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The Geologic Record of Climatic Change

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THE GEOLOGIC RECORD OF CLIMATIC CHANGE

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TABLE OF CONTENTS

ABSTRACT

This paper reviews the major results from paleoclimatic investigations, and includes background material slanted towards the reader who is not a paleoclimatologist - e.g., climate modellers, geophysicists, and space physicists. The time interval surveyed extends from the formation of the Earth 4.6 billion years ago (B.y.) to the development of the instrumental record. The principal results are:

1. 4.6 - 2.3 B.y. — the earth was apparently ice-free, despite a substantially lower solar luminosity. An enhanced atmospheric greenhouse effect may have compensated for the decreased insolation receipt.

2. 2.3 B.y. — evidence for the first glaciation seems to mark a threshold temperature through which the atmosphere passed.

3. 2.3 - 0.9 B.y. — The earth was apparently ice-free, despite low luminosity <u>and</u> a presumably depleted greenhouse effect. A suitable explanation for this phenomenon remains one of the more important unattended questions in paleoclimatology.

4. 0.9 - 0.6 B.y. — three major phases of glaciations occurred, with paleomagnetic data suggesting ice in low latitudes. This problem has also not received much attention. Lack of evidence concerning the presence of low-latitude sea ice implies that low-latitude continental glaciation should not be construed to indicate an ice-covered earth.

5. 600 - 100 million years ago (M.y.) — climates were generally mild, but punctuated by two major phases of ice growth. Paleomagnetic data indicate a strong correlation between the location of continents in high latitudes and the formation of ice sheets.

6. 100 - 50 M.y. — mild, generally non-glacial climates prevailed. The Late Cretaceous (100 M.y.) interval has served as a testing ground for atmospheric and oceanic circulation models of a non-glacial world. A substantial amount of the changes in solar insolation receipt can be accounted for by paleogeographic factors (land in high latitudes, the amount of land flooded by shallow seas, and the amount of land in tropical latitudes). However, simple atmospheric circulation models still cannot account for the relatively high temperatures in polar latitudes. Additional factors (cloudiness, oceanic circulation changes) may be important for balancing the heat budget.

7. 50 - 3.0 M.y. — sequential cooling of the globe occurred, with the Antarctic Ice Sheet forming at least by 14 M.y., and perhaps as early as 30 M.y. Mid-latitude northern hemisphere glaciation commenced at about 3 M.y.

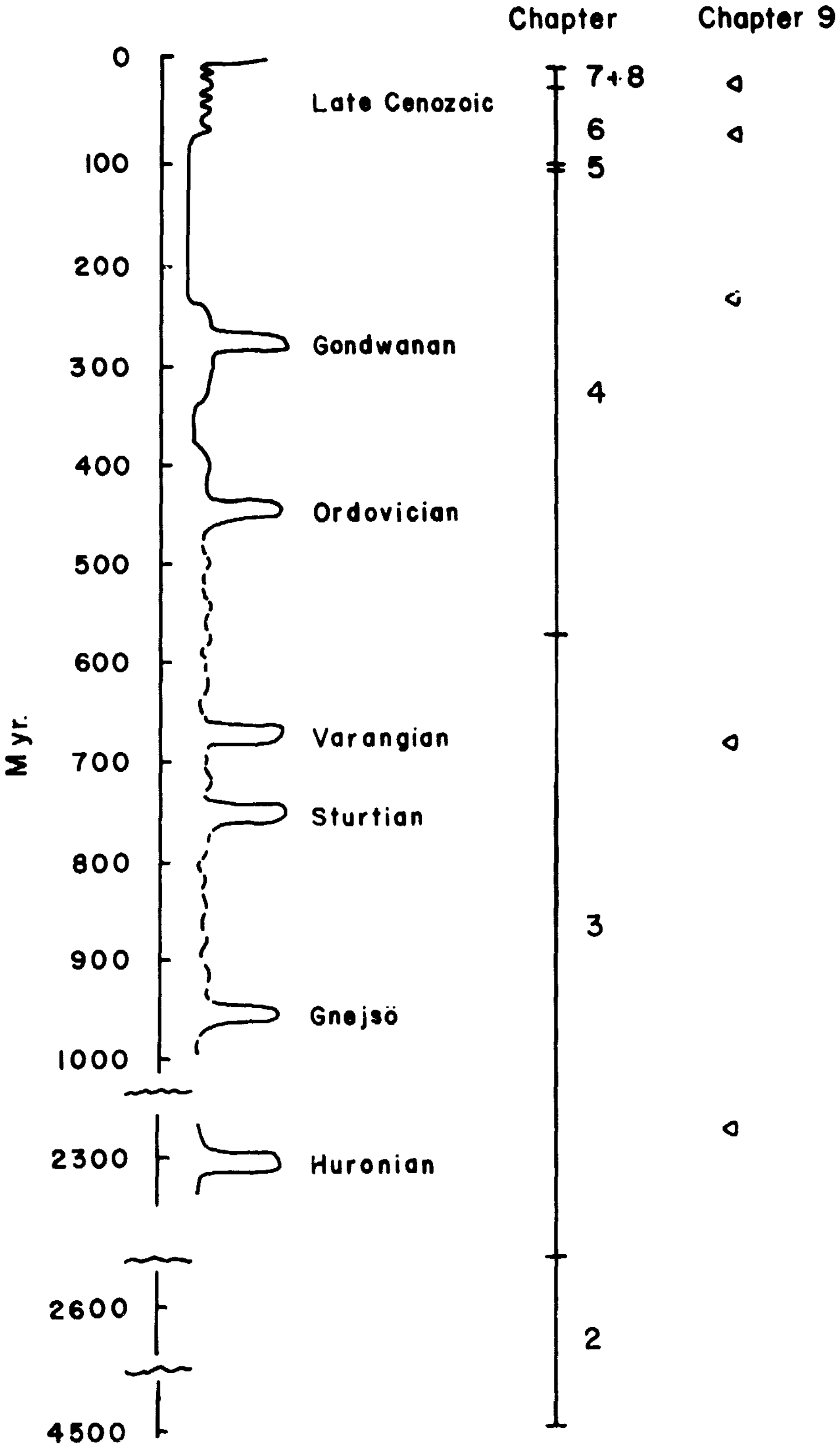
8. 3.0 - 0.0 M.y. — numerous oscillations of northern hemisphere ice sheets occurred, with orbital perturbations affecting the fluctuations in a manner that is presently not well understood. Equatorward of the sea ice/ ice sheet margins, ocean temperature changes were generally small, and the surface of the planet was significantly drier. The ocean appears to have played a major feedback role in both glacial inception and disintegration.

9. A re-interpretation of the pre-Pleistocene oxygen isotope record implies that the frequency of glaciation through the last 600 M.y. may be significantly greater than the 15-20% figure suggested by direct glaciological evidence.

10. The variance spectrum of the climatic record indicates an important stochastic contribution to climatic fluctuations that have periods longer than a few centuries.

11. Geological data provide a useful frame of reference for studies attempting to evaluate the impact of a future CO_2 - induced warming. Greenhouse warming estimates of 2-3°C represent one-fourth of the total range estimated for the last 100 million years. The estimated rate of temperature change for the CO_2 effect exceeds by an order of magnitude the changes that occurred during the glacial terminations of the Quaternary. Analyses of Late Quaternary intervals that were slightly warmer than the present provide support for the following anticipated changes: increased drought on the North American plains, increased moisture in North Africa and the Middle East, and a global sea level rise of about 5 m. However, geological data suggest that the CO_2 - warming would not result in the complete removal of the Arctic Ocean ice cover — this region has apparently had at least a seasonal ice cover for the last five million years.

12. The inferred planetary temperature pattern during the last four and a half billion years has both secular and fluctuating components. Secular changes have been influenced by changes in solar luminosity. Fluctuations have been influenced by paleogeographic changes, orbital perturbations, and feedback interactions within the land-sea-air-ice system. Glacial ages may be due to orbitally-induced forcing superimposed on a global cooling of terrestrial origin.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THE GEOLOGIC RECORD OF CLIMATIC CHANGE

I. INTRODUCTION

Available geological records provide direct evidence of glaciation for only about 5 - 10% of earth history (Fig. 1). The science of paleoclimatology attempts to describe and explain these occurrences. The purpose of this paper is to provide a survey of the major results, including background material slanted toward the reader who is not a paleoclimatologist - e.g., climate modellers, geophysicists, solar physicists, etc. The scope of the study extends from the formation of the earth 4.6 billion years (B.y.) ago to the development of a relatively widespread instrumental network during the last century (see Fig. 2 for the geological time scale). Space limitations will necessarily dictate some restrictions on subject matter and dissenting opinions. To compensate, a brief annotated bibliography of the major reference works in the field is included in the Appendix.

The paper is divided into ten sections, each of which may be read separately: 1) Introduction; 2) Early Precambrian (4.6 - 2.5 B.y.); 3) Late Precambrian (2.5 - 0.57 B.y.); 4) Paleozoic and Late Mesozoic (570 - 100 M.y. - million years ago); 5) Late Cretaceous (100 M.y.); 6) Late Cretaceous to Late Cenozoic (100 - 1.7 M.y.); 7) Pleistocene (1.7 - 0.01 M.y.); 8) Holocene (the last 10,000 years); 9) Geological evidence of solar variability;

Fig. 1 Ice ages through geological time, indexed to chapters discussed in text. The curve refers to periods of major ice sheet formation. Open triangles for Chapter 9 refer to sections of the geological record that contain unusually high-resolution records of inferred climatic fluctuations (see text). (After D. H. Tarling, Climatic Change; reproduced with permission of Cambridge Univ. Press).

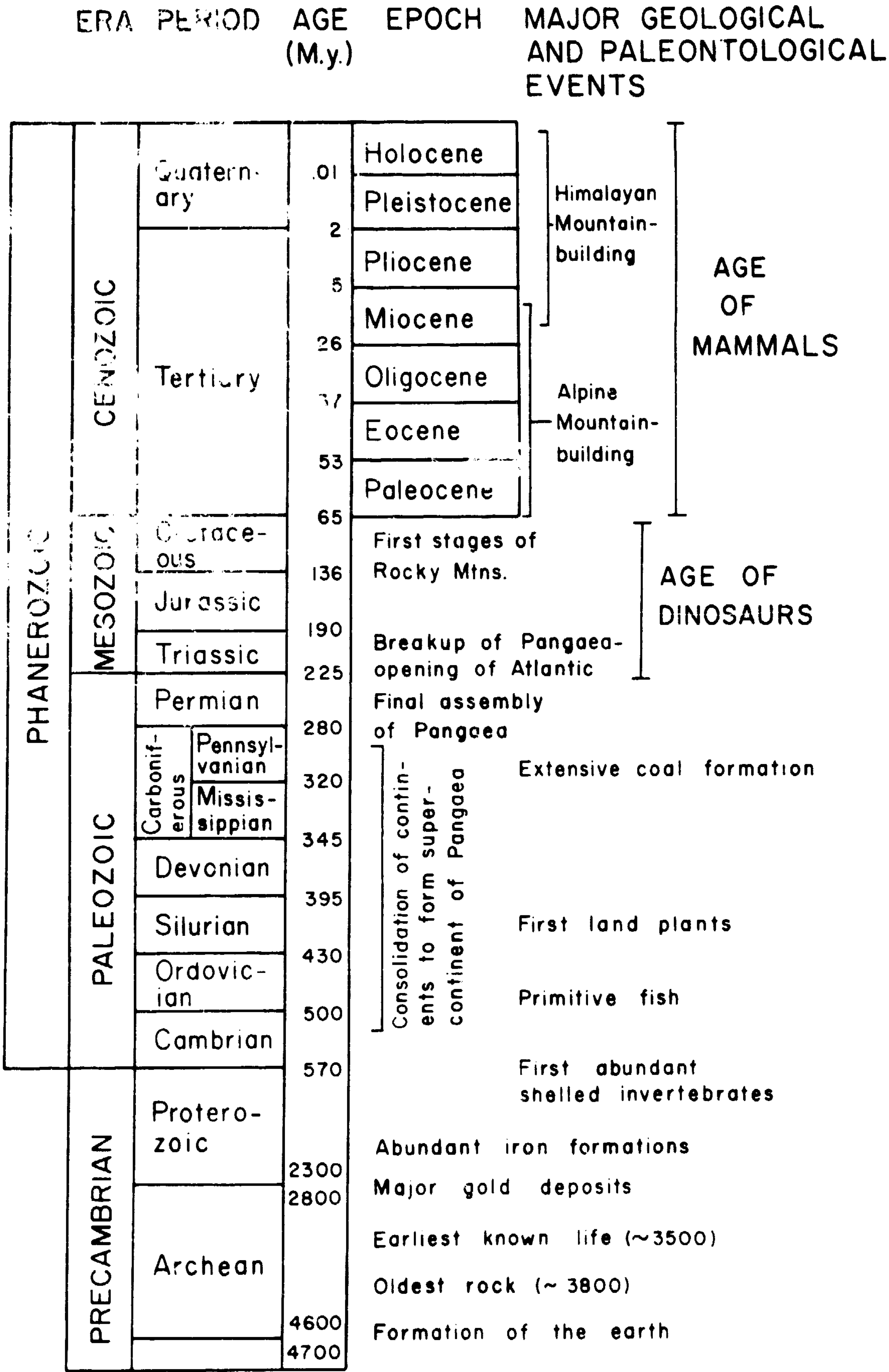

Fig. 2 The geological time scale.

10) Summary and Conclusions. The uneven partitioning of the time intervals is reflective of the different types of climatic results available; each of the intervals encompasses one of the major themes of paleoclimatology.

Before embarking into the nature of the evidence, it is first important to remind the reader of a fundamental feature of the geologic record — most rocks that have formed during earth history have been subsequently eroded. The older the rocks, the greater the chance for erosion. Thus, there is a disproportionate abundance of relatively young rocks in the record. Consequently, there is always the chance that the evidence at hand is not a faithful reflection of the average climatic conditions throughout earth history. It is essential that the reader bear in mind this feature, which Darwin called "the imperfection of the geologic record". The problem assumes its greatest significance in the discussion on the Precambrian (4.5 - 0.6 B.y.) - i.e., the first 85% of earth history.

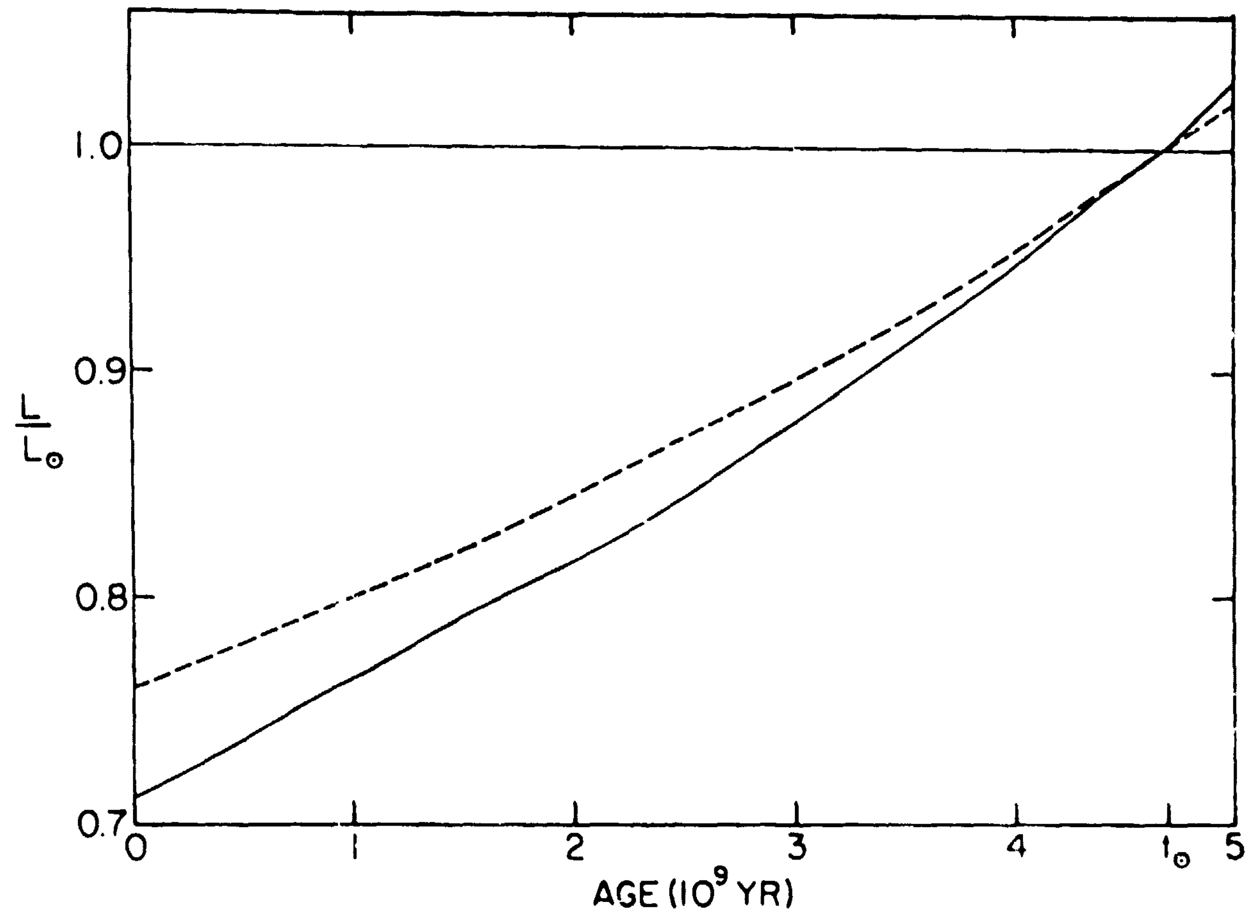

Fig. 3 Long-term evolution of the solar luminosity. The dashed line shows evolution predicted by the scaling model of Endal (1981). The solid line illustrates predictions from a detailed computer model (Endal and Sofia, 1981). (From A. Endal, 1981; reproduced with permission of the author).

II. EARLY PRECAMBRIAN (4.6 - 2.5 B.y.)

Perhaps the most challenging climatological question of the Early Precambrian (Archean) involves the "faint young sun" paradox (Ulrich, 1975). Virtually all models of solar evolution (e.g., Newman and Rood, 1977) indicate a solar luminosity increase of 20-30% during the past 4.7 billion years. Endal (1981) has shown that the luminosity estimates (Fig. 3) are not affected by more questionable estimates of nuclear reaction rates, which are used to predict phenomena such as neutrino flux (see, e.g., Davis et al, 1978; Bahcall, 1979): scaling arguments identify luminosity as being primarily dependent on the mean molecular weight of the sun. This quantity increases with time due to the conversion of hydrogen to helium in the sun's core. (Because the sun can be considered an ideal gas, an increase in molecular weight requires an increase in temperature in order to maintain hydrostatic equilibrium. In turn, increased temperatures cause an increase in nuclear reaction rates.)

The above results are of significance because models of climatic change (e.g., Wetherald and Manabe, 1975; North et al, 1981) indicate that a 5-10% reduction of the solar constant from present values should result in an ice-covered planet. Strong ice-albedo feedback effects might then maintain this condition even if the solar constant were to increase to values greater than the present. Terrestrial feedback processes are therefore required in order to explain an apparently ice-free earth for the first one-half of its history (Fig. 1). The terrestrial mechanisms will be discussed at some length in order that the reader may develop some sense of the processes and subtle complexities of the geological system.

Relevant details of early earth history involve evolutionary features of the lithosphere, hydrosphere, and atmosphere. For example, the areal extent of continents during the first half of earth history may have been significantly less than at present. From 4.6 - 3.9 B.y. the crust was probably in a high-temperature state as a result of: 1) radioactive heat flux 3-4 times greater than present values (e.g., Goodwin, 1981); 2) intense meteorite bombardment from 4.2 - 3.9 B.y. (Goodwin, 1976); 3) gravitational energy released during the formation of the earth's core. RamaMurthy (1976) has calculated that the latter effect alone could have raised the average temperature of the planet by 1200°C. Nevertheless, radiometric dates from southwest Greenland (Moorbath et al, 1975) and eastern India (Basu et al, 1981) indicate crustal formation by 3.8 B.y. Rounded sedimentary particles in the Greenland rocks record the presence of a hydrosphere.

Between 3.8 - 2.7 B.y. several models of crustal evolution (e.g., Burke et al, 1976; Goodwin, 1981; cf. Ashwal, 1981) suggest only microcontinent-scale land masses until the end of the Archean. These models are supported by $^{87}Sr/^{86}Sr$ ratios of Archean carbonates (Veizer, 1976), which [illegible] a minimal contribution of continental source rocks to the Archean ocean (continental and mantle rocks have different ratios). The formation of extensive continental rocks may have been a consequence of the thermal evolution of the planet. For example, the virtual absence of a very high-temperature (Sleep, 1979) type of volcanic rock after the Archean suggests a transition to lower thermal regimes after 2.5 B.y. (Goodwin, 1981). The weakening of the geotherm would have allowed segregation of lighter, silica-rich, granitic magmas (Lambert, 1976). This reaction may then have triggered the massive thermal event at the end of the Archean - an event that produced 50–60% of the present granitic continental crust (Ronov, 1968). Subsequent to 2.5 B.y., there has been no shortage of continent-sized land masses.

An enhanced greenhouse effect (e.g., Hart, 1978; Owen et al, 1979) may have played a pivotal role in maintaining moderate surface temperatures during the early stages of earth history. Sagan and Mullen (1972) proposed that ammonia may have been the agent responsible for the greenhouse effect. However, Kuhn and Atreya (1979) note that NH_3 photodissociates after only 40 years. Increased concentrations of CO_2 and H_2O may be more likely candidates for a greenhouse effect. Higher partial pressures of CO_2 may be attributed to increased early outgassing. Atmospheric concentrations may also have been affected by absence of an effective crustal sink for the carbon. This statement requires some explanation.

It has long been suggested (Urey, 1952; Pollack and Yung, 1980) that the CO_2 was removed via the breakdown of silicate minerals and the precipitation of inorganic calcium carbonate. Schematically,

$$CaSiO_3 + CO_2 \rightarrow CaCO_3 + SiO_2 \tag{1}$$

This process could have operated in the pre-biotic era (the earliest fossils now date at about 3.5 B.y. -- Walter et al, 1980). However, early carbonate precipitation should have been inhibited by low pH values (Garrels and Perry, 1974; Schopf, 1980), and carbonates dissolve at pH < 7.0.

An additional problem concerning early removal of CO_2 stems from the fact that virtually all organic and inorganic carbon now in the crust seems to have originated via biological mediaries (e.g., Matthews, 1974). Even the fine-grained constituents of limestone are of biologic origin. For example, on the present south Florida carbonate shelf, a single genus of calcified

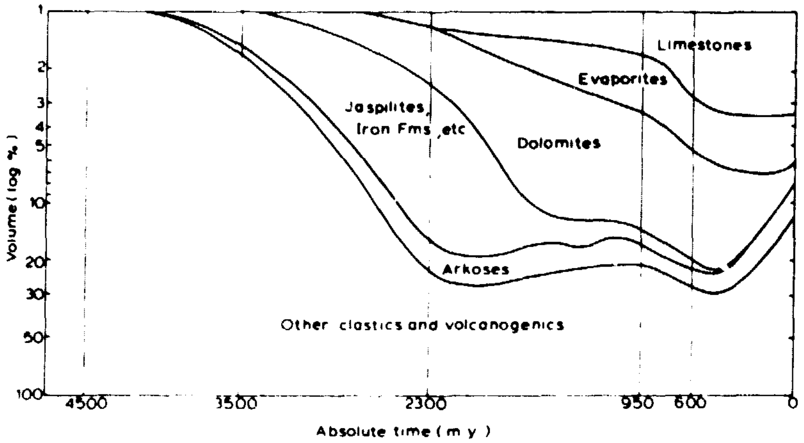

Fig. 4 Estimates of the abundance of sedimentary rock types (log scale) throughout earth history. Dolomites are magnesium-rich carbonates. Note that the abundance of carbonates increases markedly after an expansion of life dated at about 2.3 B.y. (Frakes, 1979). (After Ronov, 1964; reproduced with the permission of the American Geological Institute.)

green algae produces the vast majority of the carbonate mud in the surrounding environments (Stockman et al, 1967). Precambrian carbonate formation may also have been controlled by organisms, for the carbonates commonly occur in the form of stromatolites (finely-laminated rock). The origin of this rock type seems dependent on the presence of bacteria or blue-green algae (Schopf, 1980). Carbonate abundances that parallel the expansion of life after 2.3 B.y. (e.g., Frakes, 1979 - see Fig. 4) further support the concept of a biological control on the carbon sink.

Since the crust contains over 99.9% of the carbon in the ocean-atmosphere-crust system (Garrels and MacKenzie, 1972), material balance considerations therefore suggest a pre-biotic re-partitioning of the carbon into either the atmosphere or ocean. CO_2 partial pressure estimates of 0.01-0.10 atm. exceed present values by 30-300X (Garrels and Perry, 1974; Holland, in Pollack and Yung, 1980). The high Mg concentrations in Precambrian carbonates (Fig. 4) are also consistent with an increased partial pressure of CO_2 (Holland, 1976). Owen et al (1979) calculate that an enhanced CO_2-H_2O greenhouse effect could have produced a mean temperature of 310°K at 4.2 B.y. (the present mean is 287°K). Empirical support for high temperatures is provided by preliminary isotopic analyses of cherts, which indicate ground water temperatures of 340°K as late as 2.8 B.y. (Knauth and Epstein, 1976).

In summary, climatic cooling induced by a faint sun may have been offset by an atmospheric greenhouse effect, caused by high CO_2-H_2O concentrations. The first glaciation occurred about 2.3 B.y. This date coincides with a rapid expansion of stromatolites (e.g., Frakes, 1979) — an event that may have signaled an increased withdrawal of CO_2 from the atmosphere by photosynthetic organisms.

III. LATE PRECAMBRIAN (2.5 - 0.57 B.y.)

The Late Precambrian (Proterozoic) contains evidence for two phases of continental glaciation — at 2.3 and 0.9-0.6 B.y. (Frakes, 1979). Before discussing these events, it is useful to digress briefly and explain some of the geological criteria used to infer ancient glaciations. Three important criteria are: 1) tillites (mixtites) — a mixture of fine-and coarse-grained sedimentary particles that are deposited by glaciers; the debris in the base of the glaciers is left behind when the glacier melts; 2) dropstones (ice-rafted detritus) — fine-to-medium grained layered sedimentary rocks that contain large particles with associated impact structures; the features form when icebergs melt, allowing their enclosed sedimentary particles to plummet through the water column, impacting onto the soft sediment below; 3) striations — grooves engraved in hard rock pavements by grains ground into the base of moving ice sheets; the subparallel lineations enable determination of glacial flow directions.

