references).

In other words, the definition of the Last Interglacial and Last Glacial advocated above (Fig. 3) is both reasonable and desirable. However, the designation Eemian for the Last Interglacial is inappropriate, since the mid-latitude terrestrial record carries insufficient information on the 35,000 to 40,000 years of post-Eemian time that preceded the inception of continental glaciers in Scandinavia and Canada. Because the terms "Würmian" and "Wisconsinan" were coined with reference to glacial conditions, their use should also be restricted to true glacial episodes. Present evidence suggests that although much of the period about 115,000 to 75,000 B.P. was cooler than today in mid-latitudes, with moderate expansion of highland glaciers around the world, deep-sea oxygen-isotopic as well as glacial-eustatic records show that there was no ice over Sweden, Québec, or Keewatin. Thus the Brörup (ca. 100,000 B.P.) and Odderade (ca. 80,000 B.P.) can hardly be called interstadials but instead must be considered as submaxima of the Last Interglacial.

CLIMATIC PATTERNS OF THE LAST INTERGLACIAL

The latest is the best understood of all the Pleistocene interglacials. Although it cannot be considered a model for all Pleistocene interglacials, it can be used as a representative example. In the most general of terms, it is true that at its warmest, the Last Interglacial was similar to the Holocene. Yet when examined regionally and temporally, the Last Interglacial was complex indeed.

Mid-Latitude Europe

The mid-latitude pollen sequence in Europe begins with a vegetation mosaic of birch/pine woodland and grassland on unleached, calcareous loess or till; as warming proceeded, plant migrations and soil development continued, and a mixed oak-pine-elm forest succeeded on the high base-status, brown forest soils. Subsequently, as soil leaching progressed and base-status declined, a forest of hornbeam, fir, and oak followed; ultimately climatic cooling favored increasingly oligotrophic conditions, with podsolic soils and a vegetation of pine and birch, interspersed with tracts of acid moors and heath (Turner, 1975). At its "optimum," the 10,000-year span of warm-temperate, mixed-deciduous woodland was at least 2° C warmer than today, as evidenced by anomalous plant distributions and other biological indicators from the North Sea through to western Siberia (Butzer, 1971, p. 380ff.). It appears that the ensuing submaxima of the Last Interglacial, namely the Brörup and Odderade, were characterized by cold-temperate woodlands of spruce, pine, larch, and birch or fir in England, the Low Countries, Denmark, and in the wetter, high country of France and central Europe; the lowlands of France and most of central and eastern Europe were covered with a forest-steppe mosaic, dominated by spruce, pine, birch, and fir (Frenzel, 1968, p. 242ff.; van der Hammen *et al.,* 1971). The intervening cooler episodes saw most of temperate Europe reduced to a forest tundra, steppe, or forest-steppe.

The mid-latitude pollen data are complemented by the palaeosol record. In the now-subhumid loess plains of east-central Europe, there is a unique record of the detailed evolution of the Last Interglacial (Kukla, 1975). The first and maximum warm interval, centered about 125,000 B.P. (Eem), was marked by development of a deep, completely leached, podsolic forest soil. This was followed by an interval of chernozem pedogenesis, then a layer of long-distance eolian silt, and a thick horizon of interbedded loess and reworked soil aggregates ("pellet sands"), which is attributed to erosion of hard, dry soil particles from underneath an incomplete mat of vegetation by torrential rains. This suggests a semiarid climate, with a pronounced dry season, rather than exceptional cold. The second warm interval, at about 100,000 B.P. (Brörup), saw a partly leached, brown forest soil succeeded by a chernozem, and finally a low-humus steppe soil—indicating a return to a mesic environment, but under conditions somewhat drier than during the first of the interglacial maxima. A protracted interval of colluvial reworking followed, until another chernozemic soil formed during the last interglacial maximum at about 80,000 B.P. (Odderade). This was succeeded by an eolian silt, a thick accumulation of "pellet sands," and only then by the first appearance of soil-frost disturbance. These general inferences for Czechoslovakia and eastern Austria are supported by interpretation of molluscan assemblages, as well as by palynological evidence, and are dated by interpolation from geomagnetic marker horizons.

The implications are that extensive areas of central and eastern Europe were covered by grassy steppes or parkland for most of the interval, approximately 115,000 to 75,000 B.P. Europe was evidently somewhat cooler than today but, above all, drier. The composition of the snail faunas suggests that the nature of the climatic shift was from a maritime to a decidedly continental regime, with very cold winters and dry, hot summers; precipitation was probably concentrated during the transitional seasons.