The first evidence for glaciation occurs in 2.3 B.y. old rocks from North America, South Africa, and Australia (Frakes, 1979). The areal extent of the individual glaciers is not well known. Their presence seems to mark a threshold temperature through which the atmosphere passed. The combination of low luminosity (Fig. 3) and a "thinned" greenhouse shield could explain the low temperatures. However, a puzzling dilemma arises when comparing the 2.3 B.y. glacial event with the climate of the subsequent 1.4 B.y.: there is little evidence for glaciation between 2.3 - 0.9 B.y. (Frakes, 1979). As mentioned in the introduction, the gap may be due to erosion. Nevertheless, aluminous clay minerals (kaolinite) indicate intense chemical weathering,

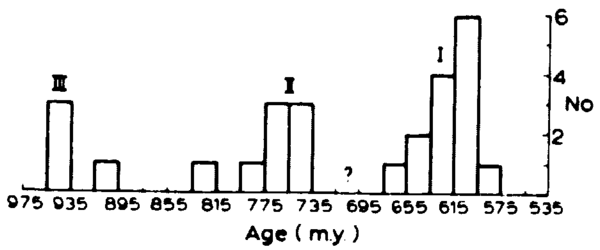

Fig. 5 Histogram of age estimates for Late Precambrian glaciations (from G. Williams, 1975; reproduced with permission of the Cambridge Univ. Press).

typical of tropical environments, in sediments overlying the Canadian glacial deposits (Young, 1973; see also Frakes, 1979). Preliminary oxygen and hydrogen isotope values from cherts also suggest ground water temperatures as high as 20-33°C at 1.2 B.y. (Knauth and Epstein, 1976). Physical explanations for the conjectured moderate climates of the period 2.3 - 0.9 B.y. have not been adequately explored.

One of the most unusual climatic events in earth history took place between about 0.9 - 0.6 B.y. At least three glaciations of continental scale occurred (Fig. 5). Virtually all regions on the earth that contain Precambrian rocks show some evidence for glaciation during this time (Fig. 6). Furthermore, paleomagnetic evidence suggests that many of the glaciations may have occurred in low latitudes (McWilliams and McElhinny, 1980; Christie-Blick, 1982; see also Frakes, 1979). Land elevation may have been relatively low, for marine sediments are interlayered with some of the glacial deposits (Frakes, 1979). In fact, Frakes, (1979, p.88) has stated that "...if mixtites were not known from the Late Precambrian, the proportions of shelf carbonate could be taken as evidence for widespread and continuously warm climates."

The causes of Late Precambrian glaciations are poorly understood. That the glaciers penetrated into lower latitudes than their recent counterparts may not necessarily be very significant, for luminosity output was still 10% less than the present (Fig. 3). Nevertheless, it is important to add two notes of caution in interpreting the glaciations: 1) apparent near-global distribution of glacial deposits (Fig. 6) does not necessarily imply nearly-synchronous worldwide glaciations. The continents could have drifted extensively during the period of 300 M.y. For comparison, 100 million years ago India was in the southern tropics; 200 millions years ago the continents

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Fig. 6-a

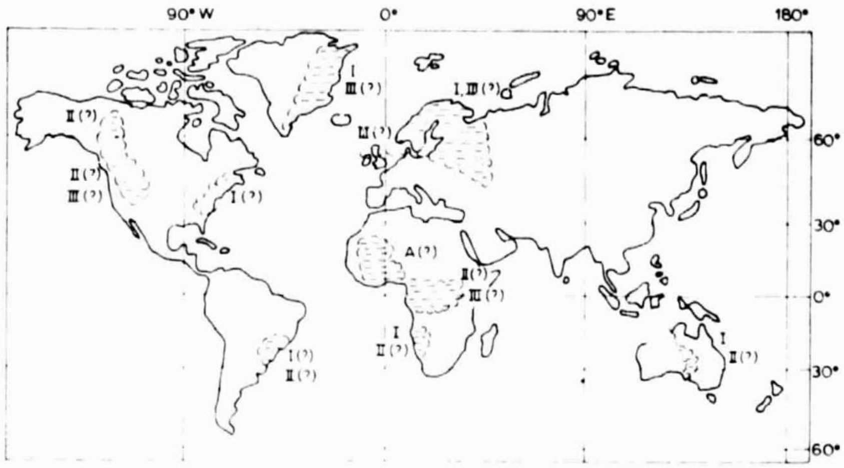

Fig. 6-b

Fig. 6 Comparison of Late Precambrian glacial distributions with the world distribution of Precambrian rocks in modern continental stable areas.

(a) Global distribution of major glacial centers on a map showing the present dispersal of continents, and their assignment to the time intervals (I, II, III) of G. Williams (1975) - see Fig. 5. The letter A signifies that all three time intervals may be represented (from Frakes, 1979; reproduced with the permission of the author and Elsevier Scientific Pub. Co.).
(b) World distribution of Precambrian rocks in modern continental stable areas. Note that some of the glacial deposits are from marginal marine locations, which are not classified as stable areas. (From Eicher and McAlester, 1980; reproduced with the permission of Prentice-Hall Pub.)

were all united into the supercontinent of Pangaea; and 300 million years ago North America straddled the equator; 2) even a synchronous low-latitude glaciation does not imply an ice-covered earth (a condition considered stable to large perturbations: see North et al, 1981). The planet was presumably still covered by 70% water, and the factors responsible for sea-ice formation and preservation are delicate enough to elicit considerable doubt about the validity of extrapolating ice-covered continents to imply an ice-covered ocean. Studies of present sea-ice formation illustrate this point. For example, Gordon (1981) has suggested that, in the Southern Ocean, the upwelling of unusually warm subsurface water is an important factor in the summer meltback of sea ice. In the North Atlantic, sea ice occurs only where a very low-salinity surface lens overlies distinctly higher-salinity water (Weyl, 1968).

In summary, the interval from 2.5 - 0.57 B.y. records four glacial events, one near the beginning and three near the end. Observations indicate two perplexing paleoclimatological problems: 1) the occurrence of moderate climates during a period of presumably low luminosity and low greenhouse effect; 2) the possibility of extensive low-latitude glaciations. The onset of Late Precambrian glaciations coincides with one of the major crust-forming events in earth history (Goodwin, 1981). These tectonic events link the Late Precambrian with the present tectonic regimes of sea-floor spreading (Goodwin, 1981). Viewed from the perspective of tectonic regimes, continental glaciations have occurred for at least 20% of the most recent (900 M.y.) regime. This figure may be an order of magnitude larger than that conjectured for the first 3.7 B.y. of earth history, and implicates plate tectonics as one of the key underlying causes of glaciation. Further evidence for such a connection is forthcoming in the following sections.

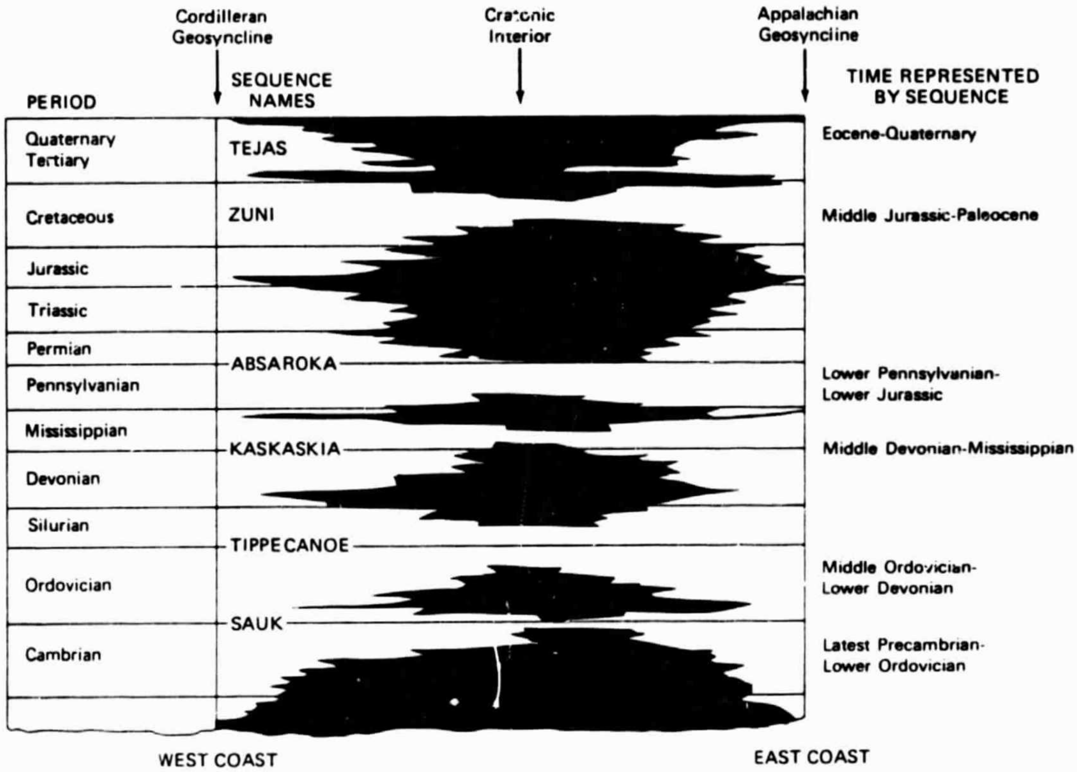

Fig. 7 Evidence for large-scale flooding of continental interiors through geologic time. Dark areas represent large gaps in the rock record, which become smaller toward the continental margin. White areas (preserved strata) represent flooding of the continental interiors by marine transgressions. (From Sloss, 1963; reproduced with permission of the author and the Geological Society of America).

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IV. PALEOZOIC - LATE MESOZOIC (570 - 100 M.y.)

Most textbooks on geology describe the interval from 570-100 M.y. as generally mild, punctuated by two major glacial phases. The impression of moderate climate is derived from both lack of evidence for frequent widespread glaciations, and in the environmental interpretation of sedimentary rocks, particularly limestone. Today, shallow water lime sediments are accumulating primarily in the tropics (e.g., Matthews, 1974). Between 570-100 M.y. geological evidence indicates that tropical seas were much more widespread. Large parts of continental interiors were covered by these shallow, epicontinental (epeiric) seas during marine "transgressions" (sea-level rises). For example, more than two-thirds of North America was covered by a shallow sea 430 M.y. ago (e.g., Dott and Batten, 1976, Fig. 12.16). In a landmark paper, Sloss (1963) showed that the sea level history of North America for the last 570 M.y. (the Phanerozoic Eon - cf. Fig. 2) could be summarized as six sequences of transgressions and regressions (Fig. 7). Pitman (1978, Fig. 8) illustrates a correlation between North American and African sea-level records for the two most recent sequences, thereby indicating that the fluctuations are of a global scope.

Although sea-level fluctuations of the last few million years have been dominated by fluctuations of continental ice volume, it is generally thought that the Phanerozoic transgression - regression sequences are not of "glacio-eustatic" origin. Rather, they may be related to variations in the rate of crustal generation along mid-ocean ridges. Times of increased upwelling of magma from the mantle are associated with a thermal expansion of ridges, displacing water onto the continents (e.g., Hays and Pitman, 1973; Pitman, 1978).

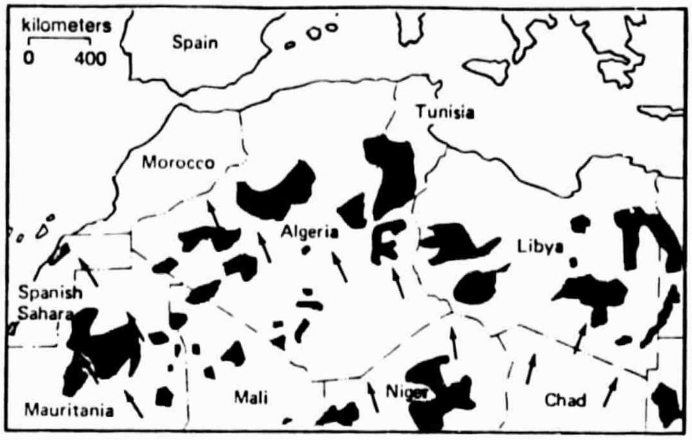

Fig. 8 Evidence for an Ordovician glaciation in the Sahara. Striations and grooves in the bedrock immediately below Ordovician tillites reveal the direction of flow (arrows) of a great ice sheet. Darker areas are sand-covered today. (From Eicher and McAlester, 1980; after Fairbridge, 1970; reproduced with the permission of the author, the American Geological Institute, and Prentice-Hall).

Fig. 9 Evidence indicating that continental ice sheet formation is linked to the location of land masses in high latitudes. Figure illustrates the path of the Gondwanaland supercontinent across the South Pole during the Paleozoic, and inferred distribution of Permian-Pennsylvanian ice sheets. (From Eicher and McAlester, 1980; reproduced with the permission of Prentice-Hall Pub.).

The "maritime" climates of the Paleozoic were interrupted by two major phases of ice growth at about 450 and 250-320 M.y. (Frakes, 1979). Evidence for a major Ordovician ice sheet at 450 M.y. can be found in rocks from the Sahara Desert (Fairbridge, 1970, 1979). Ice flow indicators suggest flow from the vicinity of the present equator (Fig. 8). However, paleomagnetic reconstructions of paleolatitudes indicate that northern Africa was in the vicinity of the South Pole during the Ordovician (Fig. 9). In fact, Figure 9 illustrates a remarkable correspondence between occurrences of Paleozoic glaciations and paleopole positions. The Late Paleozoic glaciations also indicate a polar position for a different part of the Gondwanaland supercontinent (which moved as one unit through most of the Paleozoic). The concept that continents in polar latitudes are essential for ice sheet growth has been strongly influenced by the above result. The reader might object that this result should carry no more weight than the opposite indications from the Late Precambrian. However, geological data for the Late Paleozoic is both more extensive and of higher quality than the Late Precambrian data -- thus the stronger influence.

The final suturing of the continents to form Pangaea (Fig. 10) was completed during the Triassic (200 M.y. ago). Since that time continental fragmentation has proceeded to the present day. The combination of a gigantic landmass and emergent shorelines should have resulted in extremely continental climates. The widespread aridity to be expected from such a climate is reflected in the rock record (Fig. 11). Permian-Triassic evaporite (salt)

Fig. 10 A reconstruction of the probable relationships of the continents in Triassic time before they drifted into their present configurations. (From Dietz and Holden, 1970; reproduced with permission of the American Geophysical Union).

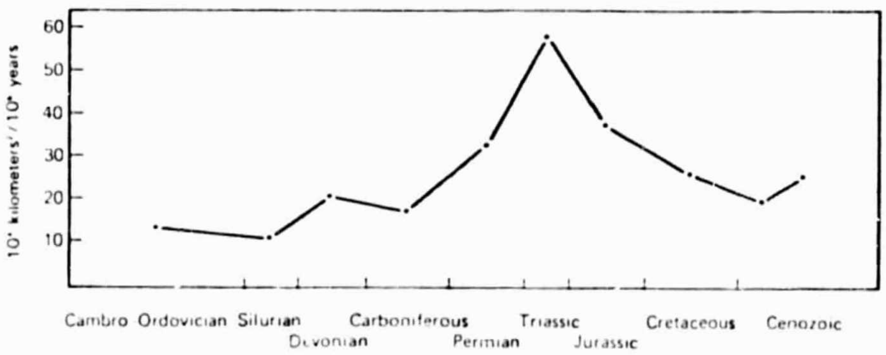

Fig. 11 Areas covered by evaporite deposits throughout the Phanerozoic. Evaporite deposits form in arid environments, and the Triassic peak in formation rate may be due to extreme aridity on a very large land mass. (From Gordon, 1975; copyright Univ. Chicago Press).

deposits - which form in arid environments - are more extensive than at any other time in the last 600 M.y. (Gordon, 1975). Stevens (1977) estimates that the salt content of the evaporite deposits is equivalent to 10% of the present ocean's salt. Average oceanic salinities prior to 200 M.y. should therefore have been 3.0 0/00 greater than the present average of about 35.0 0/00. This increase is more than twice that estimated for Pleistocene ocean enrichment due to removal of fresh water for ice sheet construction (Worthington, 1968).

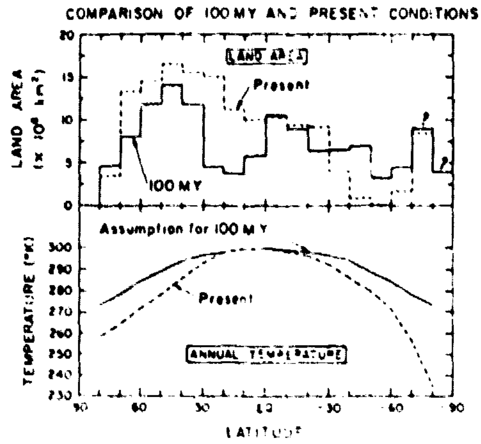

Fig. 12 The area of land in each 10° latitude belt as measured by planimeter for the present and 100 million years (Barron et al. 1980). Present day mean annual surface air temperature with respect to latitude and the temperature assumption for the Cretaceous. (From Thompson et al, 1981; copyright Univ. Chicago Press).

V. LATE CRETACEOUS (100 M.y.)

Evidence presented above indicates that non-glacial climates have prevailed for much of earth history. The Late Cretaceous represents one of the last times for which there is widespread evidence of a non-glacial state. Coral reefs extended to 30° latitude in both hemispheres, and coal deposits formed at 70°N paleolatitude (Frakes, 1979). Breadfruit trees grew at 60°N (paleolatitude) in Greenland (Nathorst, 1911), and palms were present in Alaska (Gordon, 1973). The Arctic Ocean was probably ice-free (Clark, 1977). Large parts of western Europe, North America, and Africa were flooded by shallow epicontinental seas (see below, Fig. 15). Oxygen isotope evidence suggests bottom water temperatures of 14-15°C (Emiliani, 1961). Estimated temperature differences from the present (Fig. 12) were greatest at the poles, and the same or slightly warmer in equatorial regions (Thompson and Barron, 1981). The globally-averaged temperature is estimated to be 6°C warmer than present (Thompson and Barron, 1981; Barron <u>et al</u>, 1981). The more symmetric hemispheric distribution of land, and in particular the decrease in land due to high sea level, (Fig. 12) also suggests a decreased seasonal range of northern hemisphere temperatures.

Models addressing the factors responsible for nonglacial states have often focused on the Late Cretaceous. Paleogeography (e.g., continents in low latitudes) has been perhaps the major hypothesis proposed to explain the nonglacial climates (Frakes and Kemp, 1972; Donn and Shaw, 1977). However, the model based on high-latitude continents causing global cooling has been questioned by the important work of Barron and colleagues (Fig. 13). The amount of continental areas flooded by shallow seas is more important for

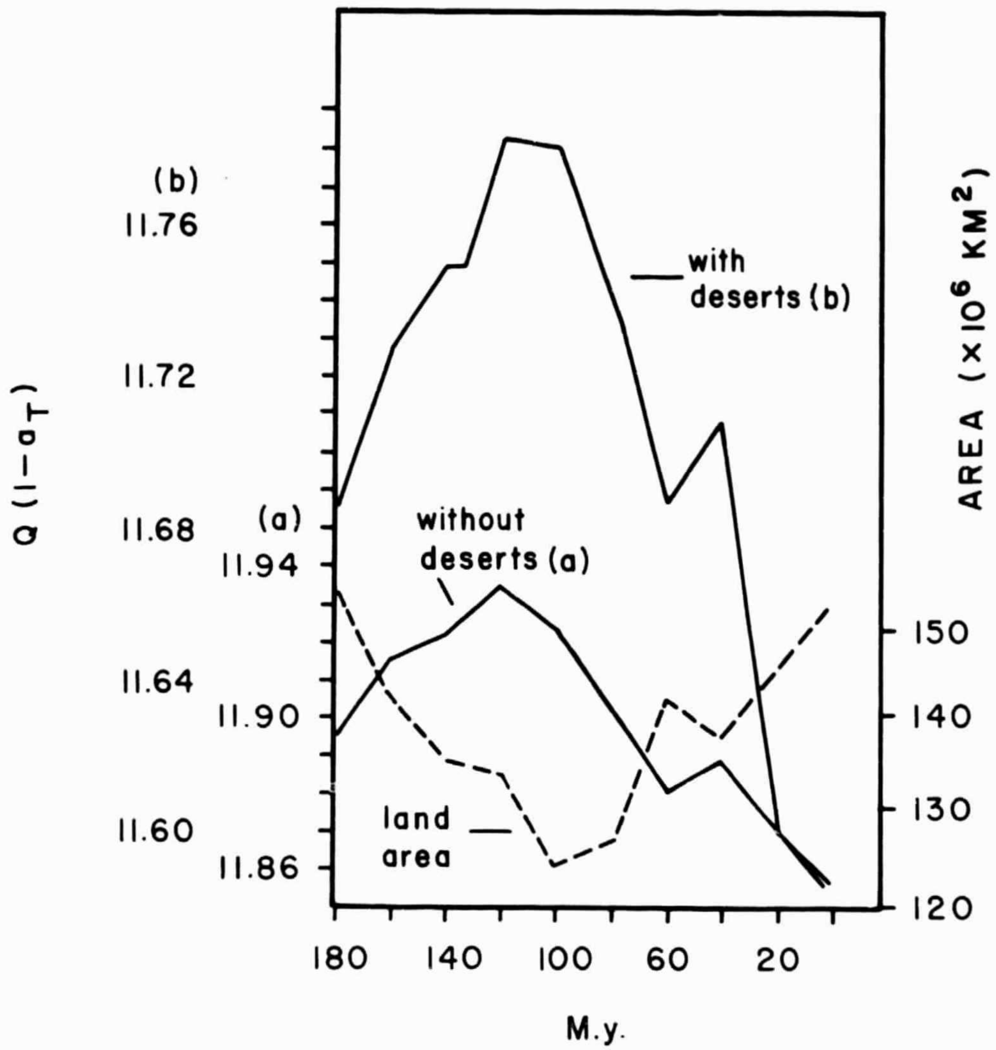

Fig. 13 A comparison of the effect of land-sea distributions and deserts on net global insolation receipt, 180 M.y. to the present. $Q(1-a_t)$ is a measure of the net total global insolation receipt, with Q the total solar radiation received and a_t the global albedo. Calculations are for an atmosphereless earth, with average latitudinal insolation values taken from Sellers (1965). (a) Variations of global insolation due to the migration of continents across latitude belts and the fluctuations of sealevel (e.g., Fig. 15). Dashed line refers to the total area of land from 180 M.y. to the present. Snow and ice prescribed for latitude 70-90°, and no deserts. (b) Total insolation receipt if the 10-30° latitude zone is considered as a higher albedo desert region with an albedo of 0.35. (From Barron et al, 1980; reproduced with the permission of the authors and Elsevier Scientific Publishing Company).

energy receipt than is the amount of land in high latitudes (Barron et al, 1980). The areal extent of deserts also plays an important role. Barron et al (1980) have calculated that large land masses situated beneath the descending branch of the Hadley cell (10-30° latitude) are responsible for a greater reduction in insolation receipt than land masses at high latitudes — reduction is greater for high insolation receipt/relatively low albedo (0.35) than for low insolation receipt/relatively high albedo (0.65). Burrett (1982) has recently extended the above paleogeographic approach to records spanning the last 600 M.y.

The Late Cretaceous is also of special climatological interest because the relatively widespread distribution of rocks provides good spatial control, allowing for more sophisticated climatic analyses. The subject of this section will therefore focus on atmospheric and oceanic circulation models that attempt to describe a more physical picture of a non-glacial climatic state. In a recent study, Barron et al (1981) have employed a zonally-averaged energy balance model (cf. North et al, 1981) in their examination of the Cretaceous climate. The model stipulates the surface temperature of a latitudinal strip as being a function of land/sea heat capacity, incoming and outgoing radiation, and the divergence of poleward heat transport:

$$\frac{\partial}{\partial t}[R(\phi)T_S(\phi,t)] = Q(\phi,t)[1 - \alpha(\phi,t)] - F_{1r}(\phi,t) - \text{div } F(\phi,t) \qquad (2)$$

where R = thermal inertia, T_S = surface air temperature, ϕ = latitude, t = time, Q = incoming solar radiation, α = albedo, F_{1r} = outgoing infrared radiation (parameterized), and F = poleward energy transport by the atmosphere

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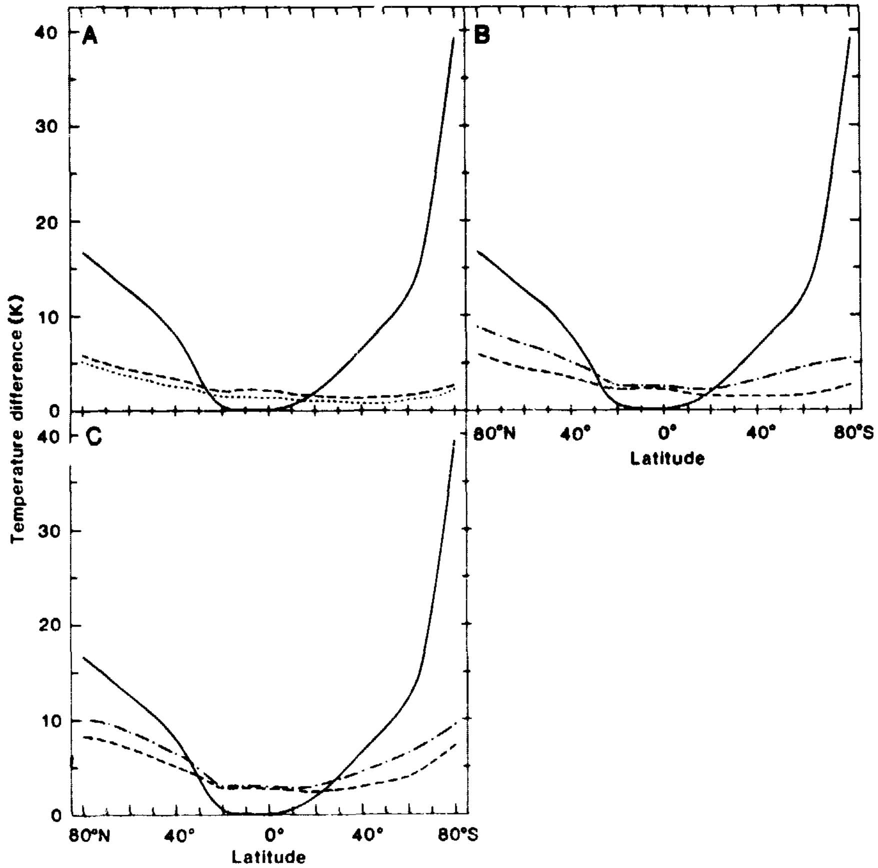

Fig. 14 A comparison of Cretaceous temperature simulations with the hypothetical difference between Cretaceous and present-day temperatures. This figure illustrates the inability of a simple energy balance model to reproduce the major features of the Cretaceous temperature distribution. The solid line illustrates the temperature difference between the present-day control simulation and the nominal Cretaceous distribution (see Fig. 12). The temperature difference between the present-day control simulation and various Cretaceous simulations, with Cretaceous land-sea distribution, are indicated as follows: (dotted line), present base land albedos and present cloud amount; (dashed lines), "wet" base land albedos and present cloud amount; (dots and dashes), "wet" base land albedos and uniform 0.5 cloud amount. (A) Comparison of land albedo assumptions. (B) Comparison of present and uniform 0.5 cloud amounts. (C) Comparison of present and uniform 0.5 cloud amounts with a hypothetical Cretaceous distribution of surface temperature minus cloud top temperature. The closer the dotted or dashed lines curves are to the solid curve, the "better" the simulation (see Barron et al, 1981, for details). (From Barron et al, SCIENCE, 212:501-508, 1 May 1981; copyright 1981 by the American Association for the Advancement of Science).

and oceans. Atmospheric transport is modeled as a diffusive process. Due to the difficulty of reliably estimating even present ocean transport values, Cretaceous values are specified as present annual mean values.

The principal results (Fig. 14) of the Late Cretaceous analysis are: 1) the different land/sea distributions can account for only 40% of the difference between the globally-averaged temperature and the inferred Cretaceous temperature in the model; 2) realistic adjustments to model parameterizations for decreased high-latitude cloudiness, decreased cloud-top radiating temperature, or decreased albedo (due to lush vegetation) — all result in planetary warming but do not achieve the hypothesized distribution of temperature with respect to latitude (Fig. 14); 3) either the geological data is open to major re-interpretation, or important non-diffusive heat transport processes must have been operating to explain warm polar temperatures. Two studies have addressed the latter question by focusing on possible changes in the deep and surface ocean circulation. It is already known that today the ocean is at times responsible for as much as one-third of the poleward energy transport (von der Haar and Oort, 1973). Changes could therefore greatly modify the amount of heat transported to polar regions.

The Cretaceous surface circulation analysis (Luyendyk et al, 1972) involves a reconstruction that used a rotating tank model of the northern hemisphere, complete with realistic continental configurations (Fig. 15). In the 1950's, von Arx (1952, 1957) used this same apparatus to experimentally reproduce the major features of the present oceanic circulation. Variations

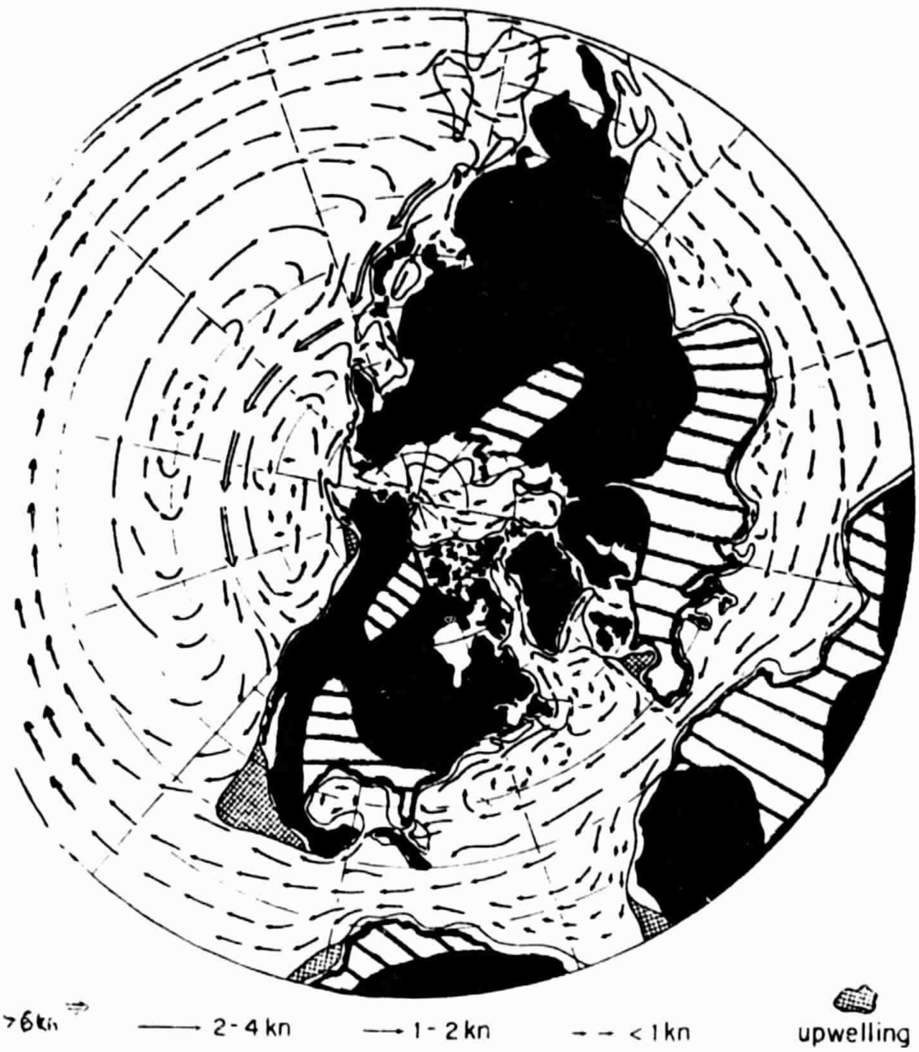

Fig. 15 Vector interpretation of the nonglacial ocean circulation at 100 M.y. Double arrows indicate currents greater than 6 knots. Upwelling regions are indicated by cross-hatching. Diagonal hachures indicate continental regions flooded by shallow seas as mapped by Barron et al (1980). (From Luyendyk et al, 1972; reproduced with the permission of the authors and the Geological Society of America).

in planetary vorticity are simulated with a variable-depth floor in the tank (poleward movement causes vortex shrinking). Winds are simulated by pumping air through an array of copper tubes suspended over the dishpan. The westerlies are displaced 10° of latitude poleward of their present position, with velocities the same as at present. Adjustments to wind velocities affected ocean current velocities, but did not substantially alter patterns.

Results indicate: 1) the Gulf Stream penetrated to the latitude of Newfoundland before turning eastward towards Europe; because the Norwegian and Labrador Seas had not yet opened, warm water could not penetrate to the pole by means of the Atlantic; 2) a circum-equatorial zonal current flowed through an equatorial seaway (Tethys Sea) from Malaysia to Gibralter; the current continued eastward across the Atlantic, through an open Isthmus of Panama, and across the Pacific; 3) in the Pacific, the Kuroshio Current was stronger, and the subpolar gyre was similar to that which occurs there today. This gyre should have effectively blocked advection of warm Pacific waters to the Arctic Ocean. Thus, a surface circulation reconstruction cannot produce warm currents penetrating to the North Pole.

The Cretaceous deep-water circulation was probably different from the present. In addition to warmer deep waters, the turnover rate of the deep waters was also occasionally low: Deep-Sea Drilling Project results from the Atlantic, Pacific, and Indian Oceans reveal widespread occurrences of Cretaceous black shales. The coloration indicates large amounts of organic matter, which in turn probably reflects anoxic conditions in old bottom water. In fact, Ryan and Cita (1977) have estimated that the organic carbon content of the shales exceeds the organic carbon content of all known coal and hydrocarbon reservoirs.

Kraus (1978) has suggested that a feasible mechanism for transporting heat poleward is via warm, saline deep-water plumes. Today, such features form in the Mediterranean, contributing water to the upper levels of the North Atlantic Deep Water. Some of the Mediterranean water can be traced along density surfaces that shallow to the north and outcrop in the Norwegian Sea (Reid, 1979). At present, the volumetric output of Mediterranean water is 1×10^6 m^3/s (Worthington, 1976), which is only about 1/30 the average transport of, for example, the Florida Current (Niiler and Richardson, 1973). Kraus (1978) has suggested that a larger input rate of saline water could substantially affect the amount of heat transported poleward. A broad Tethys Sea, with adjacent flooded regions of Europe and Africa (Fig. 15), would be an ideal locus for the formation of the highly-saline waters. Preliminary calculations suggest that an ice-free Arctic could be maintained by such a process. More detailed investigations are now in progress on this intriguing possibility.

In conclusion, the warmer temperatures of the Cretaceous are partly due to changes in albedo and land-sea distribution. Despite the sizeable contributions of paleogeographic factors, simple energy balance models still cannot account for the global distribution of Cretaceous temperatures (Barron et al, 1981). Clouds may have played a more important role in modifying the temperature patterns. Changes in surface and deep-ocean circulations may also have been important. At present the problem is unresolved.

A final, perplexing feature of non-glacial climates involves the discrepancy between insolation receipt at high latitudes and the distribution of fauna and flora north of the Arctic Circle (Donn and Shaw, 1977; Donn, 1981). The most reliable evidence pertaining to this phenomenon actually comes from a rich fossil site from the Eocene (about 50 M.y.) of Ellesmere Island, west of Greenland (Dawson et al, 1976). The present latitude of the fossil site is about 78°. Paleomagnetic results indicate the paleolatitude of this site was also about 78° (McKenna, 1980). Some of the vertebrates from the site include alligators and "flying lemurs", the latter occurring today only in southeast Asia (Estes and Hutchinson, 1980; McKenna, 1980). There is some question as to whether the broadleaf tropical plants at this site could have survived under an insolation regime similar to today (Wolfe, 1980; Donn, 1981). In other words, even if polar temperatures were in fact warmer during non-glacials, the year-round insolation receipt might still have been insufficient to support the type of photosynthetic activities that tropical plants are adapted to. This objection may not impress a geophysicist, but biologists would point out that photosynthetic mechanisms ground into the genes at fundamental levels are unlikely to radically transform even over periods of 20-40 million years. Thus, although it may be possible to re-tune climatic models to produce more heat at high latitudes, no amount of re-tuning can produce more light in the winter months. Resolution of this paradox may well provide important information for reasonable physical models of nonglacial climates.

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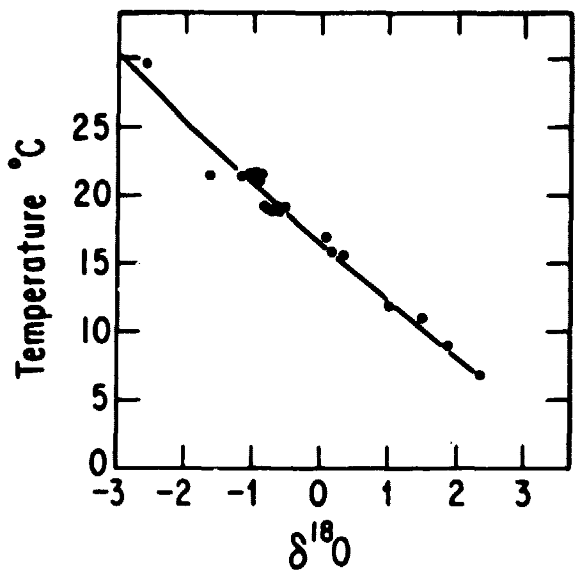

Fig. 16 Temperature of growth of mollusk shells vs. 0-18/0-16 ratio in the shells. This figure illustrates that measured variations in past 0-18 ratios have the potential for providing information about paleotemperatures. (From Epstein et al, 1953; reproduced with the permission of the Geological Society of America).

VI. LATE CRETACEOUS - LATE CENOZOIC (100-1.7 M.y.)

It is an inevitable feature of geologic data that the younger the rocks, the higher the data quality. Increased quality results in model revisions. The period 100-1.7 M.y. is a classic example of an interval that has witnessed a large increase of new, high-quality data. As a result, there are at present two interpretations of the climatic history of the interval, and resolution of the differences must await more data. Consequently, this section will present two summaries of climate history, 100-1.7 M.y.

A fundamental tool in both interpretations involves a technique that has been an invaluable aid in examining the climatic history of the last 100 M.y. — $^{18}O/^{16}O$ stable isotope analysis. In 1947, Harold Urey proposed that carbonates precipitated from different water temperatures should have different ratios of $^{18}O/^{16}O$. Epstein et al (1953) verified that theory with laboratory mollusks grown at different temperatures (Fig. 16). A 1.0 0/00 increase in ^{18}O results from a 4°C decrease in water temperature, or a 0.4 0/00 increase in salinity. Shackleton (1974; Shackleton and Opdyke, 1973) greatly expanded the usefulness of the method by developing a high-precision, small-sample, rapid analytical technique. The reader is referred to Duplessy (1978) for a recent discussion of the method.

The most commonly used fossils for ^{18}O analyses are foraminifera (Fig. 17) - single-celled amoeba-like organisms with $CaCO_3$ shells. There are both planktonic (surface and near-surface) and benthonic (bottom dwelling) varieties, and thus both surface and bottom paleotemperatures can be recorded. The planktonic variety has been a major constituent of deep-sea sediment for over 100 M.y.

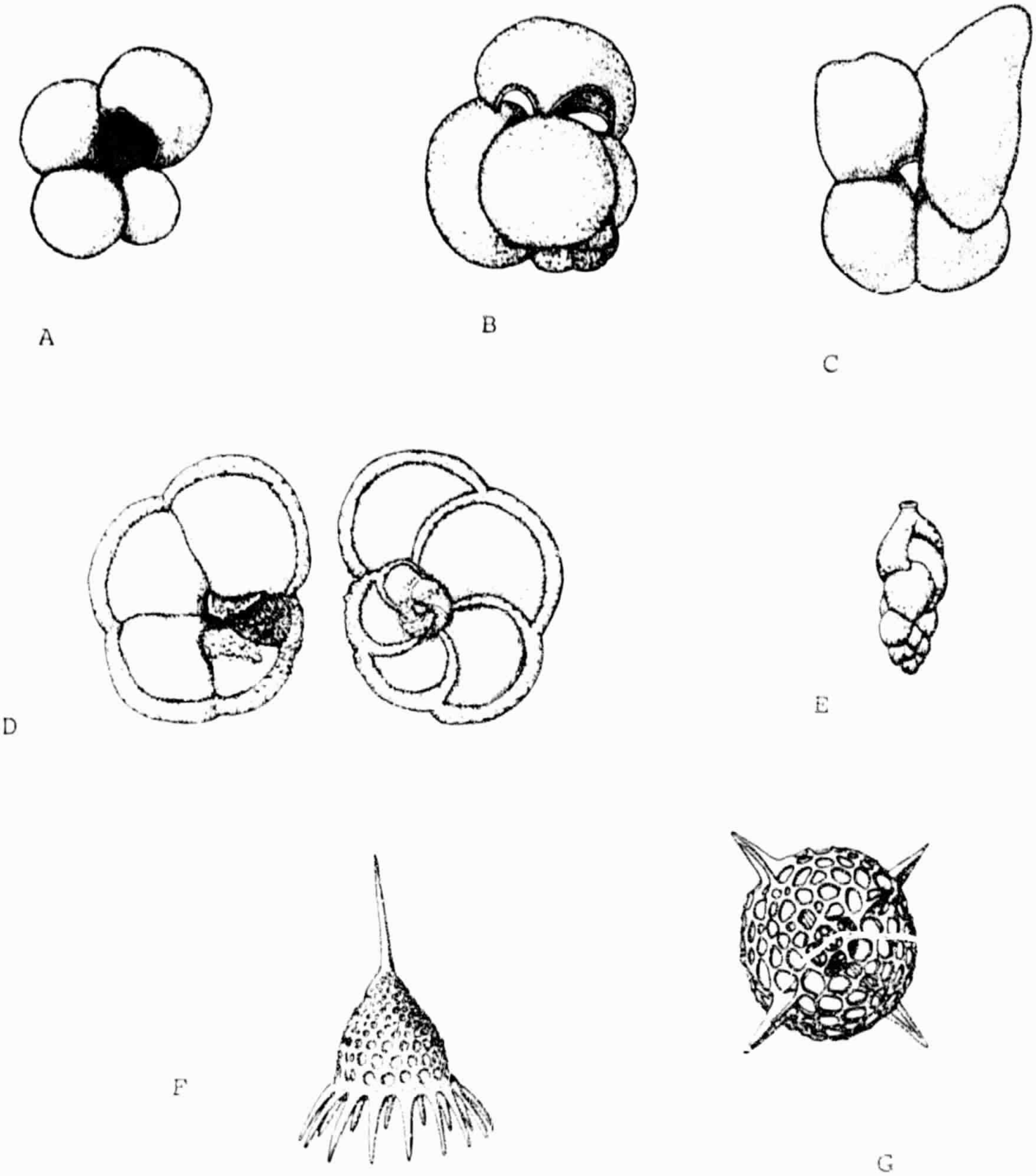

Fig. 17 Common microfossils in deep-sea sediments. A-D: planktonic foraminifera. A: Globigerina bulloides, 43X; B: Globigerinoides ruber, 64X; C: Globigerinoides sacculifer, 35X; D: Globorotalia menardii, 35X; D: benthonic foraminifera Uvigerina, 50X; F & G: siliceous radiolaria. F: Anthocyrtium, 140X; G: Hexalonche, 200X. Species A-C & E are commonly used for oxygen isotope analyses. Sources: A-D (Parker, 1962); E-G (Easton, 1960). (Figures E-G reproduced with permission of W. H. Easton.)

The fundamental geological problem associated with ^{18}O interpretation involves the realization by Emiliani (1955) that the ^{18}O content of seawater can also be affected by the formation of ice sheets. Due to the differences in atomic mass of ^{16}O and ^{18}O, seawater is preferentially enriched in ^{18}O by evaporation processes that supply H_2O for ice sheets. Thus, past ^{18}O changes can be due to either temperature, salinity, or ice volume changes. Because of the conservative properties of salinity, temperature and ice volume represent the two variables most susceptible to fluctuation. A great deal of effort has been expended in the last 25 years in separating the two effects. Shackleton (1967) succeeded in achieving this goal for the Pleistocene (see next section). However, failure to definitively isolate the ice volume effect for pre-Pleistocene sediments has resulted in two interpretations of climate, 100-1.7 M.y. Since the models diverge primarily for data covering the last 40 M.y., the inferred history of the preceding interval will first be discussed.

The principal climatic results of the interval 100-40 M.y. are as follows (Fig. 18): a cycle of mild cooling followed by warming covered the interval 100-50 M.y. (e.g., Savin, 1977). Bottom water temperatures were as high as 15°C as late as 50 M.y. (Shackleton and Kennett, 1975a). Shackleton (1978) notes that Reid and Chandler (1933) report a rich tropical flora from London Clay samples of this age. Soil profiles originating under tropical conditions extended to 45° paleolatitude in both hemispheres (Frakes, 1979).

The interval 100-40 M.y. also contains evidence of an apparent encounter with a large Earth-crossing asteroid (Alvarez et al, 1980). The now-famous impact occurred at the Cretaceous-Teriary boundary (65 M.y., cf. Fig. 2), and was associated with a longer period of time that witnessed the extinction of

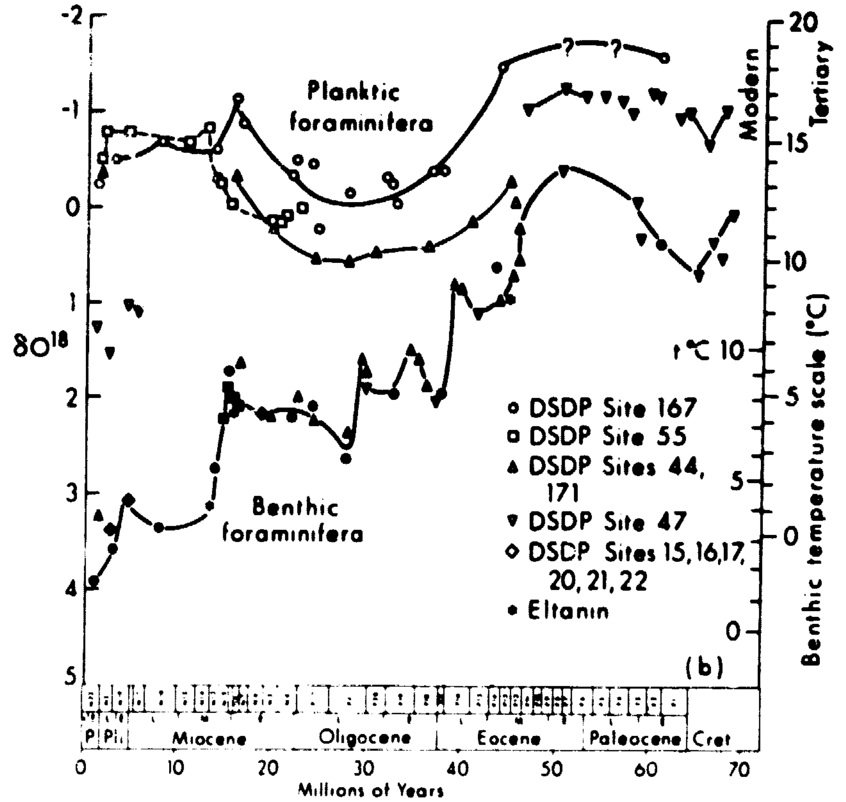

Fig. 18 Variations of surface and bottom-water O-18 ratios in the North Pacific for the last 70 M.y. In one model of Cenozoic climate, the pre-14 M.y. O-18 changes are interpreted in terms of temperature changes (see text). (From Savin, 1977; reproduced, with permission, from Annual Reviews of the Earth and Planetary Sciences, v. 5. Copyright 1977 by Annual Reviews Inc.).

an estimated half of all living species (Russell, 1979). The event presumably injected a large amount of dust into the upper atmosphere. Since analyses of historical data (e.g., Mass and Schneider, 1977) indicate a correlation between increased volcanic dust and decreasing temperatures, it would be informative to determine the climatic response associated with a very large dust veil event. Surprisingly, ^{18}O analyses indicate that, if anything, the Cretaceous-Teriary boundary was associated with a slight warming of ocean waters (Boersma et al, 1979).

Cenozoic Glaciation - Model I The main feature of Model I involves the conclusion that the Antarctic Ice Sheet formed about 12-15 M.y. In order to place this conclusion in perspective, the ^{18}O record from 40-15 M.y. must be discussed. A dramatic ^{18}O change at 38 M.y. (Fig. 18) may have marked the development of a more vigorous, colder deep-water circulation as a result of Antarctic Bottom Water Formation (Kennett and Shackleton, 1976). This water mass is one of the two main water masses in the world ocean (Sverdrup et al, 1941), and presently forms primarily (e.g., Gordon, 1966) in the Weddell Sea (see below, Figure 22). Closely-spaced core samples (Fig. 19) indicate that the cooling may have taken place in less than 100,000 years (Kennett and Shackleton, 1976). Kennett et al (1974) cite geological evidence that suggest the separation of Australia from Antarctica (Fig. 20) at about this time may have contributed to southern cooling by inhibiting poleward transport of warm ocean currents (Frakes and Kemp, 1972). The Drake Passage between Antarctica and South America appears to have opened later, at perhaps about 30 M.y. (Barker and Burrell, 1977). At this time, the Antarctic Circumpolar Current was established, and the thermal isolation of Antarctica completed.

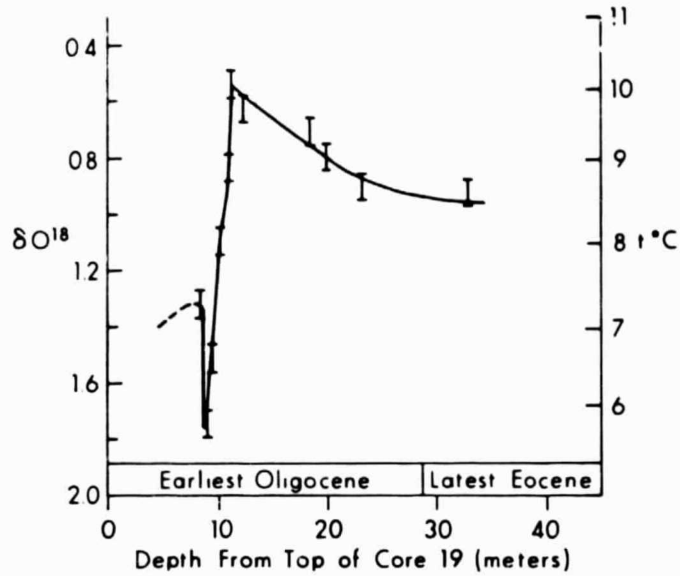

Fig. 19 Evidence for a very rapid change in bottom water 0-18 ratios in the Lower Oligocene (about 38 M.y.). In one model of Cenozoic climate, this result is inferred to indicate the development of a more vigorous deep-water circulation at this time. (From Kennett and Shackleton, 1976; reproduced with the permission of the authors and NATURE).