The Mediterranean Region

In the Mediterranean Basin, the three "high" sea levels of 125,000 to 75,000 B.P. (Y1, Y2, Y3) are paralleled by three long episodes of warm-temperate to subtropical woodland in complete pollen cores from southern Spain (Florschütz *et al.,* 1971) and Macedonia (Wijmstra, 1969; van der Hammen *et al.,* 1971), as well as by partial sequences from Catalonia, central Italy, northwestern Greece, and the Levant (discussion, Butzer, 1975a). Floristically these three woodland phases were complex, with perhaps only the first being truly subtropical. They were interrupted only by brief periods of predominantly open vegetation. These interruptions were also cold, with an amplitude of at least 4° C, judging by oxygen isotopic temperatures obtained from a segment of cave stalactite in the Languedoc, which are dated about 97,000 to 92,000 B.P. (Duplessy *et al.,* 1970). Comparison with Figure 3 suggests identification with the brief but marked cold episode apparent in the Greenland isotopic record of about 90,000 B.P., allowing for a reasonable amount of dating imprecision. This interpretation is strengthened by the fact that the cave stalactite record indicates a fairly constant temperature from about 120,000 to 97,000 B.P., much like the Greenland profile.) Interestingly, too, the Y1 and Y2 beaches were associated with a thermophile molluscan fauna now restricted to the Senegal coast of Africa; these mollusca had been exterminated in at least the western Mediterranean Basin before the Y3 substage of approximately 80,000 B.P. (Butzer, 1975). Significantly, the last phase of intensive biochemical weathering, associated with formation of *terra rossa*-type soils, certainly predates the Y3 beaches and appears to be temporally linked to the preceding cool phase.

Despite the increasing evidence for stratigraphic comparability of the European and Mediterranean records, there were significant differences between the areas. In particular, the Czechoslovakian palaeosol record suggests that forests prevailed for only 25% of the 50,000-year span of the Last Interglacial (Kukla, 1975), whereas Macedonia was forested for some 75% of the same time range (Wijmstra, 1969). Therefore the cooling and drying evident in Europe after 115,000 B.P. was restricted largely to mid-latitudes, with temperate mesic conditions continuing to dominate south of latitude 45° N. Unfortunately these tendencies cannot yet be verified in North America, because of the fragmentary nature of the Last Interglacial record there (Butzer, 1971, p. 387ff.). However, what evidence there is does not contradict such a generalization.

Africa

Limited dating possibilities make the African record uncertain for any detailed cross-correlations prior to about 30,000 B.P. It appears that lakes in East Africa, parts of the Sahara, and southern Africa might have been high for parts of the Last Interglacial time span, but this is not demonstrable with any level of confidence. The same applies for various deep palaeosols in the interior. Only on the southernmost Cape coast can broad parallelisms be tentatively recognized with the Mediterranean Basin record (van Zinderen Bakker and Butzer, 1973). This collection of impressions is completed with the suggestion that there is as yet no evidence to link periods of substantially intensified aridity with the Last Interglacial (*e.g.,* Parmenter and Folger, 1974).

Consequently, present understanding of Last Interglacial climatic patterning is restricted to Europe and the Mediterranean borderlands. It would appear that the Eem maximum of the Last Interglacial was basically similar to the Holocene, but that summers were relatively wet in central and eastern Europe and winters were somewhat milder. This maritime trend suggests a greater persistence of zonal circulation patterns during summer and winter than is the case today, possibly linked with a reduced winter snow cover and albedo over Siberia, and a corresponding weakening of the semipermanent Asiatic anticyclone during that season. On the other hand, the Brörup and Odderade maxima were decidedly colder in winter, as well as drier in summer, inferring a greater frequency of meridional circulation patterns, with high-pressure centers over Scandinavia or central Europe. It is probable that winter cyclonic rains were augmented in the Mediterranean Basin and its desert margins. Although valley glaciers and ice caps must have expanded or formed in most of the critical highland areas, it is far from likely that these last two maxima of the Last Interglacial were cold in mid-latitudes (Fig. 3). Only the intervening cold interludes were significant, and a departure of 4° C in the interior of a Mediterranean cave suggests double that value in the open air of northern Europe. But here again the isotopic data from Greenland (Fig. 3) caution that such events were rare, brief, and seldom of comparable amplitude.

Although far from satisfactory, this evidence from Europe and the Mediterranean area serves to illustrate that the 50,000 years of the Last Interglacial saw repeated and significant long-term readjustments of the general atmospheric circulation. The wavelength of the significant trends varied from about 3000 to 10,000 years, and the maximum amplitude was perhaps 50% of that between the Holocene and the last glacial maximum of the Pleistocene. Since these atmospheric anomalies did not permit large-scale glaciation of the northern hemisphere continents, secondary feedback mechanisms were clearly less significant than during a glacial era with established continental glaciers. This shows that the primary atmospheric impulses of the Last Interglacial were surprisingly powerful. But at the same time, it implies that a convergence of multiple factors was necessary to push the global climatic pendulum over from an interglacial to a glacial.

CLIMATIC PATTERNS OF THE LAST GLACIAL

After some strong initial fluctuations apparent in the Greenland isotopic curve (Fig. 3), rapid northern hemisphere glaciation began about 65,000 B.P. Once formed, the Fennoscandian and Laurentide ice sheets remained in place until the 7000-year period of deglaciation that marked the Pleistocene-Holocene transition. Fluctuations of ice volume and of high-latitude temperatures were relatively minor, and the warm-ups were limited to temporary thinning and marginal retreats of the ice sheets. These "interstadials" led to interruptions of loess accumulation and soil-frost activity in mid-latitudes, producing a variety of weakly developed soil horizons. However, no coherent chronology of these interstadials has yet been produced, and critical examination of the different lines of evidence, as dated by different techniques and with varying success, shows a great degree of contradiction. Thus in higher latitudes, it is undesirable to offer strong generalizations about the palaeoclimatic evolution of the Last Glacial.

The apparent discrepancies between different lines of evidence may well be significant and may be the results of differing response levels and trends between ice volume, planetary temperatures at different latitudinal belts, and patterns of atmospheric and oceanic circulation. The fluctuations of North Atlantic meridional heat flux implied by the Greenland ice core (Fig. 3) are of a greater amplitude but shorter wavelength than the deep-sea and continental evidence for glacier fluctuations, but they do correspond closely to the scale of major climatic oscillations in lower latitudes. In other words, the violent pulsations of the general circulation, that began about 95,000 B.P. and accelerated by about 75,000 B.P., continued—despite the inherent stability of continental glaciers, once established—until approximately 11,000 B.P. It is this category of oscillation, with a wavelength of only a few millenia, that characterizes the primary variability of the general circulation, and which, first and foremost, must be adequately explained by any theory of Pleistocene climatic change.

The nature of latitudinal zonation during the Last Glacial can best be illustrated on the basis of several lines of evidence from the key environments.

Mid-latitude Europe

The geomorphic landscapes, vegetation, and climate of the colder episodes of the Last Glacial in mid-latitude Europe have been described in detail elsewhere (Butzer, 1971, chap. 18). In effect, much of northern Europe and the Alps and their foothills were glaciated. Permafrost conditions, implying a mean annual temperature of less than -6° or -8° C (Brown and Péwé, 1973), were found throughout the lowlands north of about 45° N latitude. This implies a temperature depression of at least 16° to 18° C in western, 15° to 17° C in eastern Europe. In central Europe, between the Fennoscandian and Alpine glaciers, average thaw depth was only 70 to 150 cm, about what it is in high arctic regions today. Some form of mixed tundra-steppe, with semidesert shrub (*Artemisia*) and eutrophic grasses, extended across the European plains from the Atlantic coast to central Russia, with no arboreal species other than dwarf willows and birch. Another broad belt of loess steppe, with some tree growth in river valleys and sheltered topographic concavities, reached from east-central Europe to the Black Sea and well into central Asia. Furthermore, steppic grasslands or forest-steppe mosaics reached southward to the shores of the Mediterranean Sea. The polar tree line appears to have extended from southern France eastward to the central Volga Basin. If we assume that this limit coincided with July mean temperatures of 10° to 12° C, then the summer temperature depression was only in the order of 7° to 10° C. This infers that winters were particularly cold, with deviations of 26° C or more.

The Mediterranean Region and Near East

In the Mediterranean Basin, the Last Glacial was both cold and dry. The pollen evidence (Wijmstra, 1969; Florschütz *et al.,* 1971; Butzer, 1975a) basically argues for an open, grassy, *Artemisia*- chenopodiaceae steppe from sea level to 500 m elevation, north of about 35° N latitude; at intermediate elevations of 500 to 1000 m, there was a forest-steppe mosaic, dominated by pine, oak, and beech; at higher levels still, an alpine grassland prevailed. This pattern was maintained with next to no change through the course of the Last Glacial. Only in the Levant (Butzer, 1975b) are there indications that aridity was less accentuated. In view of this incontrovertible evidence for a dry climate, with a dry season perhaps twice as long as today, the geomorphic evidence for snowline depressions, soil-frost activity, and accelerated soil erosion is particularly interesting.