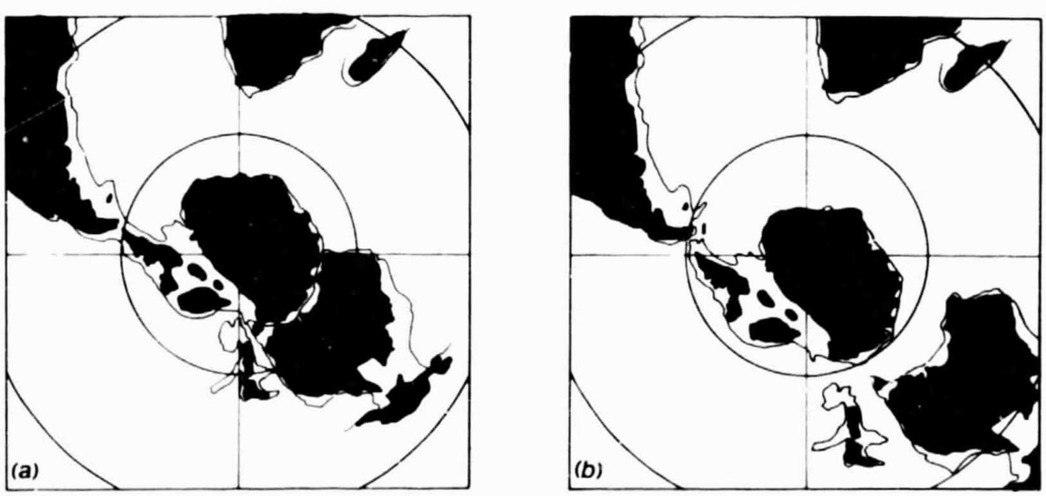

Fig. 20 Paleogeographic evolution of the Southern Hemisphere during part of the Cenozoic. (a) Eocene (about 50 M.y.): (b) Oligocene (about 35 M.y.). The breakup of the continents in the Southern Hemisphere is thought to have had a profound effect on the circulation of the Antarctic Circumpolar Current, and on the climate of Antarctica. (From Tarling, 1978, in Climatic Change; reproduced with the permission of Cambridge Univ. Press).

With the establishment of the Antarctic Bottom water, Shackleton and Kennett (1975a) reasoned that the next significant change in ^{18}O at 12-15 M.y. should indicate the formation of the Antarctic Ice Sheet. (Other geological data, discussed below, indicated that Northern Hemisphere glaciation did not commence until about 3.0 M.y.)

Cenozoic Glaciation - Model II The principal feature of Model II involves a re-interpretation of the low-latitude isotope data illustrated in Figure 18. Matthews and Poore (1980) note that the planktonic (surface) ^{18}O record 30-40 M.y. also varies significantly in this record. Since CLIMAP (1976) results imply that low-latitude sea-surface temperatures varied little even during the last glacial maximum, then the Cenozoic planktonic ^{18}O excursions may be of ice-volume origin. This interpretation implies significant ice volume for at least the last 40 M.y. Matthews and Poore suggest that ice-volume related sea level changes may be in part responsible for large sea level fluctuations during that time period as mapped by seismic techniques (Fig. 21). Their interpretation is also supported by glacial geologic work on Antarctic, which suggests that the Antarctic Ice Sheet may have been at least as large as at present at 30 M.y. (LeMasurier, 1982).

The present Antarctic Ice Sheet (Fig. 22) consists of both East and West Antarctic Ice Sheets, separated from each other by the Transantarctic Mtns. The East Antarctic Ice Sheet is an order of magnitude larger than the West Antarctic Ice Sheet (Denton et al, 1971). Together, the ice masses comprise a volume equivalent to 55 m of sea level (Denton et al, 1971). Glacial studies

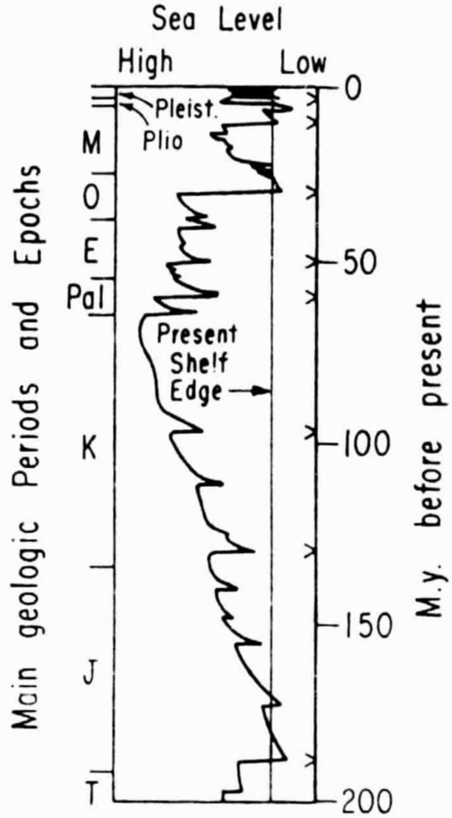

Fig. 21 Inferred sea-level curve for the last 200 Million years. Curve is based on interpretation of seismic stratigraphic horizons from continental margins. Matthews and Poore (1980) have proposed that significant sealevel fluctuations during the last 200 M.y. may be due in part to ice volume fluctuations. (After Vail et al, 1977; reproduced with the permission of the American Associations of Petroleum Geologists).

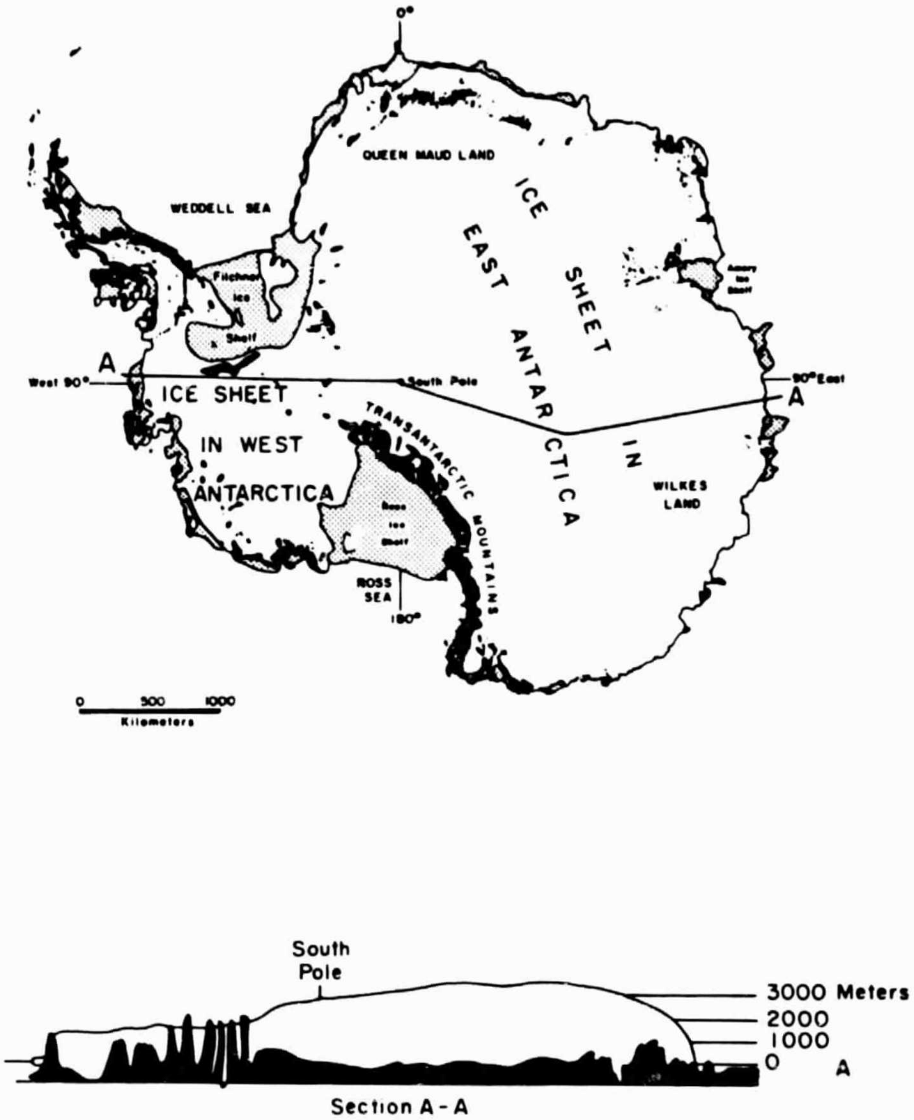

Fig. 22 Antarctic Ice Sheet. Ice shelves indicated in stipple; exposed bedrock shown in black (after Bentley, 1965). (From Denton et al, 1971, in The Late Cenozoic Glacial Ages; copyright 1971 by Yale Univ. Press).

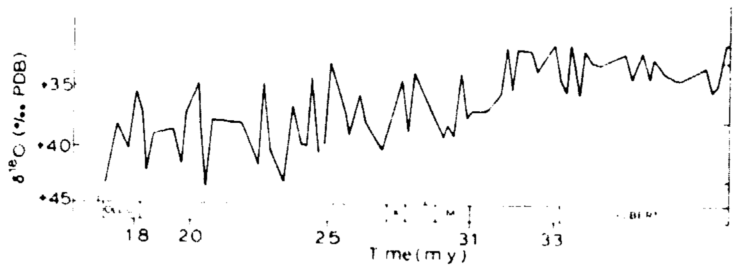

Fig. 23 Oxygen isotope evidence for the initiation of Northern Hemisphere glaciation. Fluctuations of 0-18 beginning about 3.1 M.y. are inferred to indicate fluctuations of Northern Hemisphere ice sheets. (From by N. J. Shackleton and N. D. Opdyke, 1977. Reproduced with the permission of the authors and NATURE).

in ice-free valleys of the Transantarctic Mtns. record times when the East Antarctic Ice Sheet was larger than present (Denton et al, 1971): thickening of the ice sheet caused spillover of glaciers into the dry valleys, and interbedded basalt flows provide radiometric dates for the ice advances. From these and other studies, it is known that the Antarctic Ice Sheet was at least as large as at present prior to 4.0 M.y. (Denton et al, 1971). As previously noted, one study suggests a large ice sheet as early as 30 M.y. (LeMasurier, 1982). Some studies suggest that between 4.0-5.5 M.y. the ice sheet may have been as much as 50% larger than present (Shackleton and Kennett, 1975b; see also Frakes, 1979): the Ross Ice Shelf (Fig. 19) expanded to the edge of the continental shelf (Hayes and Frakes, 1975), and there was a strong northward penetration of the Antarctic Circumpolar Current (Kemp et al, 1975).

Northern Hemisphere Glaciation The history of northern hemisphere climate must be considered in two parts - evidence for cooling, and evidence for glaciation. Analyses of North Atlantic and North Pacific records indicate some pronounced temperature decreases that extend back to at least 10 M.y. (Keller, 1979, 1980; Poore, 1981). Additionally, the Arctic Ocean has been ice-covered for at least the last 5 M.y. (Clark, 1971; Clark et al, 1980).

Northern hemisphere glaciation was apparently initiated several million years after Antarctic glaciation. The present best estimate is about 3.0 M.y. ago. This figure is based on distribution of ice-rafted detritus in the North Pacific (Kent et al, 1971) and Labrador Sea (Berggren, 1972), tills in Iceland (McDougall and Wensink, 1966), and ^{18}O (Shackleton and Opdyke, 1977). North Atlantic Deep Water production also may have increased at this time (Barker et al, 1981). Shackleton and Opdyke (1977) also show that between 2.5-3.0 M.y., closely-spaced, large-scale oscillations were initiated (Fig. 23). This pattern has persisted to the present day. Ice volume estimates indicate maxima about two-thirds of those for the Late Pleistocene.

Although the causes for northern hemisphere ice initiation cannot be established with confidence, the study of ocean circulation models (Luyendyk et al, 1972; cf. Fig. 15) suggests one possibility. Prior to 3.0 M.y., experiments conducted with an open Isthmus of Panama indicate that there was a throughflow of warm Gulf Stream water from the North Atlantic through the Arctic to the Pacific. With closure of the Isthmus, the current withdrew to the present position in the Norwegian Sea. Luyendyk et al (1972) note that the geological evidence for Central American uplift coincides closely with the initiation of northern hemisphere ice cover.

Prospectus - The above models of pre-Pleistocene climatic fluctuations may be subjected to more rigorous testing and revision in the next few years, with the principal agent of change being the newly developed hydraulic piston core (see, e.g., LaBrecque, 1981). This device can be fitted to the rigs of the deep-sea drilling ships Glomar Challenger or Glomar Explorer, and for the first time enables recovery of undisturbed sequences of soft-sediment between the surface and the hard-rock layer, which is already retrievable by means of conventional rotary-drilling techniques. High-quality continuous records that extend millions of years beyond the previous limit of about 2 M.y. will soon be available for paleoclimatic analysis, and prospects for model revision are excellent.

VII. PLEISTOCENE (1.7 - 0.01 M.y.)

Although comprising less that 0.1 % of Earth history, the Pleistocene Epoch has been studied more intensively than the rest of the paleoclimatic record. This imbalance is due to the juxtaposition of Pleistocene climates with the present Holocene interglacial climate, relatively easy access to a wide variety of geological exposures, and amenability of the data to more rigorous, quantitative analysis. In keeping with this ratio, most of the remaining part of the text will examine techniques and results from the Pleistocene and Holocene Epochs. Together, these epochs comprise the Quaternary Period (see Fig. 24 for an illustration of the most common Quaternary terms). The Pleistocene section will be sub-divided into six subjects: a) development of a Pleistocene chronology; b) methods of faunal analysis; c) results of the 18,000 YBP (years before present) CLIMAP Project; d) temporal trends of some Pleistocene records; e) correlations with orbital perturbations; f) models of glacial inception.

A. Pleistocene Chronology

The development of a Pleistocene timescale is perhaps best understood when viewed from an historical perspective of the entire Pleistocene discipline. Early studies in North America and western Europe suggested four major episodes of continental glaciation. The "multiple glaciation hypothesis" was based on evidence such as the interbedding of glacial tills with deeply-weathered soils, which required long intervals of interglacial climate in order to develop. The chronology of the events was essentially

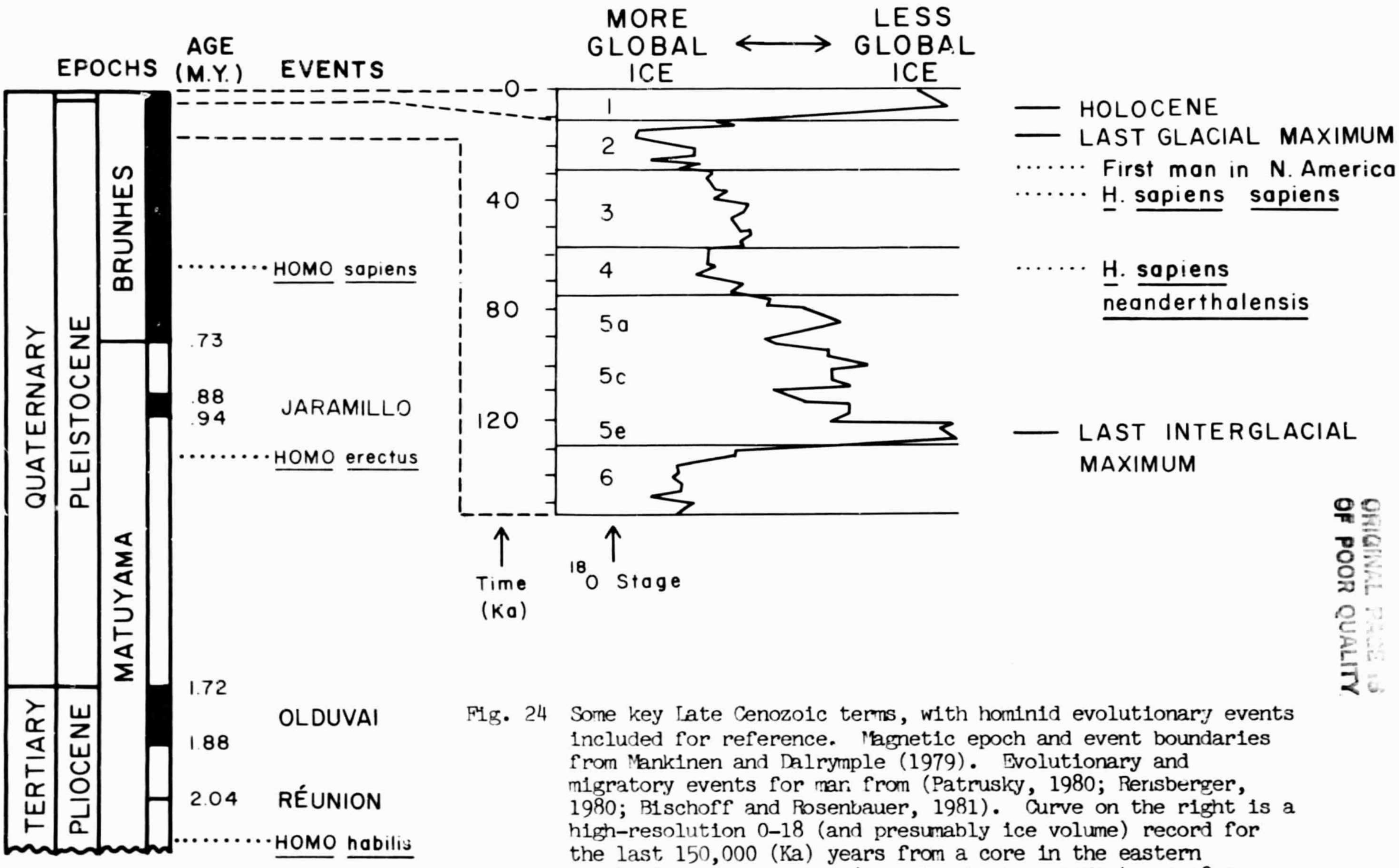

Fig. 24 Some key Late Cenozoic terms, with hominid evolutionary events included for reference. Magnetic epoch and event boundaries from Mankinen and Dalrymple (1979). Evolutionary and migratory events for man from (Patrusky, 1980; Rensberger, 1980; Bischoff and Rosenbauer, 1981). Curve on the right is a high-resolution 0-18 (and presumably ice volume) record for the last 150,000 (Ka) years from a core in the eastern tropical North Atlantic (from Shackleton, 1977a). 0-18 Stages are from Emiliani (1955); substages from Shackleton (1969). Present best estimates of isotope stage boundaries are 11, 27, 58, 72 and 128,000 YBP (Morley and Hays, 1981). Boundaries at 11,000 and 128,000 YBP are Terminations I & II (Broecker and van Donk, 1970). The interval between these two terminations is known as the last glacial cycle.

unknown beyond the range of ^{14}C dating (about 50,000 years on land). In the mid-1950's, Cesare Emiliani, a student of Harold Urey's, greatly modified this picture by application of the oxygen isotope technique to deep-sea sediments (see Section V). In many cases, the sediments are comprised chiefly of the remains of microscopic planktonic organisms. Their near-continuous accumulation on the seafloor offered the potential for a continuous record of climatic change spanning several hundred thousand years. Emiliani (1955) discovered evidence for many more glaciations than had previously been anticipated. The continental record should therefore be considered a fragmentary history of glacial fluctuations, with detailed correlations over wide areas often highly suspect beyond the range of ^{14}C dating (see Kukla, 1977). Because of these uncertainties, workers who are not actively involved in Quaternary land geology are advised to refrain from using <u>any</u> proper names of continental glacial events that pre-date 50,000 YBP.

Although Emiliani made many valuable contributions to concepts of ice-age climates, more than fifteen years passed before two critical questions concerning the ^{18}O record could be successfully interpreted: 1) how much of the ^{18}O changes were due to sea-surface temperature (SST) changes, and how much were due to the concentration of isotopically light ^{16}O in continental ice sheets during glacial advances? 2) what was the absolute chronology of the glacial fluctuations? This latter problem had to be approached circuitously, for direct dating of deep-sea sediments could not in itself provide a satisfactory chronology: ^{14}C dating of marine carbonates is generally unreliable beyond about 30,000 YBP. Uranium-series techniques can only be utilized to get approximate age estimates, because the methods are beset with questionable assumptions concerning constancy of sedimentation rates (e.g., Broecker, 1974).

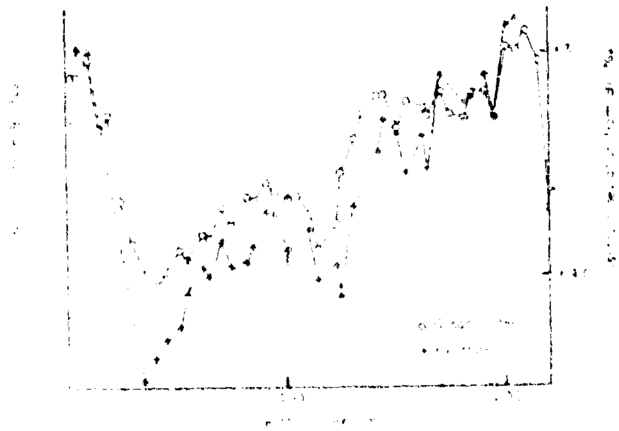

Fig. 25 Oxygen isotope evidence for the dominant control of ice volume fluctuations on the 0-18 record. Measurements are of surface and bottom-dwelling foraminifera in the same core. The two sequences are plotted to the same scale of isotopic change, but with scale zero-points differing by 5.3 0/00 (the present-day surface-bottom difference). The approximately equivalent magnitudes of change imply the same origin for both records. (From Shackleton and Opdyke, 1973; reproduced with the permission of QUATERNARY RESEARCH).

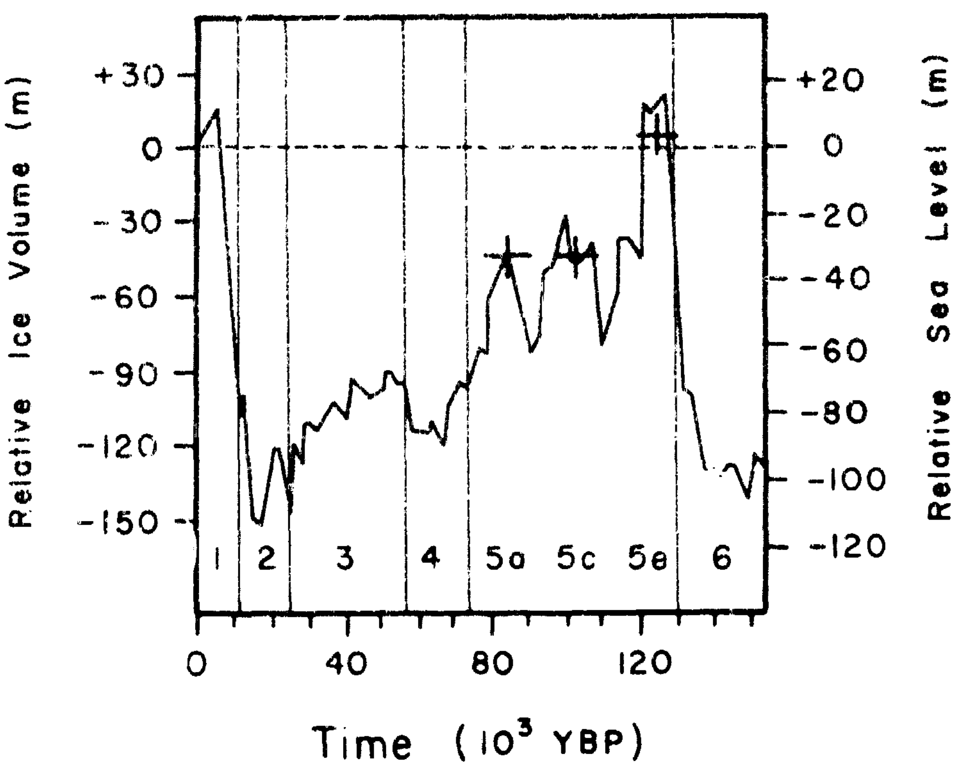

Fig. 26 Ice volume fluctuations for the past 130,000 years derived from oxygen isotope measurements in MD73125, compared with sea level estimates (crosses) derived from work on Barbados, West Indies. Note that, except for oceanic islands, the ice volume effect does not equal the sea level curve (isostatic rebound on continental shelfs reduces the effect by about 24%). Sea level estimates, based on coral reef terraces, from Mesolella et al (1969) and Fairbanks and Matthews (1978). 0-18 record from Shackleton (1977a).

The first problem was resolved by two different approaches — ^{18}O analysis of benthonic foraminifera in deep-sea sediments (Shackleton, 1967; Shackleton and Opdyke, 1973), and development of a transfer function to estimate SST from foraminiferal abundances (Imbrie and Kipp, 1971). Discussion of the latter approach will be deferred to the methods section. The ^{18}O analysis of benthonic foraminifera provided the first indication that most of the isotopic change in seawater was due to the "ice volume" effect (Fig. 25). The reasoning is as follows: since bottom water temperatures vary by only a small amount and are near their minimum value, glacial ^{18}O changes in bottom-dwelling benthonic foraminifera must primarily reflect changes in continental ice volume. Global synchroneity of isotopic changes is then only constrained by the mixing time of the oceans (about 1000 years - Broecker, 1963). The ice volume effect of about 1.6 0/00 accounts for about 80% of the change in many planktonic ^{18}O records. The above approach yielded similar results when ^{18}O variations in surface-dwelling foraminifera were measured in a core from a region with a stable SST history (Crowley and Matthews, 1982).