Numerous mountain glaciers argue for a snowline depression of 1200 m for most of the region (Butzer, 1971, chap. 19), which, in view of reduced precipitation, implies temperatures at least 9° C lower than today; fossil solifluction processes of the alpine ecozone imply a similar magnitude of change. These deductions find support in the deep-sea faunal record: surface waters of the Bay of Biscay were at least 12° C colder (J. Imbrie, pers. comm.), those of the eastern Mediterranean 5° to 10° C cooler (Vergnaud-Grazzini and Herman-Rosenberg, 1969).

The pedogenetic record is informative from a different point of view: interdigitations of beach, colluvial, and littoral-eolian deposits show that, in relation to the glacial-eustatic trace of sea levels, each glacial advance in higher latitudes led to accelerated sheetwash activity, soil erosion, and colluviation. This was followed by eolian activity, for dunes often swept far inland, with facies counterparts in eolian silts of coastal or fluvial derivation, ultimately mixed with colluvial deposits (Butzer, 1975a). As geomorphic activity subsided, semiarid soils formed, only to be largely eroded during the next hemicycle of erosion (Brunnacker and Ložek, 1969; Butzer, 1975a). This would suggest specific atmospheric circulation patterns: (a) Intensive, persistent rains during each glacial advance or readvance, despite a prevailing semiarid climate of marked rainfall periodicity; it is probable that such rains were primarily associated with the intensified meridional circulation of autumn and spring. (b) Reduced rainfall amounts, intensity, or duration during each glacial submaximum, increasingly so during the second half of the Last Glacial; very possibly eolian activity was favored by increased dry-weather storminess, *i.e.,* during summer. (c) Limited geomorphic activity during each glacial retreat, allowing for some pedogenesis in a slightly less harsh environment.

These patterns of the Mediterranean Basin recall the alternations of pellet sands, hillwash, or solifluction; eolian loess; and pedogenesis in the central European record (Kukla, 1975). The close stratigraphic and genetic parallels are hardly fortuitous and argue for comparable weather anomalies or even circulation patterns, particularly during the warmer half of the year.

It requires little emphasis that there was no Last-Glacial "pluvial" in the Mediterranean area. Nor was there a significant increase of rainfall at comparable latitudes in the Near East, where closed lake levels, such as those of the Dead and Caspian seas, were higher—solely in response to lower evaporation ratios (review, Butzer, 1975b). Only in the northern and central Saharan desert belt were there more frequent rains, but the climate there remained arid (Butzer, 1971, p. 315ff.). In other words, the secondary impact of a large ice sheet over Europe did not produce a "wet belt" of intensified precipitation in lower middle latitudes. Nonetheless it should not be assumed that the characteristic vegetation of the Mediterranean region did not survive in a plethora of local refuges. This can be deduced from the limited amount of local extinction in the Iberian Peninsula (Florschütz *et al.,* 1971) and from the persistence of limited numbers of various warm-temperate forms within the submontane woodlands of northwestern Greece throughout the vicissitudes of the Last Glacial (Bottema, 1974).

North America

Last-Glacial environmental zonation of North America is considered in detail by Davis (this volume) and Wright (this volume) and has also been reviewed earlier by this author (Butzer, 1971, Chap. 21). Only a few comments need to be made at this point: (a) Permafrost (see also the recent compilation of ice-wedge occurrences by Brown and Péwé, 1973) was restricted to ice proximity and was absent south of the modern 24° C July isotherm. This suggests a temperature depression of 15 to 18° C in the northern United States. (b) The extensive area of closed, cool-temperate forest in the southeastern quadrant of the United States has no parallels in Europe, but may well have had counterparts in the comparable continental situation of Eastern Asia. (c) The high lake levels of the Great Basin (Lakes Bonneville, Lahontan, etc.), coincident with the coolest phases of the Last Glacial, can be adequately explained by lower temperatures and reduced evaporation.

Africa

The continental record of the last 75,000 years in Africa includes a wide range of phenomena, many of them difficult to interpret and few that allow precise information about the amplitude of climatic changes. Exception to this are situations where multivariate parameters can be narrowed or simplified by reasonable assumptions, for example, in closed lake basins, where hydrologic budgets can sometimes be reconstructed. Less satisfactory but also of semiquantitative value are changes in vertical ecozonation or other evidence of specific cold-climate phenomena, including mountain glaciation. Least informative may be changes of the weathering/erosion balance, stream regimen, or spring discharge. Yet each category of information ultimately contributes to filling out the qualitative picture and, in conjunction with radiocarbon dating, is now providing an increasingly detailed impression of the directions and wavelengths of environmental changes during at least the last 20,000 to 30,000 years.

Several key regions on a north-south transect can be briefly outlined:

Egypt: Poorly dated as yet are a number of periods of augmented stream and spring discharge—on the Red Sea coast, in the desert catchment of the Nile, and in the Libyan Desert—that are all older than 25,000 to 30,000 B.P. (Butzer and Hansen, 1968; Butzer, 1975b). Thereafter the period 25,000 to 17,500 B.P. was hyperarid, with increased water flow about 17,500 to 8500 (interrupted by several brief but dramatic interludes of reduced discharge) and again 6000 to 5000 B.P., with a period of biochemical weathering about 7000 B.P. These late-glacial to mid-Holocene moist intervals were modest, however, and they did not compare with the intensity of earlier wet phases that temporarily produced subarid conditions in favored topographic areas of the Sahara.

East Africa and the Southern Sahara: With some short breaks, Nile discharge—as derived from the monsoonal rains of tropical Africa—was greater than today through most of the Last Glacial, until approximately 4500 B.P. In detail the high floods about 24,000 to 18,000 B.P. can be satisfactorily explained by reduced evaporation over highland Ethiopia, while floods about 17,500 to 12,000 B.P. were sufficiently violent to sweep gravel, derived by accelerated local runoff from Sudanese tributaries, right through to Cairo, thereby indicating unusually strong monsoonal rains (Butzer and Hansen, 1968; Butzer *et al.,* 1972). A "wild" Nile, with repeated catastrophic floods of twice the present amplitude, is indicated around 12,000 B.P.

This nilotic record is only partly compatible with the lake record: Chad, Rudolf, Nakuru, and Naivasha were high at some time prior to 20,000 or 30,000 B.P., then low until 12,000 B.P., reaching maximum levels about 10,000 to 7500 and again 6600 to 3000 B.P. (Butzer *et al.,* 1972). A similar history is indicated for Lake Victoria. Computations for the Nakuru-Naivasha basin indicate 50% to 65% increase of precipitation for these early Holocene high lakes. The dry interval between perhaps 30,000 and 12,000 B.P. saw a significant expansion of dune fields along the southern margins of the Sahara, and desert varnish and calcic soils developed in the Rudolf Basin. Yet, it was also coeval with a major vertical depression of the alpine vegetation in Kenya and Ethiopia (Coetzee, 1967; van Zinderen Bakker, pers. comm.), as well as with an advance of the high mountain glaciers. Snowline depressions of 1000 to 1200 m, coeval with a drier climate, infer a temperature lowering of, at the very least, 6.5° to 8° C.

Indian Ocean: It is fortunate that a number of deep-sea cores, off the coast of East Africa, which span several oceanic circulation segments, have been studied (Olausson *et al.,* 1971). There each of the two cool interludes of the Last Interglacial saw the trade-wind circulations weakened for some 2000 years, with a greater expanse of cold, upwelling waters off the Somali coast. The onset of the Last Glacial was almost instantaneous, less than 1000 years, leading to an 8000-year span with increased upwelling. The second submaximum of the Last Glacial was preceded by a 1500-year transition and lasted only 3000 years. Whereas summer temperature deviations (related to weakening of the July southwest monsoon) were greatest during the first pleniglacial phase, winter-cold anomalies (related to weakening of the January northeast monsoon) were preeminent during the later pleniglacial. All but the very last of these cooling trends were surprisingly rapid, evidently in response to sudden, primary changes of the general circulation. The abruptness and brevity of the oceanic circulation shifts documented in these Indian Ocean cores compares well with those of climatic changes in the continental record of lower latitudes.