Development of an absolute chronology for deep-sea cores has received impetus from two different approaches, one that focuses on the Late Quaternary, the other that addresses the timing of older events. In a study of uplifted coral reef terraces on Barbados, Mesolella et al (1969) utilized uranium-series dates on isotopes trapped in the carbonate crystal lattice to determine that the ages of some interglacial-reef formations centered on 82, 105, and 125,000 YBP. ^{18}O analyses of corals and mollusks in the reefs were then used to match the patterns of isotope ratios with the deep-sea ^{18}O record (Shackleton and Matthews, 1977; Fairbanks and Matthews, 1978). For example, the samples in the 125,000 YBP terrace consistently had the lightest ^{18}O

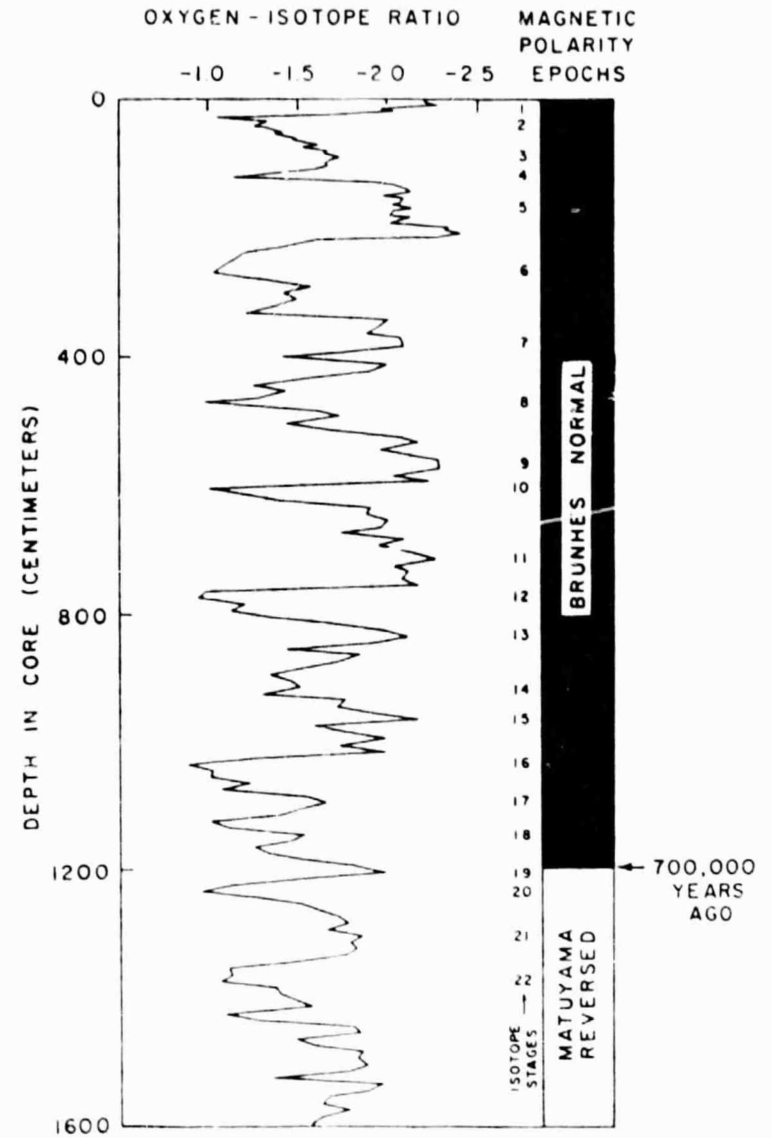

Fig. 27 A graph of the isotopic and magnetic measurements made in 1973 by N. J. Shackleton and N. D. Opdyke on a Pacific deep-sea core (V28-238). These observations--which established that isotope Stage 19 occurs at the boundary between the Brunhes and Matuyama Epoches--provided the first accurate chronology of Late Pleistocene climate. (From ICE AGES Solving the Mystery, by John Imbrie and Katherine Palmer Imbrie, Enslows Pubs., Box 777, Hillside, New Jersey, 07205.)

values (i.e., most negative). These samples correspond to ^{18}O Stage 5e in the deep-sea record (cf. Fig. 24). ^{18}O Stage 5e therefore became a critical tie-point in the development of an absolute chronology.

The clarification of the chronology for the last 125,000 yrs. has enabled calibration of the ^{18}O record in terms of absolute ice volume changes. For example, various techniques have been used to estimate that a volume of water equivalent to approximately 150 m of sea level was evaporated to construct the ice sheets (Hughes et al, 1981). Utilizing an average glacial-interglacial ^{18}O range of 1.6 0/00 (Shackleton, 1977b), every change of 10 m in sea level produces a 0.10 0/00 change in ^{18}O. The method predicts ice volume changes during ^{18}O Stages 5a, 5c, & 5e (Fig. 26) of - 45m, - 45m, and + 15m (with respect to the present). These values agree well with estimates of - 45m, - 43m, and + 5m from coral reef terraces on Barbados (Fairbanks and Matthews, 1978).

Paleomagnetic studies have provided the key to age interpretation of Pleistocene sections older than 150,000 YBP. The pattern of magnetic reversals in sediments is correlated to the continental pattern, where absolute ages for reversal boundaries are determined by potassium-argon dating of land basalts. Shackleton and Opdyke (1973) were the first to determine that the last major reversal, the Brunhes/Matuyama boundary, occurred within ^{18}O Stage 19 (Fig. 27). Correlations were later extended to the entire Pleistocene (Shackleton and Opdyke, 1976). Since sedimentation rates in deep-sea cores are often nearly constant, age boundaries of isotopic stages could be linearly interpolated from magnetic reversal boundaries. The stage was now set for spectral analysis of the paleoclimatic indices. Before those results are discussed, however, it is useful to review some of the techniques of faunal analysis.

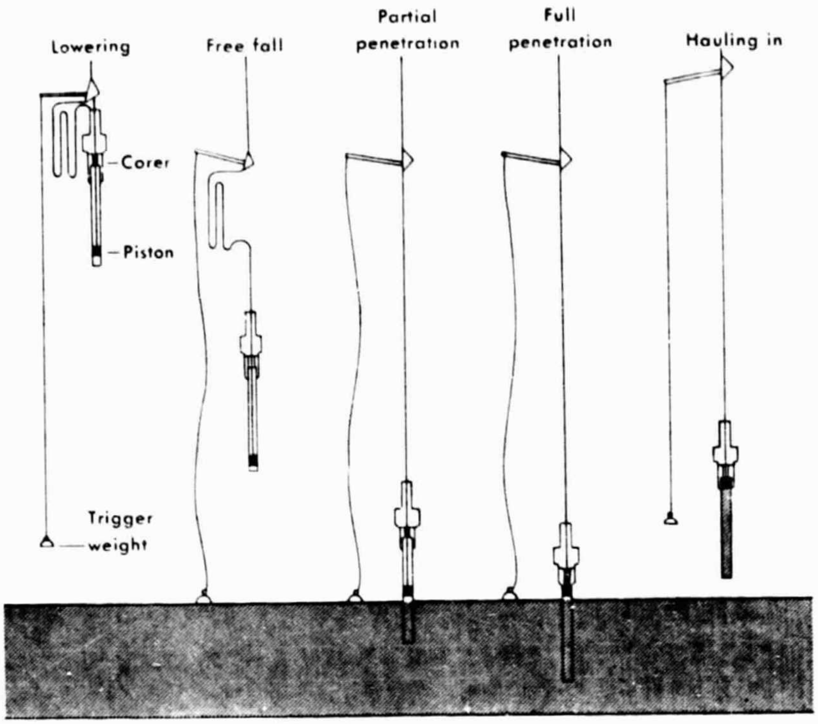

Fig. 28 Operation of the piston core. (From U. S. Naval Oceanographic Office Pub. 607, 1968).

B. Methods of Faunal Analysis

This section discusses three aspects of data analysis: 1) retrieval and processing of raw core samples; 2) statistical techniques of data manipulation; 3) problems of interpreting paleoclimatic results.

Retrieval and processing of raw samples This section recounts typical sampling techniques developed at the Lamont-Doherty Geological Observatory during the past 25 years. The Ewing piston core (Fig. 28) is a common device used for the recovery of deep-sea sediments. Cores of 8-cm diameter and 5-20 m length can be taken with relative ease. The core contains an interior, snugly-fitting piston and is fitted with a 900 kg weight and a small companion trigger weight core. When lowered from the ship, the trigger weight core extends about 5-m below the bottom of the piston core. Near the seafloor the apparatus is slowly lowered until the trigger weight penetrates the sediment to a depth of 10-40 cm. The resultant slack on the wire leading to the main core triggers a free fall, with the core barrel sliding past the piston as the barrel penetrates the sediment. A core catcher, consisting of light flexible tangs, prevents the sediment from sliding out during the process of hauling-in. The core is described and then returned to shore for storage. Half of the core is preserved as an archive record, and half is available for sampling. The present inventory at Lamont numbers over 12,000 cores from virtually every part of the world ocean.

The following discussion recounts a typical CLIMAP prescription for sampling planktonic foraminifera. Cores are sampled at intervals of 2-, 5-, or 10-cm. Sample sizes are typically about a cubic centimeter (2-3 gm). The dried sediment sample is disaggregated in water and sieved, with different size fractions preserved for different studies. The fraction > 149 μ commonly contains several thousand specimens of adult planktonic foraminifera. A sample splitter is used to pare the sample down to 300-400 specimens. Statistical tests have shown that a sample this size yields reproducible results at the 80% confidence level (Imbrie et al, 1973).

Specimens strewn on microscope slides are identified and counted. There are about 30 species of planktonic foraminifera that inhabit the surface waters of the world ocean. The census of species are then listed in tabular form. Each row represents a different sample, either from a different core depth or from a geographical array of surface samples. Each column represents a different variable (species of foraminifera). Elements of the data matrix therefore represent frequency of a particular species in an individual sample. The resultant rectangular matrix is now in a form suitable for statistical analysis.

Statistical Techniques This section discusses how environmental estimates are extracted from paleontological data. The procedure involves three steps: 1) description of floral/faunal ecological patterns in terms of end-member samples; 2) calibration of end-member samples against environmental variables (e.g., sea-surface temperature, SST); 3) estimation of paleoenvironmental values from paleoecological distributions of flora and fauna. The example in this section discusses how SST-records are obtained from the geographic distribution of planktonic foraminifera in 191 North Atlantic surface sediment samples (Kipp, 1976). However, the method has much broader applicability, and can be used for other biological groups (e.g., plants - Webb and McAndrews, 1976) and other physical variables (e.g., precipitation and ocean salinity - Webb and Bryson, 1972; Cullen, 1981). The reader is referred to Sachs et al (1977) for a more complete discussion of this subject.

1. Description of data in terms of end member samples. As discussed in the previous section, there are about 30 species of planktonic foraminifera in the world ocean. The species are not uniformly distributed. Some live in tropical waters, others in polar waters. The different species therefore tend to cluster into a smaller number of ecological assemblages (tropical, polar, etc.). An analogy can be drawn from the distribution of land plants -- e.g., tropical (rain forest), temperate (broadleaf deciduous), savannah (oak - grasses), or boreal (spruce - pine). This nearly-universal clustering tendency of organisms allows for a more economical description of sample distributions. Rather than describing each geographical sample separately, all samples can be characterized by a few average, end-member samples (tropical, subtropical, etc.) A standard procedure for objectively identifying the end members involves some form of reduction in dimensionality (e.g., principal component analysis, factor analysis). These methods have also been widely used in meteorology and oceanography (e.g., Kutzbach, 1967; Weare et al, 1976). The vectors in the principal coordinate matrix are the desired average end members, and are determined by computing the eigenvalues of the raw data matrix.

The standard starting point for the analysis involves a rectangular data matrix such as that described in the previous section. The matrix for this example represents a geographic array of North Atlantic surface sediment samples (core tops, ct). Utilizing the symbols of Imbrie and Kipp (1971), the technique describes a row-normalized rectangular data matrix (U_{ct}) as a product of two matrices

$$U_{ct} = AF + E \tag{3}$$

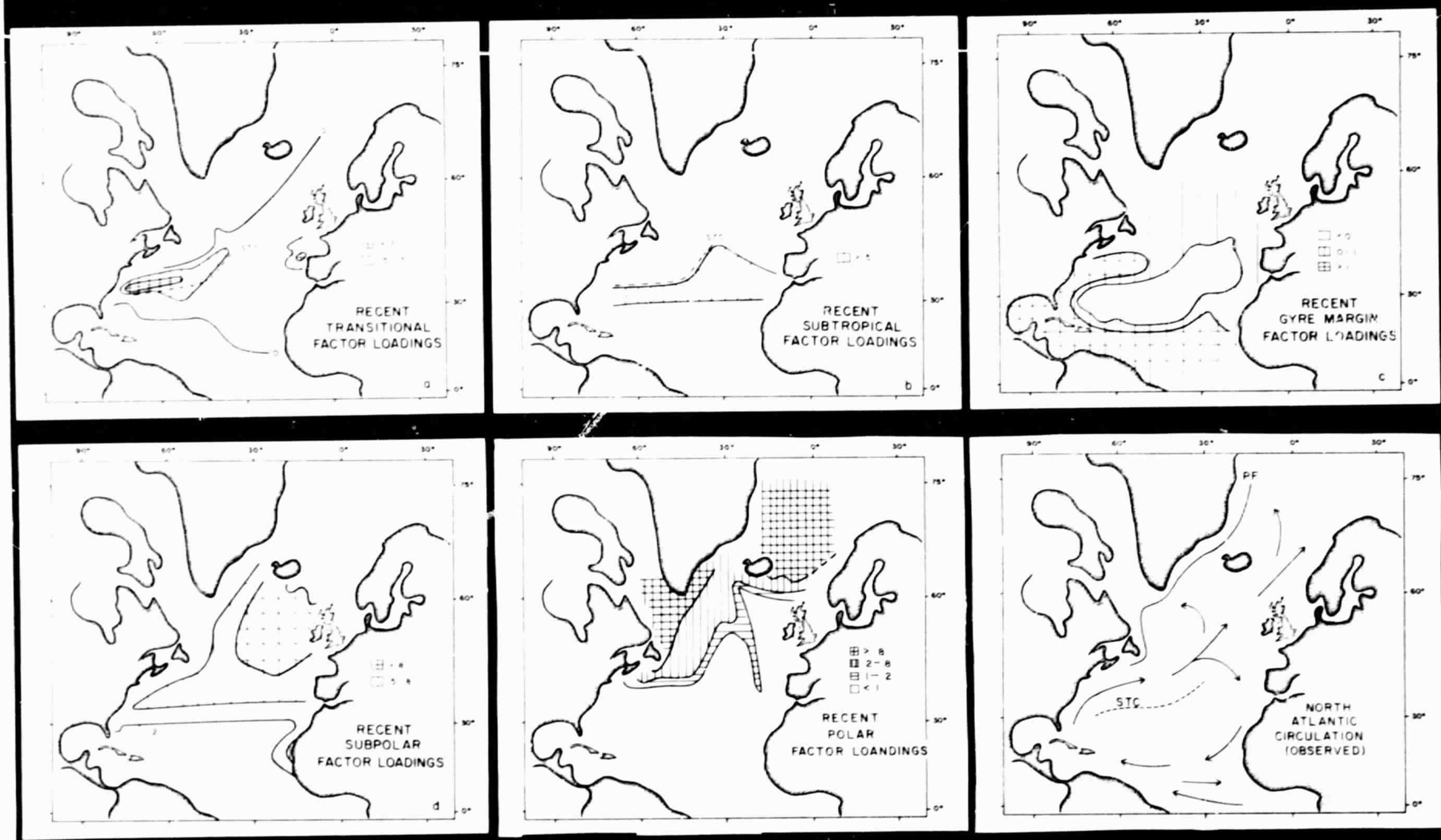

Fig. 29 Comparison of five North Atlantic ecological assemblages of foraminifera with the observed North Atlantic circulation. Each assemblage (eigenvector) has been calculated from a factor analysis of surface sediment samples. Position of the oceanic polar front (PF) and subtropical convergence (STC) from Defant (1961). (After Crowley, 1981; faunal data from Kipp, 1976; oceanographic data from Defant, 1961. Reproduced with the permission of the authors, Elsevier Scientific Publishing Company, and the Geological Society of America).

with F = an eigenvector matrix premultiplied by a matrix of weights (A), and E = the error matrix. The matrix of weights determines the proportional contribution of each of the eigenvectors to a particular sample. The term "factor loading" is the expression that describes the amount of this relative contribution.

Although the first eigenvector is constrained to explain the maximum amount of variance in a system, the physical significance of the eigenvectors may in some cases be obscure (for an illuminating example, see Walsh and Richman, 1981). To facilitate interpretation, it is sometimes useful to rotate the eigenvectors with the orthogonal varimax rotation (Kaiser, 1958),

$$B = AT \tag{4}$$

with B the matrix of rotated factor loadings, and T a transformation matrix, which rigidly rotates A through angles determined by the varimax criterion (for additional details, see Harman, 1976; Jöreskog et al, 1976).

Analysis of 191 North Atlantic surface samples indicates that six eigenvectors account for 92% of the variance (Kipp, 1976). The end members are also sensitive indicators of climatic conditions. There is a remarkable correspondence between the geographic distribution of these factors and the major features of the North Atlantic circulation (Fig. 29): the North Equatorial Current, western and eastern boundary currents, the North Atlantic Current, Sargasso Sea, polar front, and subtropical convergence can easily be distinguished by gradients in the factor loadings. Similar results have been obtained from Pacific and Indian Ocean sediments (Bé and Hutson, 1977; Moore, 1978).

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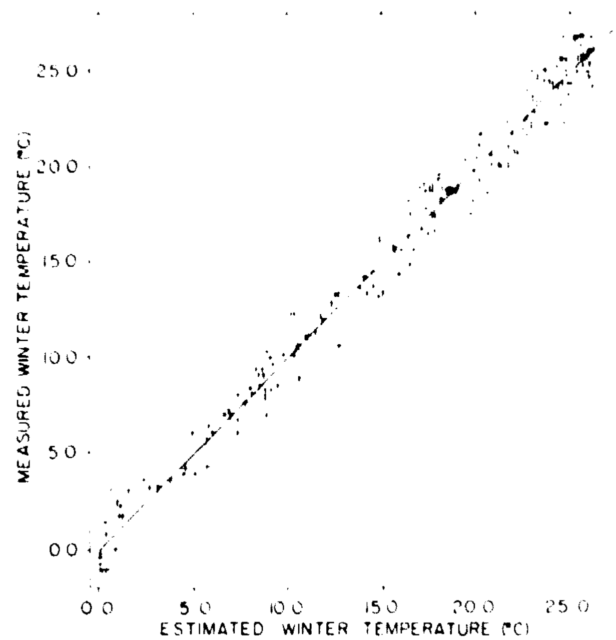


Fig. 30 Observed values of winter temperature vs. estimates calculated by a transfer function on 191 North Atlantic core top samples. Dashed lines indicate 80% confidence interval. (From Kipp, 1976; reproduced with the permission of the author and of the Geological Society of America).

2. Calibration of samples against environmental variables. Once end members have been isolated, they can be used to develop a transfer function. This technique interprets faunal abundances in terms of environmental variables. For example, a sample containing 80% tropical fauna and 20% subtropical fauna would tend to correlate with relatively high SST. Additionally, in a sample containing four different end members, some factors may correlate better with temperature than others (the other factors may be more influenced by salinity, nutrient levels in the ocean, etc.). An objective method of evaluating the relative importance of each factor is by means of a multiple regression, which compares the factors abundances with SST (and other variables), defines the relative importance of the different factors, and computes their overall correlation.

In order to derive paleoenvironmental estimates (temperature, salinity, etc.) from paleoecological data, the matrix of weights (B) for the core tops is first correlated with average observed values in the overlying water column by means of a nonlinear regression (see Imbrie and Kipp, 1971, pp. 86 ff.).

$$\hat{Y}_{ct} = B^2{}_{ct}\ K + k_o \tag{5}$$

with $\hat{Y}_{ct}$ being the matrix of environmental estimates, K the coefficients of the regression, and k_o the intercept. A scattergram of results (Fig. 30) for winter temperature attests to the power of the method: the multiple correlation coefficient is 0.991, with a standard error of estimate of 1.2°C, and an 80% confidence interval of 1.5°C (Kipp, 1976). The relatively low confidence interval seems to reflect a natural noise level in paleoceanographic data. The problem can be partially circumvented by multiple sample analyses, because the confidence interval shrinks by a factor of $N^{1/2}$ (N = number of samples).

3. Estimates of paleoenvironmental values from paleoecological distributions of flora and fauna. The principal behind paleoenvironmental estimates is simple: if a sample from the past has the same faunal composition as a sample from the present (with known correlations with temperature, salinity, etc.), then it is assumed that the past samples represent the same considiton. The manner in which the estimates are obtained involves recognition that the matrix of weights (B) is the variable required to produce the temperature estimates (cf. equation 5). In the core top equation, the B matrix is coupled with the eigenvector matrix in the form

$$U_{ct} = B_{ct}F \tag{6}$$

However, B can be easily isolated. Because F is a row-wise orthonormal matrix, multiplication of both sides of (6) by F' (= F transpose) yields

$$U_{ct}F' = B_{ct} \tag{7}$$

(note that FF' = I, the identity matrix). It is the F' matrix that is used to isolate B in independent data sets, such as a single core (each row in a core is a different core depth):

$$U_cF' = B_c \tag{8}$$

The new matrix of weights for the core is then combined with the constant and regression coefficients from (6) in order to calculate paleoenvironmental estimates for the core

$$\hat{Y}_c = B^2_c\, K + k_o \tag{9}$$

It should be noted that operation of the transposed eigenvector matrix on an independent data set assumes that ecological relations of species in the new data set are the same as in the original. One check on this assumption is to calculate the communality of the samples (equal to the sum of squares of elements in the i'th row of B). Communalities approaching 1.0 indicate that most of the original information in the data matrix has been retained. Another check on the above assumption is to note that the turnover rate of individual species of planktonic foraminifera is about 2-4 m.y. (Berggren, 1969). Therefore, relatively little evolution has probably occurred in Late Quaternary samples. However, extinctions or the appearance of new species will by necessity alter the composition of eigenvectors in unknown ways. Since one definition of the Pliocene/Pleistocene boundary at 1.7 M.y. is appearance of a new species of planktonic foraminifera (e.g., Berggren et al 1980), paleotemperature estimates with the Imbrie-Kipp method have been restricted to the last 1.7 million years.

Problems of interpreting paleoclimatic results Every worker in paleoclimatology is well aware of the difficulties involved in extracting meaningful climatic information from geological data. Marine records have often been eroded, dissolved, or mixed (either by burrowing organisms in the sediment or in the mechanical process of coring), and each of tne effects must be isolated and eliminated before a useful climatic record can be constructed. However, even at this point, serious interpretive errors can be made. In this section, I have included some examples of problems involved in interpreting completed paleoclimatic results.

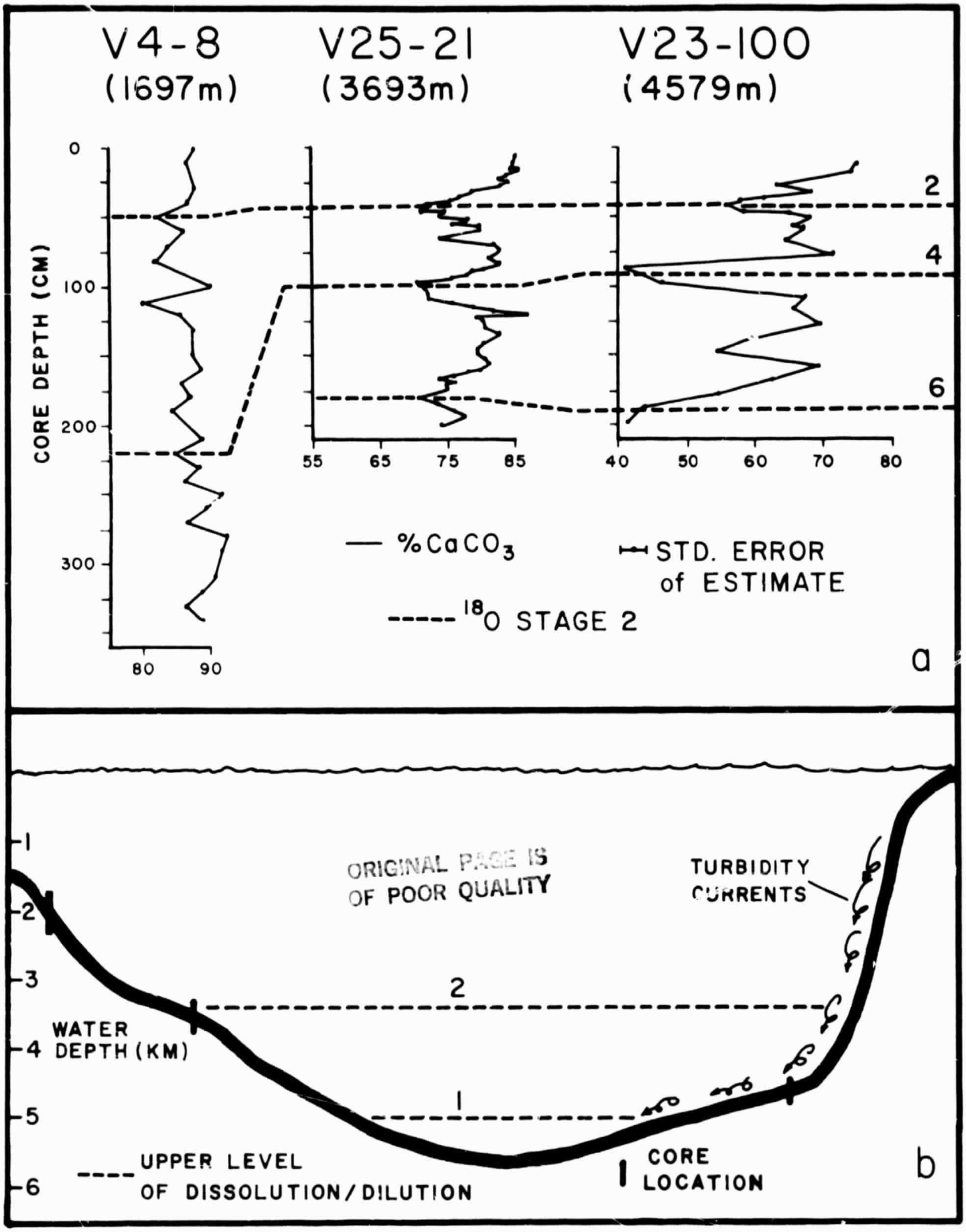
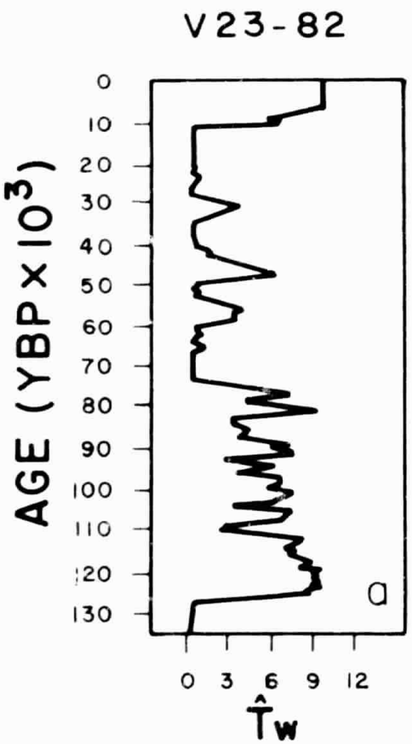


Fig. 31 Interpreting geochemical variations in terms of local dynamic processes: factors responsible for temporal changes in carbonate content in the eastern tropical Atlantic. (a) Carbonate curves from three cores, plotted vs. core depth (data from Parkin and Shackleton, 1973; Crowley, 1976; Crowley and Matthews, 1982). (b) Cross-section of the Canary Basin, with core sites from (a) plotted vs. seafloor depth. Horizontal lines indicate successive stages of dissolution/dilution events, which modify the records in (a).