Southern Africa: The dry zone between the Cape coastal ranges and the northern margins of the Kalahari underwent a different Last-Glacial "history" than did the Sahara or East Africa. In Lunda, northeastern Angola, the interval about 30,000(?) to 13,500 B.P. saw "lateritic" soil development, with high water tables or a vegetation suggesting cooler, moister conditions (Butzer, 1971, p. 342ff.). On the Gaap Escarpment, on the southwestern fringes of the Kalahari, accelerated spring discharge is documented approximately 32,000 to 14,000 and again 9700 to 7600 and about 2500 B.P. (Butzer and R. Stuckenrath, in prep.). The first of these wetter intervals was coeval with flood-plain aggradation of the Vaal River (Butzer *et al.,* 1973) and with a 12-m-deep, nonoutlet lake in the now dry Alexandersfontein Pan (Butzer *et al.,* 1973; van Zinderen Bakker and Butzer, 1973). This lake had an area of 44 km^2. Assuming a 6°-C drop of mean annual temperature, with evaporation reduced from 2120 to 1400 mm, as well as a runoff quotient of 9.3%, it is possible to show that a rainfall of 878 mm (compared with 397 mm today) would be necessary to balance with a 12-m lake. In other words, even with an appreciable cooling, a rainfall increase of at least 100% is implied for the last submaximum of the Last Glacial. This bears out the pollen records of two key spring sites (van Zinderen Bakker and Butzer, 1973; Coetzee, 1967). Contemporary "periglacial" phenomena from various parts of southern Africa are critically discussed by Butzer (1973a), and the significance of frost weathering during the first part of the Last Glacial at several coastal cave sites requires winter temperature depressions of, at the very least, 10° C (Butzer, 1973b, and in prep.). Interestingly, south of the Cape ranges, the southern coast of South Africa records soil development, under mesic conditions, during the warmer parts of the late Pleistocene and Holocene, with arid conditions and widespread eolian activity —in what are now rain forest environments—during the glacial submaxima. Unlike along the southern margins of the Sahara, major eolian activity in the lower Congo Basin and in Angola did not take place during the later pleniglacial, but for a few millenia only after 13,500 B.P. and again after 4500 B.P. An identical pattern is indicated in the Vaal Basin.

Synopsis: This overview of the African late Pleistocene demonstrates several facets of the Last Glacial time span that may be characteristic of other tropical areas as well: (a) There were major contrasts between the several climatic provinces. (b) The last submaximum of the Last Glacial (ca. 20,000 ±3000 B.P.) was relatively dry everywhere, except in the interior of southern Africa, and possibly also in the Congo Basin; the interval approximately 17,000 to 12,000 B.P. was dry or turning drier everywhere, except in parts of the Nile Basin; all areas except for the Mediterranean borderlands experienced one or more moist intervals during the early and middle Holocene. (c) There was no one-to-one correlation of African climatic anomalies with the higher latitude glacial chronology, and wavelengths, phase, and direction of such changes differed appreciably. (d) In Africa climatic changes with a wavelength of several millenia repeatedly had a greater amplitude than the median difference between "glacial" and "interglacial" conditions in this area. The secondary impact of northern hemisphere glaciation was not insignificant, but primary changes of the general circulation had a direct and far more immediate impact.

OVERVIEW AND DISCUSSION

The body of information selectively reviewed here allows a number of generalizations regarding the temporal and spatial aspects of Pleistocene climates:

1. The early Pleistocene (ca. 1.8 to 0.7 million years ago) saw a series of cool and warm oscillations that ultimately climaxed in successive spasms of continental glaciation in North America and Europe. These early cool episodes saw the onset of glaciation in suitable high-mountain and high-latitude localities, with repercussions among mid-latitude floras.

2. The middle and late Pleistocene (beginning 0.7 million years ago) witnessed a series of major glacial episodes, synchronous on all continents, and with a variable duration averaging near 50,000 years. At their maximum these glacials saw mean annual temperatures 15° to 18° C colder than today on those continents affected by major glaciation, and perhaps 6° to 10° C colder on the landmasses of lower middle latitudes and the tropics (see Fig. 1c). Probably, winter temperature anomalies were twice as great as those of summer. Because ocean-bottom waters are near the freezing point today, probably oceanic temperatures were not much lower than today, although an overall reduction of planetary temperatures is suggested, possibly in the order of 5° C. Latitudinal temperature gradients were greater by perhaps 50%, with summer thermal gradients on the glaciated northern hemisphere continents almost comparable to those of the winter season today. Correspondingly, land-sea gradients must have been exaggerated, with unusual air-mass contrasts.

3. Despite reduced temperatures and evaporation, the Pleistocene glacials were dry in mid-latitudes, with an inverse correlation between temperature and precipitation. In lower latitudes the glacial maxima were mainly dry as well, with the interstadial warm-ups relatively moist. Initially the late interglacial-early glacial transitions appear to have been wet at most latitudes. The surprising aridity of the glacial maxima (Fig. 1c) appears to owe in part to reduced evaporation over cooler ocean surface waters, greater atmospheric stability over the tropics (Kraus, 1973), and more extensive seasonal or permanent pack ice.

4. A major exception to this rule of glacial aridity comes from southern Africa (Fig. 1c). The most reasonable explanation for this is a weakening of the subtropical high-pressure cells (see Williams and Barry, 1974), with reduced upwelling and cold in the Benguela current. Thus, while contraction of the monsoonal rainfall belt (Kraus, 1973) brought West and East Africa into the dry Saharan peripheries, the Namib and Kalahari experienced greater summer rains.

5. The Pleistocene interglacials were in part comparable to the Holocene thus far, but longer and more complex, in part moister, in part either warmer or colder. Mean duration, again quite variable, was a little less than that of the glacials.

The preceding generalizations do little credit to the true complexity of Pleistocene climatic change as briefly reviewed in this paper. Nor can they hope to describe the interrelationships between the radiation budget, hydrologic balance, and the general atmospheric circulation during the course of an "extreme," *i.e.,* glacial, Pleistocene condition. For this reason, an explanatory model (Fig. 4) has been attempted for the observed zonal and regional climatic anomalies that are abstracted by Fig. 1c. This tentative scheme is specifically directed to the Last Glacial, but may serve as a working hypothesis for earlier glacials as well. The initial causative factors of northern hemisphere glaciation and global cooling are not explicitly stated by Figure 4. Presumably they are to be sought in a selection of convergent processes among a range of terrestrial, astronomical (orbital), and solar variables.

In considering such background causes or triggering mechanisms for ice ages and glacial-interglacial oscillations, it is essential to recall that global climatic changes show a wide range of periodicities. Furthermore, each wavelength category may be related to one or more different causative factors. In particular, developing a classification originally devised by Manley (1953), several orders of climative variation can be profitably distinguished:

First Order: Ice ages, such as the Eo-Cambrian, Permocarboniferous, and Pleistocene, appear to show a periodicity of about 275 million years, although there is an increasing body of information favoring late Ordovician glaciation in former South Pole proximity in the Sahara, and of early Carboniferous glaciation in several parts of Gondwanaland (Crowell and Frakes, 1970). If indeed we are dealing with brief and periodic aberrations, it is possible that external periodicities are involved, for example, the sun may be a star of variable luminosity. If, on the other hand, claims for long-term Palaeozoic glaciation in high latitudes can be verified, then the critical location of continental plates with respect to ocean currents and one or the other of the poles assumes primary importance.

Second Order: Climatic shifts with a wavelength of several million ($x.10^6$) years can be noted at several points of the Tertiary record (Fig. 2). These appear difficult to explain except in terms of tectonic history or solar variability. Speculation on these problems will remain unproductive until considerably more better-dated and less equivocal data are available for the Tertiary.

Third Order: Climatic cycles, such as the glacials and interglacials of the Pleistocene, with a wavelength of $x.10^4$ years, probably reflect a combination of primary and secondary cumulative factors. Involved may be long-term trends, such as the orbital