The first example (Fig. 31) involves interpretation of geochemical signals. The signals are variations of calcium carbonate in sediment cores from the central North Atlantic. Calcium carbonate is a major constituent of deep-sea sediments, and percent fluctuations have been used to interpret such varied phenomena as oceanic productivity, deep-water production rates, and intensity of trand trade winds (Broecker, 1971; Hays and Peruzza, 1972; McIntyre et al, 1972). An inspection of the three records in Fig. 31-a illustrates a different pattern for each core, from relatively constant percentages in the shallow core to pronounced fluctuations of differing intensity in deeper cores. In order to understand why closely-spaced samples can record such different patterns, it is necessary to place the patterns within a dynamic framework.

The carbonate decreases in the cores can be attributed to factors such as increased dissolution of seafloor calcium carbonate by bottom waters (Gardner, 1975) and increased input of non-carbonate material from the continents (e.g., Broecker et al, 1958). In general, the noncarbonate material is injected into the basins by turbidity currents moving down the continental slope (Ewing and Connary, 1970); in this case, the continental slope of Africa (Fig. 31-b). Glacial carbonate values therefore decreased under increased bottom water corrosion and dilution, with the effect penetrating into shallower levels as the intensity of the events increased (the deepest core would still record the most severe effects). Very shallow cores such as V4-8 (Fig. 31-a) would therefore be insolated from the modifying effects of the deep events.



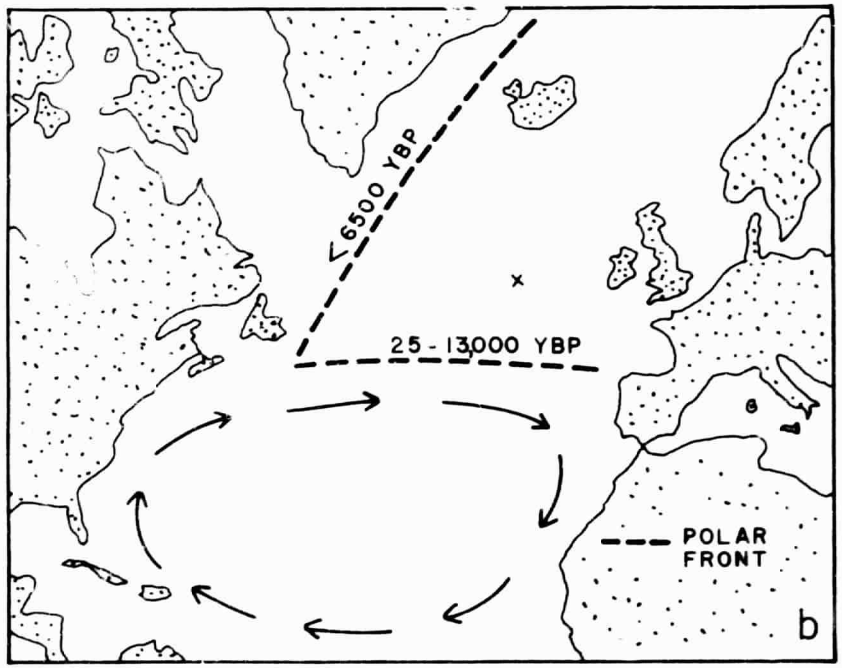


Fig. 32 Interpreting paleoenvironmental estimates in terms of local dynamic processes. (a) Winter paleotemperature record (°C) of a core from the northeast Atlantic ("X" in 32-b), plotted vs. time (from Sancetta et al, 1973). (b) Position of the oceanic polar front during the last 25,000 years (from Ruddiman and McIntyre, 1973). During the last glacial maximum, the Gulf Stream/North Atlantic Current flowed directly eastward across the Atlantic. Sharp temperature changes in V23-82 record the passage of the polar front across the core site.

Another record raises two other problems involved in interpreting climatic records (Fig. 32). A SST-record from west of Ireland records very sharp temperature changes during the last 130,000 years (Sancetta et al, 1973). However, the sharp changes are not due to an instantaneous global change, but rather local passage of the North Atlantic polar front across the core site (McIntyre et al, 1972). Although the local changes of SST were virtually instantaneous, several thousand years were required (Fig. 32-b) for the migration of the polar front from its glacial-maximum position at 42N to its present position in the northwest Atlantic (Ruddiman and McIntyre, 1973). In addition, a comparison of the phase relationships between SST and ice volume would yield different phasing, depending on whether the records were located off the coasts of Spain or Greenland (cf. Hays et al, 1976a).

The above discussion is intended to introduce a note of caution toward the interpretation of paleoclimatic patterns. The primary conclusion from the analysis is that geological records must be interpreted within a local framework of dynamic processes before it becomes justifiable to extract any trend of global significance from the results. Failure to do so may lead to erroneous conclusions. For example, a rapid shift of ^{18}O in the Greenland ice core has sometimes been cited as evidence for a very rapid climatic change (Flohn, 1979; Dansgaard, 1981). But is this conclusion justifiable if local changes in atmospheric circulation patterns are taken into consideration? The obvious counter to local impediments is the establishment of a broader geographic array of detailed records — a goal that the SPECMAP Project (Mapping Variations in Ocean Spectra, 10^{-5} to 10^{-1} cycle/yr.) is presently working toward (e.g., Imbrie, 1982).

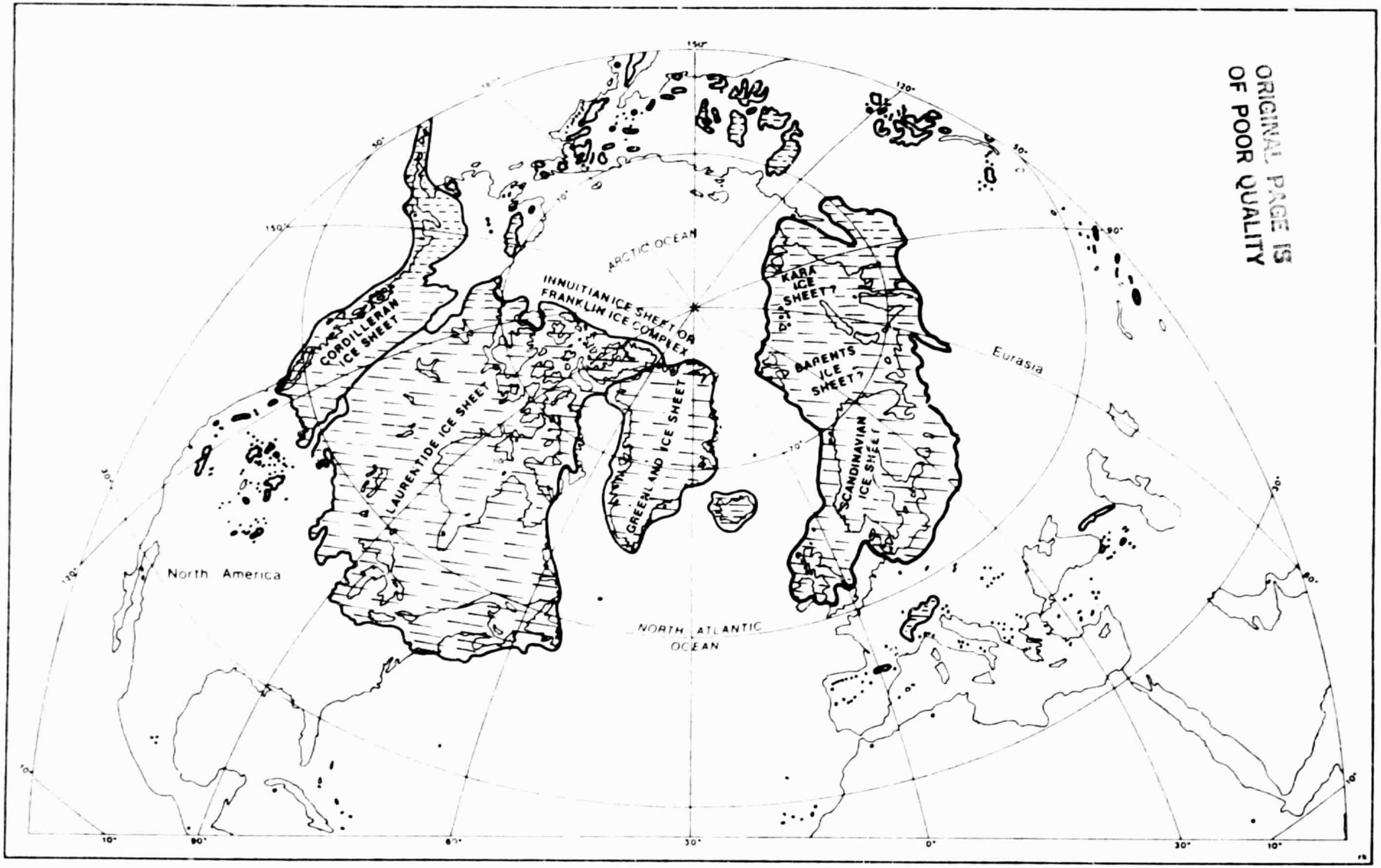


Fig. 33 Northern Hemisphere ice distribution at 18,000 YBP. (From Denton and Hughes, 1981; reproduced with the permission of John Wiley & Sons, Inc.).

C. Results of the 18,000 YBP CLIMAP Project

The last glacial maximum prevailed from about 22-14,000 YBP (e.g., Dreimanis and Goldthwait, 1973). The glaciation has been variously called the Wisconsin, Weichselian, or Würm, depending on whether eastern North American, northern Europe, or the Alps are referred to. The features of the Earth's surface have been mapped in considerable detail by the CLIMAP (Climate/Long-Range Investigation Mapping and Prediction) Project (1976). A vast ice sheet covered substantial parts of the northern hemisphere (Fig. 33), and sea ice spread equatorward in both hemispheres (Fig. 34). Approximately 18% of the earth's surface was ice-covered, with the global average surface albedo (summer) increasing from its present value of 0.14 to 0.22 (Gates, 1976). Construction of the ice sheets required evaporation of about $50\text{-}60 \times 10^6$ km^3 of water from the oceans, with world sea level dropping about 120 m (Hughes et al, 1981). Most of the ice accumulated in eastern North America and northwestern Europe, with another smaller center in Russia. There are some indications that the ice sheets may have been linked via a thicker Arctic Ocean Ice Shelf (Broecker, 1975b; Hughes et al, 1977). In eastern North America, the Laurentide Ice Sheet reached thicknesses of 2500 m, and extended from the Rocky Mountains to the Atlantic shore, and from the Arctic Ocean southward to about the present positions of the Missouri and Ohio Rivers. In Europe, the Fennoscandian Ice Sheet reached northern Germany and the Netherlands. The weight of the massive ice sheets depressed the crust by as much as 700-800 m (Flint, 1971).

In ice-free areas, large lakes developed in the Great Basin of western North America (Peterson et al, 1979) and in European Russia (Grosswald,

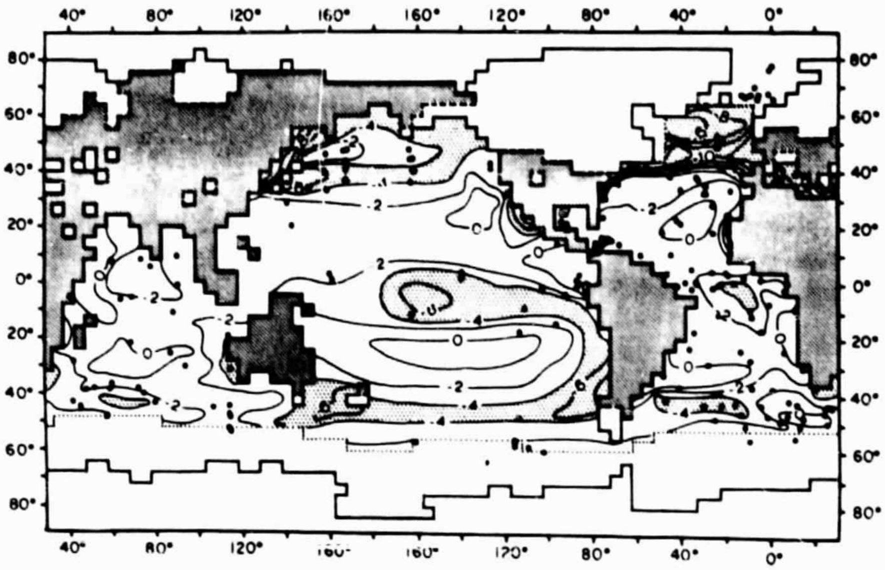


Fig. 34 Difference between August sea-surface temperatures 18,000 years ago and modern values. Contour interval is 2°C. Areas where the temperature change was greater than 4°C are shown in light stippling. Ice-free land areas are shown in darker stippling. Continental ice and outlines conform to a grid spacing of 4° latitude by 5° longitude. Heavy solid lines indicate continental outlines; dashed lines are ice margins on land; dotted lines indicate sea-ice margins. Large dots mark the locations of cores used in reconstructing sea-surface temperatures 18,000 years ago. (From CLIMAP project members, SCIENCE, 191:1131-1137, 18 March, 1976; copyright 1976 by the American Association for the Advancement of Science).

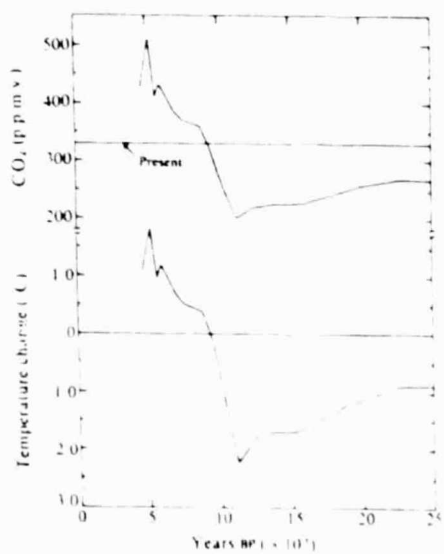


Fig. 35 Graph showing the relationship between the atmospheric CO_2 level reconstructed from Camp Century, Greenland, core measurements (upper curve) and the corresponding global change in surface air temperature estimated from current climatic models (lower curve). Recent measurements (Neftel et al, 1982) indicate that the CO_2 peak of the last 10,000 years is spurious; i.e., CO_2 values rose to about 300 ppm and have remained at about that level through most of the last 10,000 years. (Data by Berner et al, 1980; figure from Thompson and Schneider, 1981; reproduced with the permission of the authors and NATURE).

1980). Factors possibly responsible for these patterns include increased cloudiness (Benson, 1981) and/or lower air temperatures (Brakenridge, 1978) in the former, and meltwater runoff/physical blockage of drainage systems in the latter (Grosswald, 1980). Other than the above-mentioned lakes, much of the remaining surface of the planet dried out considerably. Extensive sand dunes occurred in Central America (Ward, 1973) and the sub-Sahara (Sarntheim, 1978). Lake levels in Africa were very low (Street and Grove, 1979), and the Amazon Rainforest may have been reduced to a few refugia (Haffer, 1969).

Over the oceans, low SST accompanied the equatorward migrations of the polar fronts (CLIMAP, 1976). For unknown reasons, Southern Ocean sea ice did not melt back during the austral summer (Hays et al, 1976b). In the northeast Atlantic, SST were as much as 10°C lower than today (McIntyre et al, 1976). The Gulf Stream/North Atlantic Current, which presently flows into the northeast Atlantic (Fig. 30-b, p. 58), flowed directly eastward towards Gibraltar (McIntyre, 1967). Along the eastern boundary currents, present sites of equatorward penetration of cool waters, SST decreased about 6°C. The cool waters extended into the eastern equatorial Atlantic and Pacific (Fig. 34). However, over large parts of the western tropical oceans, and extending sometimes to latitudes of 40°, SST varied by less than 2°C. Globally, SST decreased by 2.3°C.

A final, remarkable feature of the ice-age earth involves the apparent concentration of CO_2 in the atmosphere (Fig. 33). Studies of gas inclusions in the Greenland and Antarctic Ice Sheet cores (Berner et al, 1980; Delmas et al, 1980) indicate glacial concentrations of about 200 ppm, almost 100 ppm less than the estimated pre-industrial value of about 290 ppm (Broecker, 1975a). The fluctuations in atmospheric CO_2 values may be due to variations in marine biological productivity (Berger et al, 1981; Broecker, 1981). CO_2 is incorporated into the calcium carbonate shells of the organisms. Factors that control shell preservation are quite complex (e.g. Broecker, 1971), and influence the oceanic CO_2 content. Since oceanic CO_2 concentrations exceed atmospheric values by 60X (Garrels and MacKenzie, 1972), the atmospheric values are therefore susceptible to strong oceanic influence.

CLIMAP boundary conditions of SST, ice extent and elevations, and land albedo were inputs into two general circulation model (GCM) reconstructions of the ice-age atmospheric circulation (Gates, 1976, 1977, 1981; Manabe and Hahn, 1977). As model results are dependent on such factors as parameterizations, grid resolution, etc., the following results should be accepted with some reservation. Nevertheless, the calculations do provide some very suggestive trends.

The atmospheric reconstruction indicates surface high pressure systems over the Laurentide and Fennoscandian Ice Sheets, with resultant easterly winds at the southern edge of the ice margins. Westerly wind velocities increased somewhat and shifted southward. The Hadley cell was noticeably weakened, probably because lower SST in equatorial regions decreased overlying air temperatures. The lower air temperatures (with lower absolute humidity) provided significantly less kinetic energy for the mean meridional motions (Kraus, 1973). The decreased uplift, coupled with reduced land-sea differences, would also account for decreased precipitation and enhanced tropical aridity (Manabe and Hahn, 1977). The average planetary air temperature (Gates, 1981) decreased by 3.5°C (August) and 3.3°C (February).

An independent CLIMAP data set of about 20 low-and mid-latitude pollen and periglacial temperature estimates provided "observational verification" of selected air temperature estimates: comparison of the different approaches generally produced estimates similar to within 1-2°C (Gates, 1976).

A particularly interesting result from the global two-level reconstruction (Gates, 1977) indicates that there was a downward heat flux of 100-200 W/m^2 along the southern margin of the Laurentide, Fennoscandian, and Siberian Ice Sheets, but not the Greenland Ice Sheet. This flux is comparable to that occurring today in the Arctic summer, when ice melts back at a rate of about 5 cm/day. The coincidence of simulated melting with the "unstable" subset of Pleistocene ice sheets may prove to be an important element of future integrated ice age models.

Since the boundary conditions of both SST and albedo were different during the ice age, it is interesting to inquire which factor may have been more responsible for the increased tropical aridity. When present SST and glacial albedo boundary conditions were stipulated for the 12-level Geophysics Fluid Dynamics Laboratory model (Manabe and Hahn, 1977), continental runoff decreases were intermediate for South America and equatorial Africa, but substantially reduced over the region of the South Asian monsoon. This result is consistent with suggestions (Flohn, 1968) and calculations (Hahn and Manabe, 1975) that a principal factor affecting the strength of the summer southwest monsoon is the extent of the snow cover over the Tibetan Plateau.

D. Temporal Trends in Pleistocene Records

Climatic records from time intervals other than 18,000 YBP provide additional valuable information about the nature of glacial-scale climatic fluctuations. Although the last glacial maximum certainly had some impressive statistics associated with it, continental evidence indicates that some earlier ice sheets may have been about 10% larger (Flint, 1971; Frakes, 1979). Analysis of an ^{18}O record for the entire Pleistocene indicates some distinctive secular trends to ice volume patterns. The last 900,000 years has been distinguished by relatively large ice sheets, fluctuating with an approximately 100,000 year period. Earlier intervals indicate different dominant periodicities and less variability (Shackleton and Opdyke, 1976; Pisias and Moore, 1981).

Analysis of microfossil records provides evidence for some divergent local surface water trends. For example, the Arctic Ocean has apparently remained ice-covered for at least the last five million years (Clark et al 1980), while the Norwegian Sea has remained ice-free only during peak interglacial events (Kellogg, 1976). The latter region is therefore one of extreme climatic sensitivity, a result consistent with observations of air temperature trends over the northern hemisphere during the 20'th century (e.g., Mitchell, 1961). Farther south, a record from the central waters of the North Atlantic gyre (Crowley, 1976, 1981) indicates essentially constant SST during the last 150,000 years in the Sargasso Sea (Fig. 36). The pattern is somewhat similar to a SST-record for the entire Pleistocene in the

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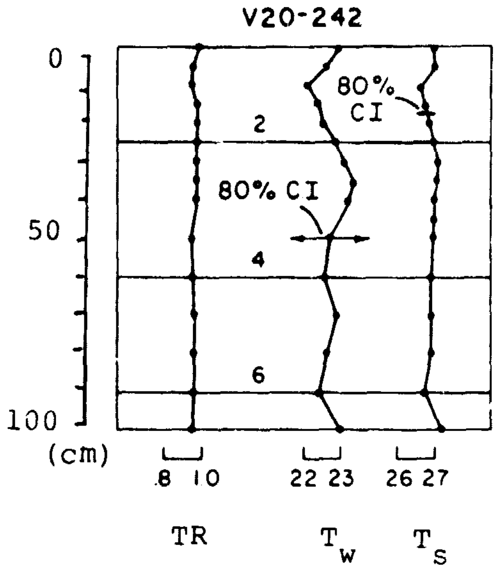

Fig. 36 A stable SST record in a core extending through the last glacial cycle (record length is about 160,000 years). The core is from the central waters of the North Atlantic gyre; tropical factor loadings (TR) and winter and summer temperature estimates (T_W, T_S) are plotted vs. core depth (cm). 80% confidence interval for estimates from Kipp (1976). Horizontal lines are 0-18 Stages 2, 4, & 6 (cf. Fig. 24, p. 46). SST varied little because the tropical factor loadings, which dominate the sample, varied little. (The squared sum of the factor loadings equals the sample variance; tropical factor loadings averaging 0.95 therefore indicate that about 90% of the sample variance is explained by this one factor; from Crowley, 1976).

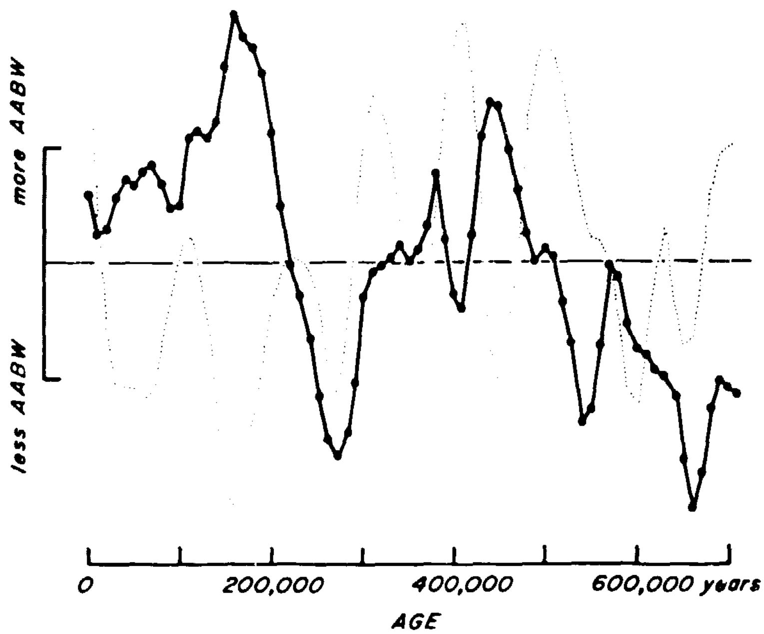

Fig. 37 Evidence for a nonlinear relationship between ice volume and bottom water production rates. Abundance of Antarctic Bottom Water indicators (solid line) and calcium carbonate, plotted vs. time. In this core, the calcium carbonate record correlates with the 0-18 ice volume record (interglacials are at the top of the figure). (From Lohmann, 1978, OCEANUS, v. 21, No. 4, p. 64).

equatorial Atlantic (Briskin and Berggren, 1975). Results from the CLIMAP 18,000 YBP study indicates that stable SST in the interiors of the subtropical gyres seems to be a general phenomenon for the world ocean (CLIMAP, 1976). In some regions, SST patterns led (Hays et al, 1976a) or lagged (Ruddiman and McIntyre, 1981c) ice volume increases, and in the central North Atlantic SST even increased during a phase of ice growth (Crowley, 1981).

Intervals as warm as the recent have occurred for only about 10% of the time during the Late Quaternary (Emiliani, 1972). During the last interglacial maximum (about 125,000 YBP; see Fig. 24), initial reconstructions indicate a SST-pattern comparable to the present (Ruddiman and CLIMAP project members, 1979). At that time, sea level is thought to have been 3-6 m higher (Mesolella et al, 1969), a figure attributed to a possible melting of the West Antarctic Ice Sheet (Mercer, 1968). It has been suggested that one consequence of a future CO_2-induced warming would be a re-melting of the West Antarctic ice sheet (e.g., Mercer, 1978).

Because of its long response time (10^3 yrs.), the deep waters of the ocean have sometimes been described as a flywheel, whose inertia dampens high-frequency fluctuations, and whose fluctuations may contribute significantly to the lower-frequency portions of the variance spectrum. Deep/bottom water production rates have also varied during the Pleistocene. North Atlantic Deep Water (NADW) production apparently decreased during part of the last glacial cycle (e.g., Streeter and Shackleton, 1979; Schnitker, 1980). Of even greater interest are results from the Vema Channel in the western South Atlantic (Lohmann, 1978), where there is an indication of an unexpectedly complex coupling of deep and surface water fluctuations (Fig. 37). Antarctic Bottom Water (AABW) production was apparently greater than today during some glacials, and less than today during other glacials. Similar conclusions have been obtained from the Southern Indian Ocean (Corliss, 1979). This type of result raises doubts about certain classes of climate models that require some type of 1:1 coupling between surface and bottom waters (e.g., Newell, 1974; Weyl, 1968).