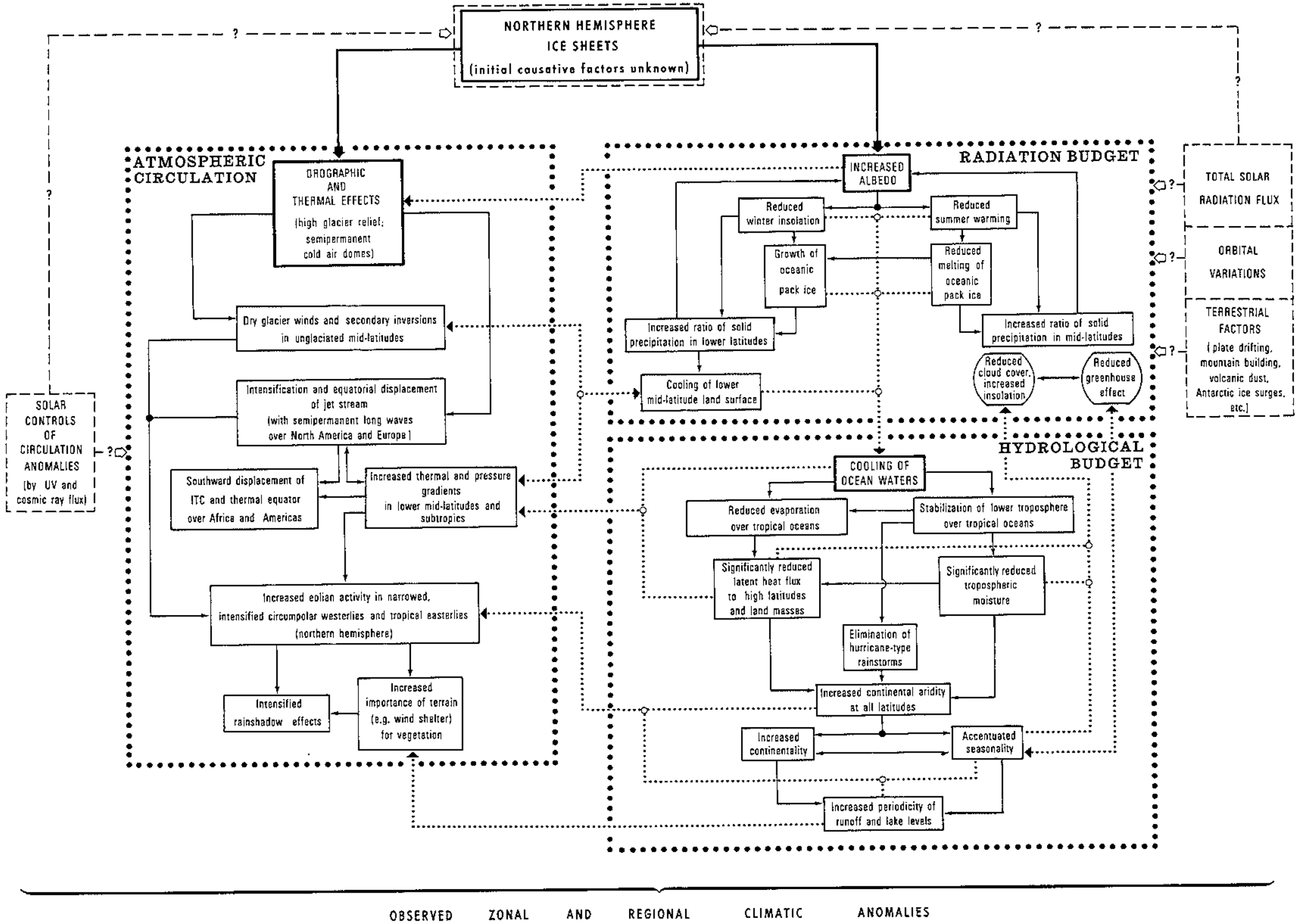

FIGURE 4. An explanatory model for glacial climates.

periodicities expressed in various recalculations of the Milankovitch "radiation curve;" short-term impulses, such as general circulation anomalies (related to solar controls); changes in atmospheric transparency (volcanic dust, carbon dioxide, ozone, water vapor); temporary opening of the Arctic Ocean; Antarctic ice surges, etc. The case for occasional openings of the Arctic Ocean, to provide a brief but potent source of precipitation for the High Arctic, has become increasingly dismal because it is now apparent that the warmest episode of the Last Interglacial predated the critical onset of the last glacial by almost 50,000 years, while the Arctic deep-sea record now shows that the Arctic Ocean has been continuously ice bound for at least 700,000 years (Hunkins *et al.,* 1971). An appealing geophysical theory for the role of the Antarctic in preconditioning and then triggering global climatic changes has been most recently outlined by Flohn (1973), but the critical Antarctic ice surges—with sudden extrusion of great masses of ice far out into the southern oceans, and a correspondingly disastrous affect on hemispheric temperature and albedo—find no geological support by way of high sea-level traces. Our model (Fig. 4) shows that no matter what the initial, cumulative factors were, once formed, the northern hemisphere glaciers maintained a remarkably stable ocean-atmospheric system. In fact, for almost 50,000 years, the resulting feedback mechanisms were able to counteract the repeated, strong warming impulses apparent in the Greenland isotopic record (Fig. 3). Ultimately the system collapsed with a sudden atmospheric reversal that nonetheless required 7000 years to warm the higher latitudes.

Fourth Order: Unduly neglected in the palaeometeorological literature are the intermediate periodicities of a few millenia (x.10^3 yr). These include all short-term impulses listed in the third order, any one of which could have been preeminent, if not instrumental, in driving the glacial-interglacial pendulum back and forth over its critical threshhold. Our discussion of the different facets of the Pleistocene record has deliberately emphasized these pulsations, in the belief that better temporal and dimensional resolution of these features will be crucial toward understanding the nature and ultimate causation of Pleistocene climatic anomalies. Just as increasingly strong second-order pulsations began in Miocene times in anticipation of the Pleistocene third-order changes, strong fourth-order oscillations began 20,000 years before the termination of the Last Interglacial and continued until the end of the Last Glacial; oscillations of comparable amplitude have not yet been experienced during the Holocene, implying that the next glacial is not imminent.

Fifth and Sixth Orders: Less significant for the purposes of this paper are the short-term changes of several centuries (x.10^2) and several decades (x.10) recognizable within the historical era. Their primary importance for the Pleistocene record lies in the actualistic experience that they can provide for explaining the complex dynamism of Earth's atmosphere.

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