E. Correlations With Orbital Parameters

A fundamental problem in climatology concerns the origin of the Pleistocene ice sheets. Substantial progress has recently been achieved in delineating the outlines of the problem. In a pivotal paper, Hays et al (1976a) provided strong empirical support for an orbital perturbation influence on glacial activity -- i.e., the Milankovitch hypothesis (1941). The perturbations are due to the gravitational effects of the different planets on three features of the Earth's orbit: 1) the tilt of the axis (the obliquity - period of 41,000 years); 2) the longitude of perihelion (the precession effect: periods of 19,000 and 23,000 years; 3) the eccentricity of the orbit (periods of about 100,000 and 400,000 years). The first two effects result only in a geographic, seasonal redistribution of solar insolation at the top of the atmosphere. Eccentricity variations produce a maximum change of 0.2% in total insolation received, which in a simple climate model produces only about a 0.4°C change in the average surface temperature of the planet (North et al, 1981). However, the eccentricity variations can strongly modulate the amplitude of the precession effect, with insolation anomalies during high eccentricity 3-4 times larger than anomalies during low eccentricity. At 60N, the difference translates into a summer seasonal difference of about 40 W/m^2 (Schneider and Thompson, 1979), which is a variation of about 8% around the mean value for that latitude (Imbrie and Imbrie, 1980). Values at the summer solstice exceed 13% of the mean (D. Short, personal communication, 1981). (It has long been thought that the summer insolation values at this latitude might be the critical index for

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Fig. 38-a

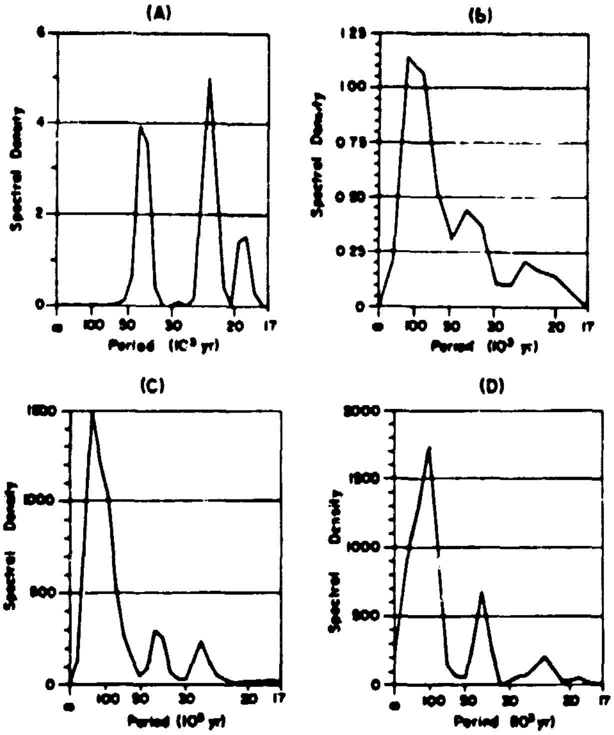


Fig. 38-b

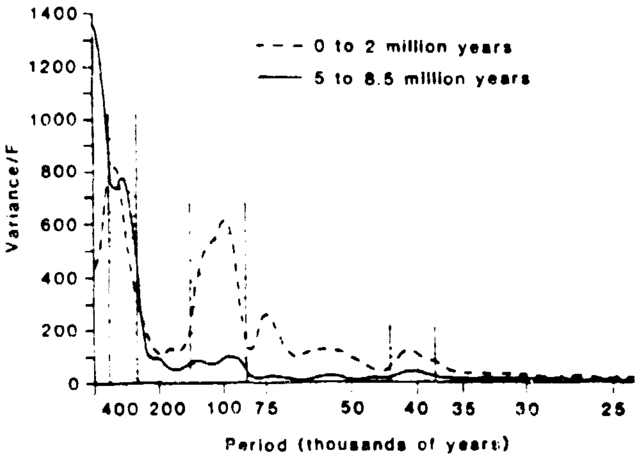


Fig. 38 Late Cenozoic spectral records. (a) Orbital input and O-18 ice volume output from three different Late Quaternary cores, each showing the dominance of the 100K peaks. Orbital data from Berger (1978). Output data from Hays et al (1976a), and Shackleton and Opdyke (1973, 1976). Record lengths are 270K (B), 700K (C), and 600K (D). (From Birchfield et al, 1981; reproduced with the permission of QUATERNARY RESEARCH). (b) Variance spectrum of carbonate content in pre-Pleistocene records. This figure illustrates the time-dependency of the 100K peak, and the relative stability of the 400K peak. (From Moore et al, 1982; reproduced with the permission of the authors and Elsevier Scientific Publishing Company).

determining glacial inception. Much of that latitudinal strip is above sea level, and cool summers would allow enhanced preservation of winter snows, triggering an ice-albedo feedback loop - from Köppen and Wegener, 1924.)

In a study of time series from the Southern Indian Ocean (e.g., Fig. 38-a), Hays et al (1976a) demonstrated: 1) that spectral peaks in the ice volume record occurred at values of 23, 41, and 105K (K = thousands of years); 2) that the 105K contribution to the total variance far exceeds that expected from a simple linear relationship between insolation and ice volume; 3) that there is a fairly coherent phase relationship between insolation, SST, and ice volume: each preceeded the next by 2-4K. Briskin and Harrell (1980) have added evidence for a 400K peak in a SST record spanning the entire Pleistocene. A. Berger (1976) has noted that the most important term in the series expansion for eccentricity occurs at this period.

Since the establishment of an orbital connection, recent studies have centered on two areas on research: coupling insolation changes with atmosphere-ice models, and explanations for the dominance of the 100K peaks. For example, Suarez and Held (1976, 1979) incorporated orbital parameters into a seasonal energy balance model, keyed to ice-albedo feedback, and produced an output qualitatively similar to the geologic record of the last 150,000 years. Pollard (1978) added Weertmann's (1976) ice-sheet model to an energy balance model and also generated some encouraging results. However, Pollard noted that the dominant 100K cycle is poorly simulated by all of these models.

The non-linearity of the 100K peak has prompted a series of explanations, none of which have unequivocal pre-eminence at the present time. In a general sense, the models revolve around the non-linearities associated with the mechanics of ice-sheet growth and decay, for there is good theoretical and

Fig. 39 Evidence for oceanic feedback mechanisms during glacial inception and disintegration. (a) Summer SST record from the central North Atlantic -"X" in 37-c. Temperature minima are primarily due to meltwater outflow from the Laurentide Ice Sheet. (b) Variance spectrum of SST record. (c) Oceanic coupling patterns during a glacial cycle. Ice-rafted debris data from Ruddiman (1977). (From Ruddiman and McIntyre 1981c, SCIENCE, 212:617-627, 8 May 1981; copyright 1981 by the American Association for the Advancement of Science).

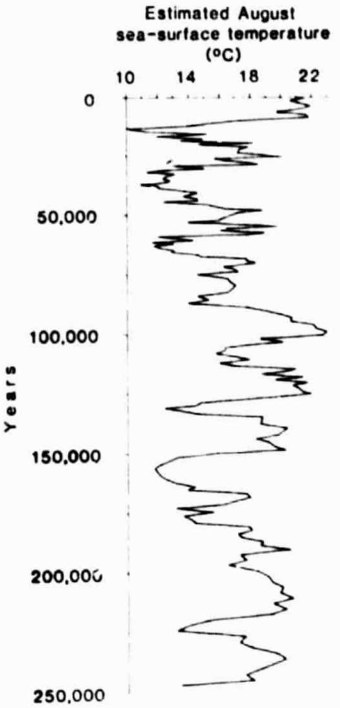


Fig. 39-a

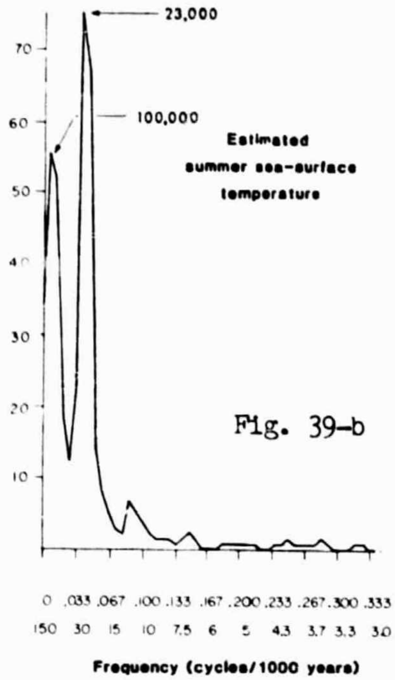
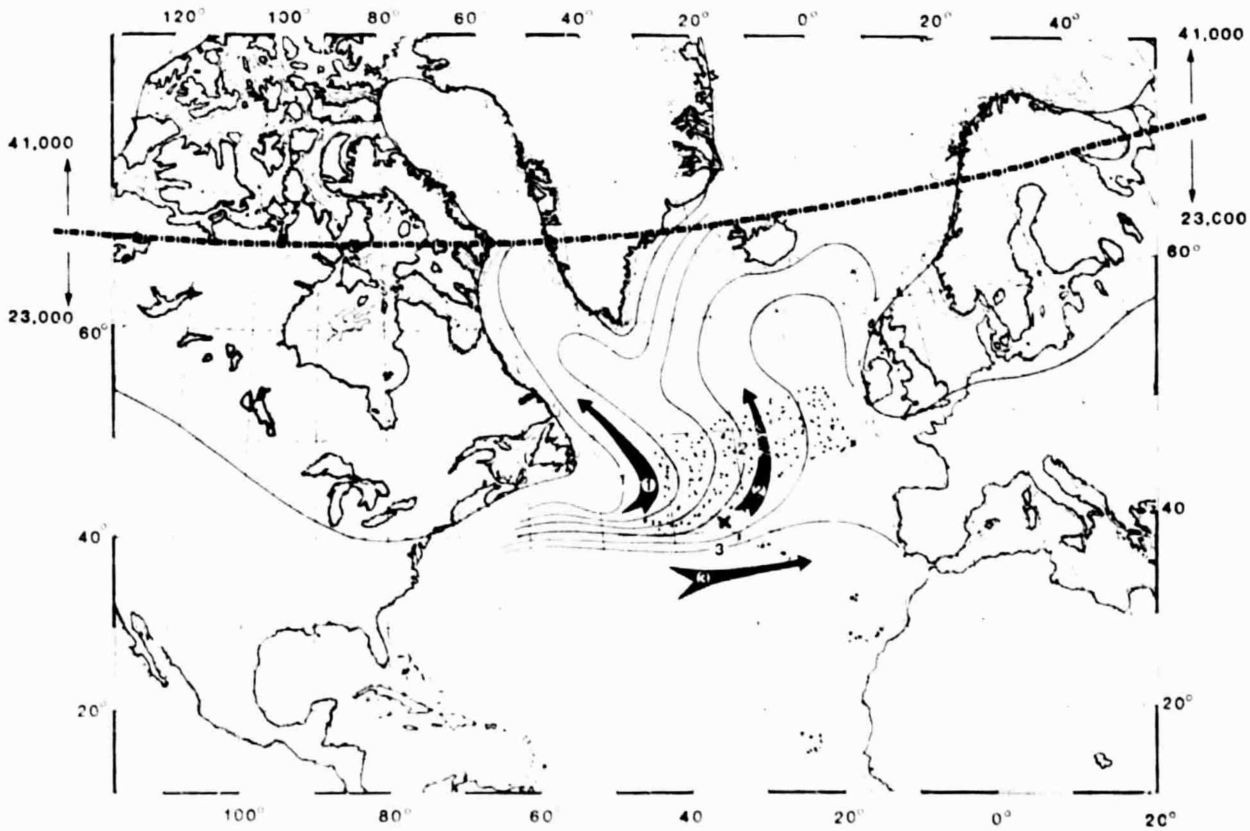


Fig. 39-b

Fig. 39-c



Schematic representation of summer iceberg and meltwater plumes formed at different times during 23,000-year cycles of fluctuating outflow from the Laurentide Ice Sheet. Latitude 65°N separates the area dominated by the 41,000-year (tilt) frequency of summer insolation change from the region dominated by 23,000-year (precession) frequency. Stipple marks outline the region of maximum deposition of ice-rafted debris, with iceberg-meltwater plumes drawn in this general area. Largest plume extents correspond to maximum iceberg-meltwater influx. Winter sea-ice limits roughly match the meltwater plume limits. Arrows show inferred winter storm tracks and moisture transport along winter sea-ice limit during melt-product outflow at minimal (arrow 1), moderate (arrow 2), and maximum (arrow 3) levels.

empirical support for the asymmetric nature of ice growth (e.g., Weertmann, 1964; Broecker and van Donk, 1970). Imbrie and Imbrie (1980) have estimated that the time constants for growth and decay differ by a factor of about 4.

Specific explanations for the large 100K variance generally fall into two classes. Each class incorporates orbital input and terrestrial nonlinearities, but the explanations differ in emphasis. In the first case (e.g., Ruddiman and McIntyre, 1981c), phase relationships between precession and obliquity still drive the system. The mechanism is as follows. The precession effect is strongest in mid-latitudes; 23K oscillations modulate the margin of the main, primarily mid-latitude, Laurentide Ice Sheet at this frequency. The resultant outflow of Laurentide meltwater into the central North Atlantic at 23K intervals bears evidence to the effectiveness of this mechanism (Fig. 39-a, b). When conjunctions between precession and obliquity occur at glacial terminations, the melting effect penetrates into high latitudes, driving the system toward full interglacial state. Feedback loops involve: 1) a large outflow of glacial meltwater into the northwest Atlantic (see Ruddiman et al, 1980); the cold water steers winter storms away from the ice sheets, further depleting the mass balance of the glaciers (Fig. 39-c); 2) catastrophic calving of glaciers along the edge of the rising sea (Hughes et al, 1977). Thus, the coupling of insolation maxima to oceanic feedback loops produces a large step-function response at deglaciations; in effect, the 100K peaks in the spectral record are produced by nonlinearities that occur for an interval of 10,000 years, once every 100,000 years.

The second class of models explaining the 100K peak addresses the inherently non-linear interactions involving ice-albedo feedback, plastic ice sheets, elastic lithosphere, viscous mantle, etc. (e.g., Weertmann, 1976; Kallen et al, 1979). Orbital forcing is portrayed as modulating a

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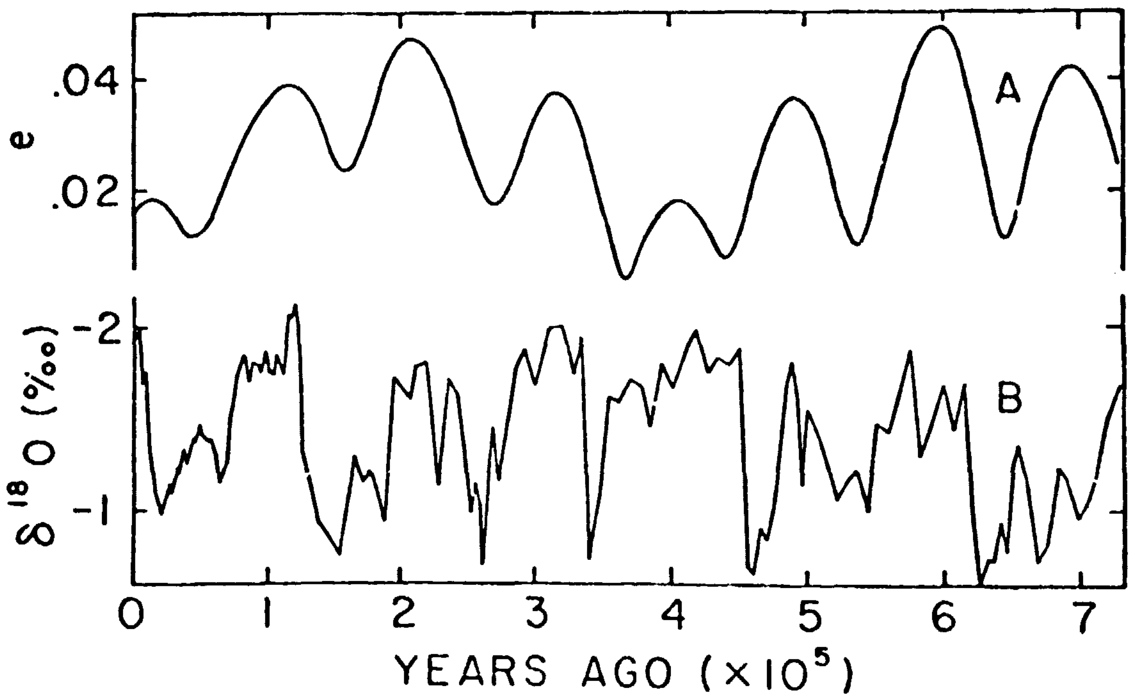

Fig. 40 Eccentricity and global ice volume (0-18) over the past 730,000 years. (a) Variations in orbital eccentricity as calculated by Berger (1978). (b) Oxygen isotope curve for deep-sea core V28-238 (from Shackleton and Opdyke, 1973). Time scale by Kominz et al (1979). This time scale has been tuned to the orbital record, but at the obliquity frequency; thus the figured coherence between eccentricity and ice volume is a significant result. Also, strict linear interpolations between 0-730,000 YBP indicate that the Kominz et al time scale does not vary drastically from linearity (D. Short, personal communication, 1981). (From Imbrie and Imbrie, SCIENCE, 207:943-953, 29 February 1980; copyright 1980 by the American Association for the Advancement of Science).

self-oscillating system. For example, Wigley (1976) noted that transmission of 19K and 23K frequencies through a nonlinear system can produce substantial power (combination tones, or subharmonics) at about 100K. Models that incorporate specific interactions support this concept (Ghil, 1981).

As each of the above models provides a reasonable explanation for climatic fluctuations of the last few hundred thousand years (e.g., Fig. 40), a critical discrimination involves comparisons of predictions against the older part of the climatic record. In fact, there is some indication (Fig. 38-b) that the dominant signals in older records may not be 100K (Shackleton and Opdyke, 1976), in which case the "combination tone" models may not be valid. For example, Pisias and Moore (1981) have analyzed the ^{18}O record for the entire Pleistocene and found that the 100K cycle is only present in records from the last 900K. Between 0.9-1.45 M.y., the records are dominated by cycles between 20-60K. Between 1.45-2.0 M.y., only the 20K cycle is present, and then at a reduced amplitude. Pisias and Moore (1981) suggest that recent modelling efforts (e.g., Imbrie and Imbrie, 1980) may provide an explanation for the above pattern. These efforts indicate that the dominant low-frequency component observed in the late Pleistocene can result from different time constants for the rate of glacial growth and decay. Pisias and Moore (1981) hypothesize that during the early Pleistocene the rate of growth and decay were more similar, and that continental erosion by successive glacial advances lowered the land surface in areas of ice-cap formation to below sea level. When the ice-caps became marine-based, more rapid decay of the ice became possible.

Pre-Pleistocene There is a small but growing number of studies that have also provided information on pre-Pleistocene cyclicity. For example, in a comparison of Pleistocene and Late Miocene (5.8 - 8.5 M.y.) records, Moore et al (1982) have shown that a strong 400K cycle can be detected in carbonate records from the equatorial Pacific (Fig. 38-b). There is also some evidence for periods of about 100K, 40K, and 20K in the Miocene records. However, Moore et al (1982) note that the total variance of Miocene records at 100K periods is substantially less than in Late Pleistocene records.

In a further comparison of the Pleistocene-Miocene records, Moore et al (1982) have touched upon an important point that must be incorporated into models of climatic change - the varying role of eccentricity in climatic fluctuations. For the late Pleistocene, the correlation between eccentricity and ice (^{18}O) seems justifiable (e.g., Fig. 40). However, Kominz and Pisias (1979) have shown that a cross-spectral analysis between ^{18}O and eccentricity for the last 2.0 M.y. shows no significant coherence between the dominant frequencies of eccentricity and the isotope record. This result is in part due to the afore-mentioned lack of any significant 100K cyclicity in early Pleistocene records. However, an analysis by Moore et al (1982) of a Pleistocene carbonate record does yield a significant 100K peak for the entire Pleistocene that is coherent with eccentricity. Moore et al (1982) point out that the explanation for this discrepancy between the carbonate and ^{18}O spectral records may lie in the fact that the different variables are monitoring different parts of the climate system (oceans and ice sheets). Peterson and Lohmann (1982) have recently provided empirical support for this idea. In a study of Antarctic Bottom Water (AABW) fluctuations in the South

Atlantic, they note that the major change in AABW formation during the late Pleistocene at about 700K post-dates by almost 200K the major change in ice volume patterns (i.e., emergence of a dominant 100K peak in the spectral record - Shackleton and Opdyke, 1976; Pisias and Moore, 1981).

In addition to the above-mentioned spectral analyses, preliminary examinations of pre-Miocene climatic records continues to indicate the presence of an oribtal influence. Dean _et al_ (1981) have determined an average cycle of about 40K in carbonate records between 20-37 M.y., and cycles of about 20K in records between about 37-45 M.y. There are also indications for a 20K cycle in lake records of about the same age (Bradley, 1930). Finally, cycles of about 20K duration have been reported from rocks of about 100 M.y. (Fischer, 1980) and 250 M.y. (Anderson, 1982).

From the above discussion, it should be apparent that the spectral signature of climatic fluctuations is far more complex than might have been anticipated. The significant progress of the last few years will be hastened by additional detailed studies based on two lines of approach: 1) detailed spectral analyses of selected Pleistocene records by the SPECMAP Project (Mapping Variations in the Ocean Spectra, 10^{-5} to 10^{-1} cycles/yr.; cf. Imbrie, 1982); 2) analysis of the new hydraulic piston core results (see Section VI), which will provide access to high-quality records extending millions of years beyong the previous limit of about 2 M.y.

Stochastic Effects A further analysis of climatic records indicates that, despite the presence of orbital influence, a large part of the total variance may be of stochastic origin: i.e., low-frequency red-noise variance

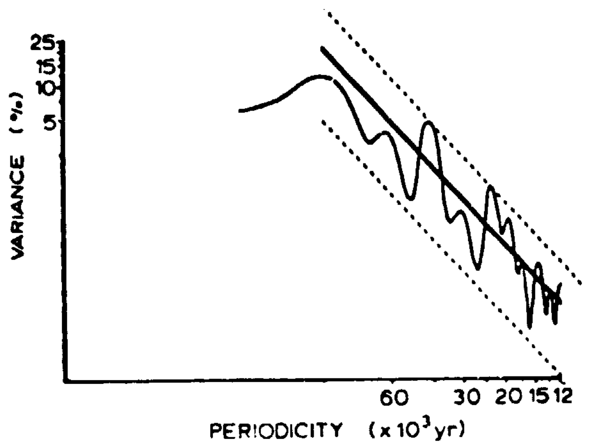

Fig. 41 Log-log plot of the variance spectrum of the 0-18 record in Fig. 40. The solid straight line has a slope of -2, while the dashed lines represent the 80% confidence interval about that line. (From Kominz et al, 1979; reproduced with the permission of the authors and Elsevier Scientific Publishing Co.).

can be produced when white-noise forcing, from a time series with a short-response time, is applied to another system with a relatively long response time (e.g., Hasselmann, 1976). Support for the presence of this process is derived from the analysis of ^{18}O ice-volume records, which indicates that a linear response model relating insolation and ice volume can explain less than 25% of the variance in Pleistocene climatic records (Kominz and Pisias, 1979). More realistic nonlinear models still only account for about 35% of the variance (N. Pisias, in Imbrie and Imbrie, 1980, footnote 45). Furthermore, log-log plots of the variance spectrum (Fig. 41) yield a red-noise spectrum with a slope of -2 between 100K and 12K (the Nyquist frequency). This trend is consistent with that predicted by white-noise forcing from time scales shorter than 12K (Hasselmann, 1976; Pisias _et al_, 1979). For a lucid discussion on the relationship between red-noise variance and climate sensitivity, the reader is referred to Section 8 of North _et al_ (1981).

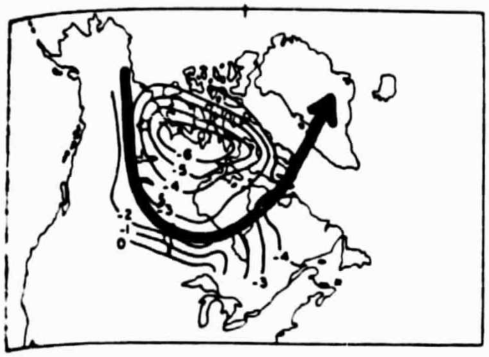

Fig. 42 A possible pattern for the geographic distribution of summer temperature anomalies during glacial inception. The values for temperature are a sum of two results: insolation-induced summer temperature changes during Northern Hemisphere insolation minima (from Shaw and Donn, 1968), and actual observed late summer temperature anomalies for Arctic Canada after unusually cold winters. Heavy arrow indicates conjectured average path of a jet stream perturbed by the surface temperature anomalies. (After Williams, 1978; reproduced with the permission of QUATERNARY RESEARCH).

F. A Model For Glacial Inception

The final feature of Pleistocene glaciation to be discussed in this survey involves the problem of formation and growth of ice sheets. An early theoretical study of glacier mechanics suggested that ice sheets may require as much as 30,000 years to reach full size (Weertmann, 1964). However, an examination of ^{18}O ice volume records has significantly revised this concept. Values subsequent to the last interglacial maximum (^{18}O substage 5-e/5-d boundary at about 120K - see Fig. 24, p. 46) indicate that as much as 50% of maximum ice volume may have been reached within 10,000 years. This figure would correspond to an equivalent ice volume of about 75-m of seawater (Hughes et al, 1981). Another study suggests that a climatic deterioration at about 230K may have reached 80% of maximum ice volume values in 10,000 years (Ruddiman et al, 1979). The above results seem to imply that snow accumulation rates during times of ice growth in regions like northeast Canada must have been much higher than the present (e.g., Andrews and Mahaffey, 1976; Weertmann, 1976).

Some simple conceptual models for enhanced moisture transport are supported by the geological record. For example, calculations of surface-temperature changes due to insolation reductions suggest July temperature changes in northern Canada of about 6°C (Fig. 42). Lower air temperatures should have allowed increased summertime preservation of snow. The expanded snow cover should deepen upper troughs during the winter (Flohn, 1974, Williams, 1975), thereby steering low pressure systems into northeast Canada. Present-day synoptic studies have indicated that the deepened troughs produce a maximum moisture transport into northeastern Canada, with southeast winds blowing in from the Atlantic (Barry, 1966; Rogers and van Loon, 1979). Furthermore, the pattern is associated with warm SST in the northwest Atlantic (Rogers and van Loon, 1979). It is interesting to note that during phases of maximum glacial growth, the geologic record indicates that the northwest Atlantic was ice-free, with SST comparable to the present (Ruddiman and McIntyre, 1979). The thermal contrast of cold land/warm sea (Fig. 39-c) should have further enhanced baroclinicity, thereby providing the hydrous fuel necessary for rapid glacial growth. The model receives additional support from indications of enhanced accumulations during a time of rapid ice growth in nearby Greenland (Andrews et al, 1974). Thus, an interweaving of data and ideas has produced a conceptual model of how the ice sheets may have developed. It now remains to be determined whether the basic elements of the model are supported by more rigorous quantitative investigations.

VIII. HOLOCENE (last 10,000 years)

Features of Deglaciation The Holocene Interglacial Epoch represents the present warm period that has prevailed since the last glacial maximum. The absolute age of the midpoint of the transition from glacial to interglacial has long been stated as 11K (e.g., Broecker et al, 1958). However, there is considerably more structure to the nature of the transition than was previously thought. For example, there is evidence that perhaps half of the volume of continental ice disappeared between 13-16K (Kennett and Shackleton, 1975). Ruddiman and McIntyre (1981a) note that this result has some bearing on the mechanics of deglaciation. Since the geographic dimensions of the ice sheet did not undergo large areal reductions (Bryson et al, 1969), downdraw and vertical thinning of the ice sheets must have been substantial. The transition from a high-profile to a low-profile ice sheet may have greatly affected the mechanics of ice motion (e.g., Berglund, 1979). The outflow of low-salinity glacial meltwater into the ocean should also have inhibited oceanic turnover (Worthington, 1968), and both W. Berger (1977) and Ruddiman et al (1980) have cited evidence from deep-sea sediments to support this effect.

A further feature of the deglaciation involves a very rapid climatic cooling in some regions between 11,000 and 10,000 YBP. The North Atlantic polar front migrated southward almost to its glacial maximum position (Duplessy et al, 1981; Ruddiman and McIntyre, 1981b). Temperatures in western Europe were very cold (see summary in Duplessy et al, 1981). South American records also indicate a cool interval at this time (van der Hammen, 1974;

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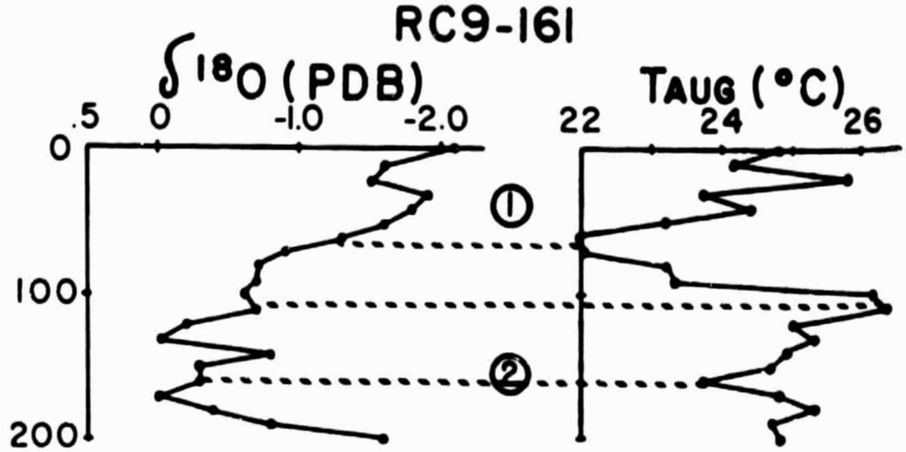

Fig. 43 An example of climatic patterns during the last deglaciation. This figure illustrates that both minimum <u>and</u> maximum SST for the last 20,000 years in the upwelling area of the Arabian Sea occurred during the deglaciation. Oxygen isotope stratigraphy and estimated August SST, plotted vs. core depth (from W. L. Prell, 1978). Isotopic stages 1 & 2 are circled. Upper dashed lines indicates latter part of the glacial termination (about 9K); middle dashed line indicates early part of the termination (about 14K), and lower dashed line indicates the last glacial maximum (18K). (Reproduced with the author's permission).

Heusser and Streeter, 1980). However, there is little evidence for any widespread re-advance of glaciers over eastern North America (Terasmae, 1980).

One of the more interesting conclusions concerning the nature of climatic change has been derived from studies of Early Holocene precipitation patterns in the Middle East, where maximum aridity <u>and</u> wetness (for the last 30,000 years) occurred on the transition to interglacial (i.e., the glacial termination). Street and Grove (1979) cite evidence for maximum North African aridity around 14-15K, about the same time that large volumes of melt-water were being added to the oceans (primarily the North Atlantic). Upwelling indicators off Saudi Arabia (Fig. 43) also indicate high SST, which in this area reflects decreased upwelling (Prell, 1978). A weak southwest monsoon can account for the latter pattern. By extension, South Asian rainfall may also have decreased.

The above Arabian Sea SST record also indicates that later in the termination (about 9K), the intensity of upwelling, and hence the strength of the southwest monsoon, increased to a maximum. At the same time, lake levels rose to a maximum in North Africa, with wet conditions prevailing from the Namib Desert at 26S to the northern tropics (Street and Grove, 1979), and eastward to northwest India (Bryson and Swain, 1981). Kutzbach (1981) has suggested that precession and tilt effects caused an enhanced solar radiation receipt during July (the rainy season). Modelling results indicate an increased low-level cyclonic inflow over the African-Eurasian land mass, with resultant increased precipitation. Thus, the model and the data support each other.

C-2

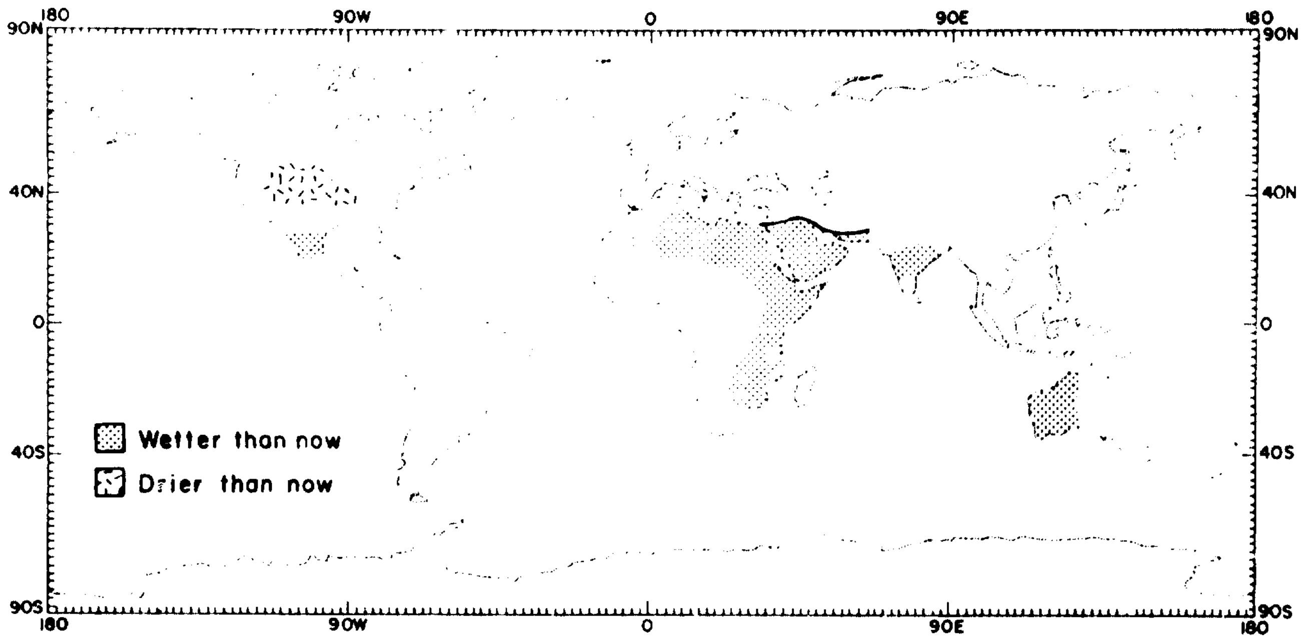

Fig. 44 A reconstruction of the continental climatic conditions during the Holocene Maximum (4500-8000 YBP). Data from Butzer (1980) and Kellogg (1977). (Reprinted with permission of Westview Press from CLIMATE CHANGE AND SOCIETY, by W. W. Kellogg and R. Schware).

Mid-Holocene Warm Period - A Possible Scenario for a CO_2-Induced Warming

The final disappearance of the Laurentide Ice Sheet at 7K ushered in an interval slightly warmer than the present in mid-latitudes. The event has been variously called the Holocene Maximum, Hypsithermal, Altithermal, etc. The thermal optimum for North America and western Europe occurred at this time (e.g., Bernabo and Webb, 1977; Wijmstra, 1978); the Great Plains were drier (Webb, 1980) and the Middle East wetter than the present (Fig. 44). SST peaked in the eastern tropical Atlantic (Kipp, 1976, Crowley, 1981). The climatic patterns from this time period are of particular climatic interest because they provide a possible picture of how the climate might differ under the influence of a CO_2-induced warming. The geological data suggest that two trends most likely to develop are increasing aridity over the Great Plains of North America and decreasing aridity in the Middle East (Kellogg and Schware, 1981). Some climate modelling results also suggest that Arctic pack ice cover may be significantly reduced (Parkinson and Kellogg, 1979; Manabe and Stouffer, 1980). However, geological data indicate that it is unlikely Arctic sea ice would completely disappear — Clark et al (1980) cite evidence for the presence of at least seasonal sea ice in the Arctic for the last five million years.

An additional consequence of a CO_2 - warming might be the disintegration of the marine-based West Antarctic ice sheet (Mercer, 1978). This event would raise world sea level about 5 m. It has been suggested (e.g., Mercer, 1968; Stuiver et al, 1981) that the disappearance of this ice sheet during the last interglacial maximum at 125,000 YBP (cf. Fig. 24, p. 46) may have been responsible for the higher sea level at that time (cf. Fig. 26, p. 48).

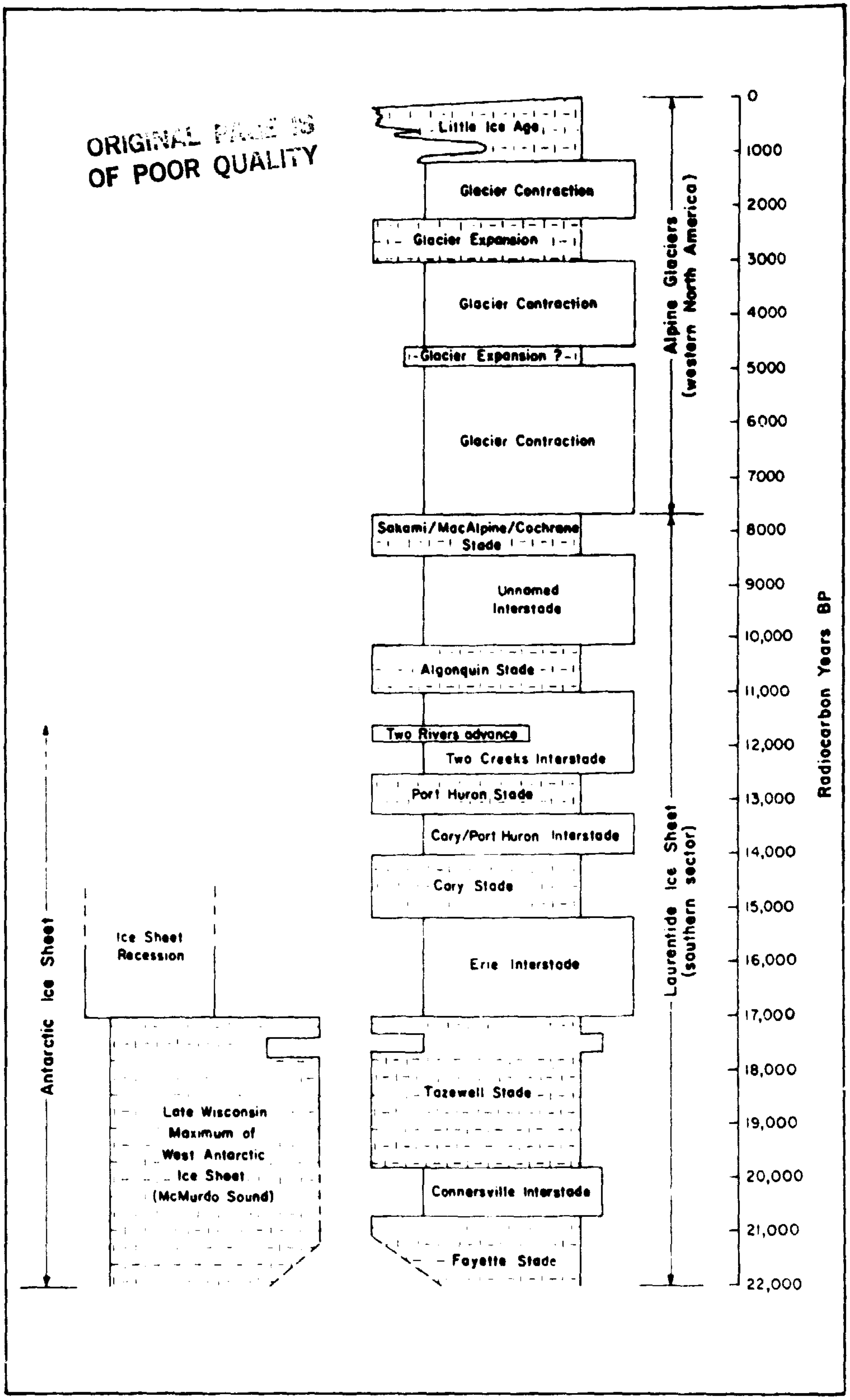

Fig. 45 Evidence for millennial-scale climatic fluctuations during the last 20,000 years. Schematic diagram of pulses of glacier acticity in North America superimposed on major Late Wisconsin and Holocene climatic changes. Antarctic information is from Stuiver et al (1981). Time scale is in radiocarbon YBP. (From Mayewski et al, 1981; reproduced with the permission of John Wiley & Sons, Inc.).

Another feature of a CO_2 - warming that warrants attention concerns the transient climatic response to the forcing. There is some reason to believe that climatic patterns during the transitional state may not be of an intermediate nature (Thompson and Schneider, 1979; Bryan et al, 1982). Indeed, the above review of the climatic response during the transition period of the last deglaciation is a source of evidence that would support such a conjecture.

Additional Features of Holocene Climate Although the general trend of Holocene climates during the last 4-5,000 years has been one of cooling (termed the Neoglaciation), higher-frequency fluctuations have occurred. There is also some indication of cyclicity in Holocene records. Mitchell (1976) cites some evidence for a 2500-yr. cycle in climatic records. In fact, Pisias et al (1973) have found this cycle in a North Atlantic temperature record extending back through the last 130,000 years. In addition, Mayewski et al (1981) have shown that millenial-scale fluctuations have occurred during the last 20,000 years (Fig. 45). Climatic oscillations (Fig. 46) of about 80, 180 and 350 years have been identified in ^{18}O records from the Greenland ice core (Johnsen et al, 1970; Dansgaard et al, 1971). The climate of the last 1000 years (Fig. 47) has been marked by a Medieval warming, followed by a cool interval (the Little Ice Age) from about 1430-1850 (Gribbin and Lamb, 1978; Bernabo, 1981). The subsequent warming that peaked in the 1940's (Mitchell, 1961) completes the picture. Evidence from North American tree rings indicates that a circulation type characterized by warmth and aridity in the west, and cold in the east, was much more common during the Little Ice Age

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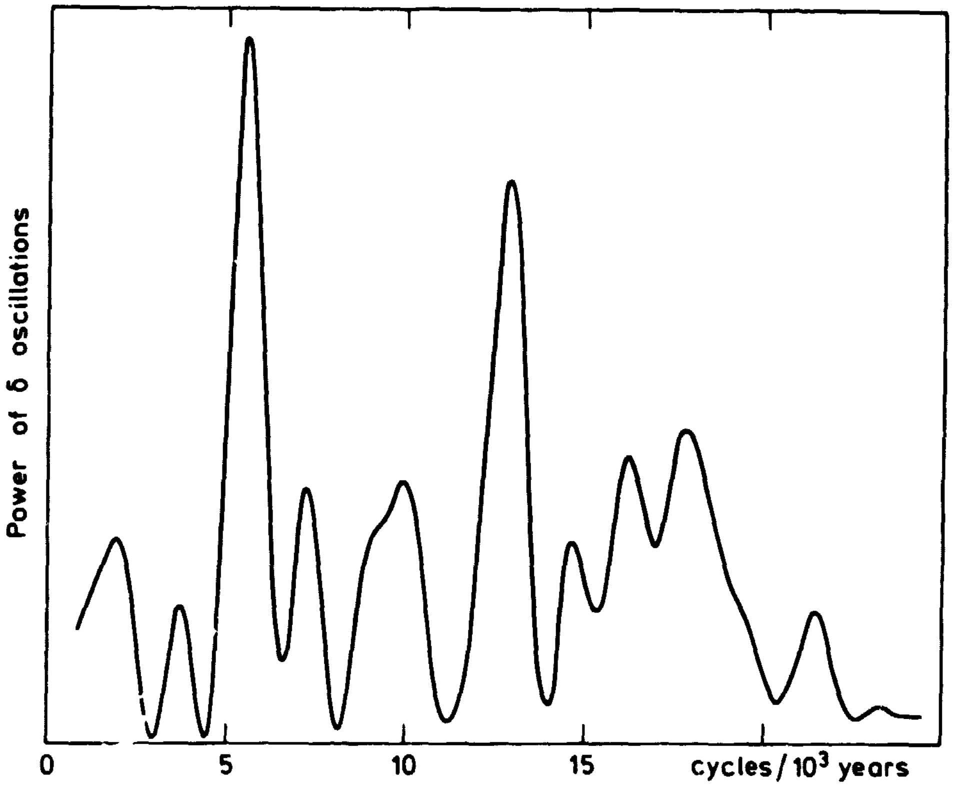

Fig. 46 Evidence for centennial-scale climatic fluctuations - the power spectrum of the oxygen isotope record from the Camp Century (Greenland) ice core. The two dominating peaks correspond to periods of about 80 and 180 years. (From Johnsen et al, 1970; reproduced with the permission of the authors and NATURE).

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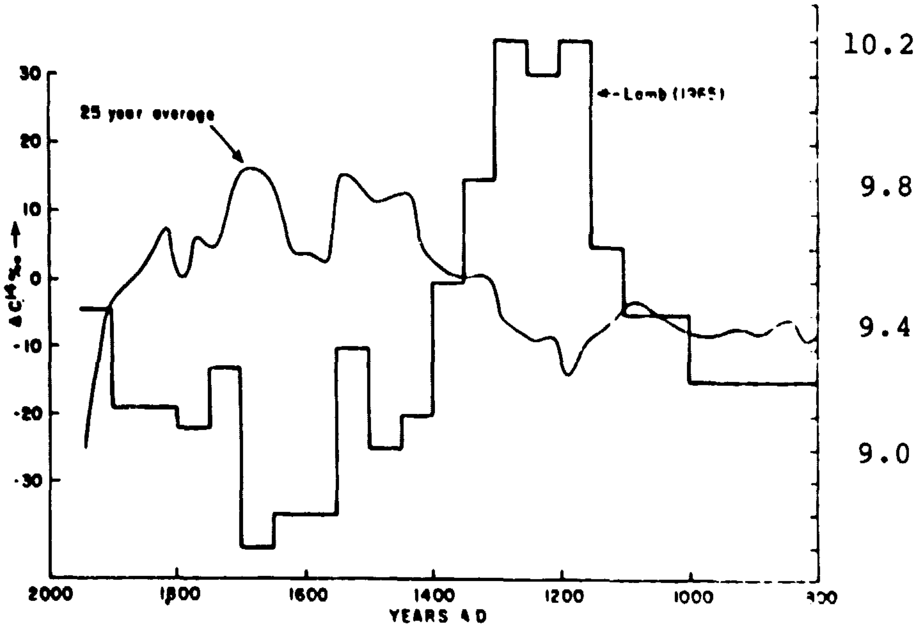


Fig. 47 C-14 variations vs. temperature during the last millennium. C-14 values are from data summarized by Damon <u>et al</u> (1978). Temperatures are from Lamb's (1965) estimates of average annual temperature prevailing in central England. Note the inverse relationship between the two records. (From Damon <u>et al</u>, 1978; reproduced, with permission, from the Annual Review of Earth & Planetary Science, v. 6. Copyright 1978 by Annual Reviews Inc.).

than during the 20'th century (Fritts et al, 1979). In fact, the circulation pattern appears to have been similar to the pattern for the severe winter of 1976–77 in the eastern U. S. (Namias, 1978).

Causes of the centennial-and millennial-scale climatic fluctuations of the Holocene are not well understood. Explanations have run the gamut from stochastic and volcanic effects to variations in solar activity (e.g., Lamb, 1970; Damon et al, 1978; Robock, 1978). The parallel between climate and ^{14}C production (Fig. 47) is particularly intriguing, because ^{14}C apparently reflects changes in solar activity (Stuiver and Quay, 1980). An active sun is associated with an increased deflection of galactic cosmic rays, with resultant decreased production of atmospheric ^{14}C. Denton and Karlen (1973) claim to have found similar correlations between ^{14}C and ice extent in records extending back to about 5,000 YBP. The coincidence of an apparent 2500-yr. cycle in ^{14}C production rates (Suess, 1980) and in paleoclimatic records is a further indication of a possible solar influence on climate.

IX. GEOLOGICAL EVIDENCE OF SOLAR VARIABILITY

In the previous sections of this review, the geologic record of climatic change has been catalogued, and causal mechanisms of terrestrial and extraterrestrial origin have been discussed. In addition to evidence for the influence of orbital variations on the earth's climate, there are also some indications for the influence of other types of solar variability (see, e.g., Newkirk and Frazier, 1982). It is the purpose of this section to review the geological evidence for any such variability.

The basis for the following survey involves historical evidence for solar variability. For example, at present there are indications for short-term variations in solar luminosity (Willson et al, 1981), possible changes in the solar radius (Eddy and Boornazian, 1979; Dunham et al, 1980), and long-term variations in sunspot number (Eddy, 1976). Stuiver and Quay (1980) have shown that some of the sunspot variations correlate to production rates of ^{14}C in the upper atmosphere during the last millennium (Fig. 48). There is also evidence for a cyclicity in the production rates of 120-and 140-years (Stuiver, 1980). Stuiver (1980) has noted that the rather orderly pattern to the fluctuations places some limitations on theories of solar dynamo behavior.

The geological evidence for solar variability involves three types of data: extraterrrestrial, and terrestrial records of annual- and longer-period fluctuations.

Extraterrestrial Evidence of Solar Variability Analyses of lunar samples and meteorites has focused on two types of evidence: heavy ion implantation tracks from solar flares, and production of cosmogenic nuclides (the reader is referred to Frazier, 1981, and Pepin et al, 1980, for a more

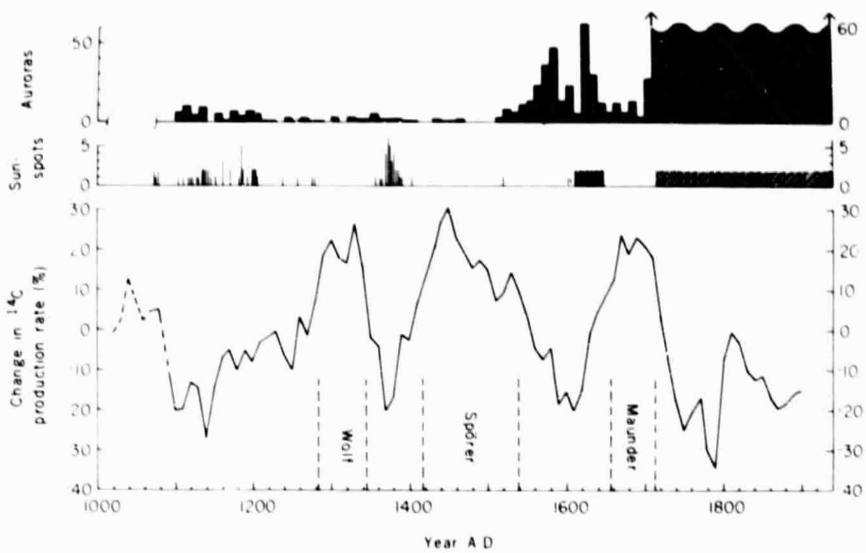


Fig. 48 Evidence for C-14 changes attributable to a variable sun. Frequency of auroral observations, sunspots, and changes in C-14 production rate calculated from a reservoir model. The number of naked eye sunspot observations per decade before 1610 is given by the vertical bars. The cross-hatched area denotes the existence of sunspots as observed by telescope. (From Stuiver and Quay, SCIENCE, 207:11-19, 4 January 1980; reproduced with the permission of the authors; copyright 1980 by the American Association for the Advancement of Science).

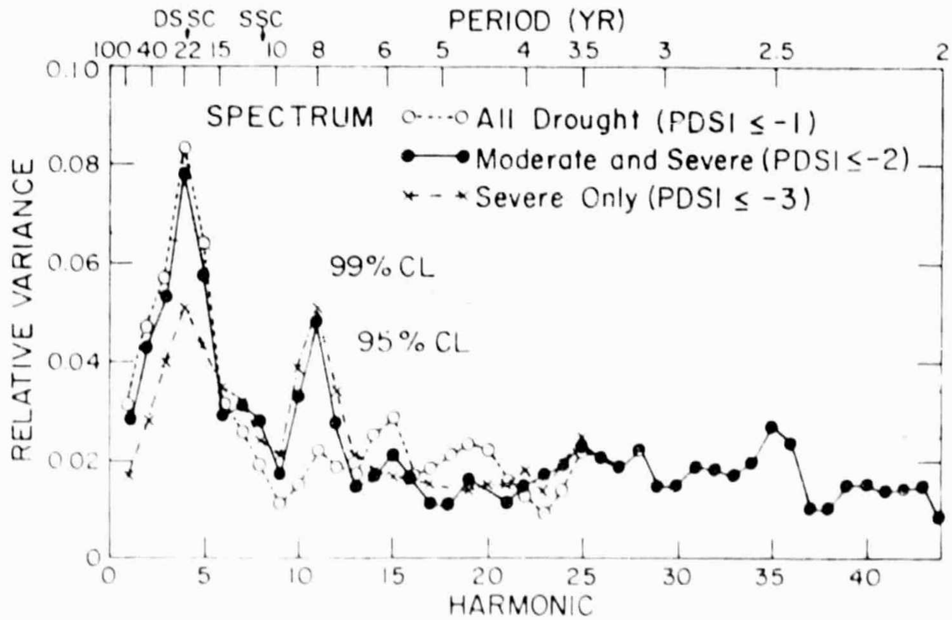
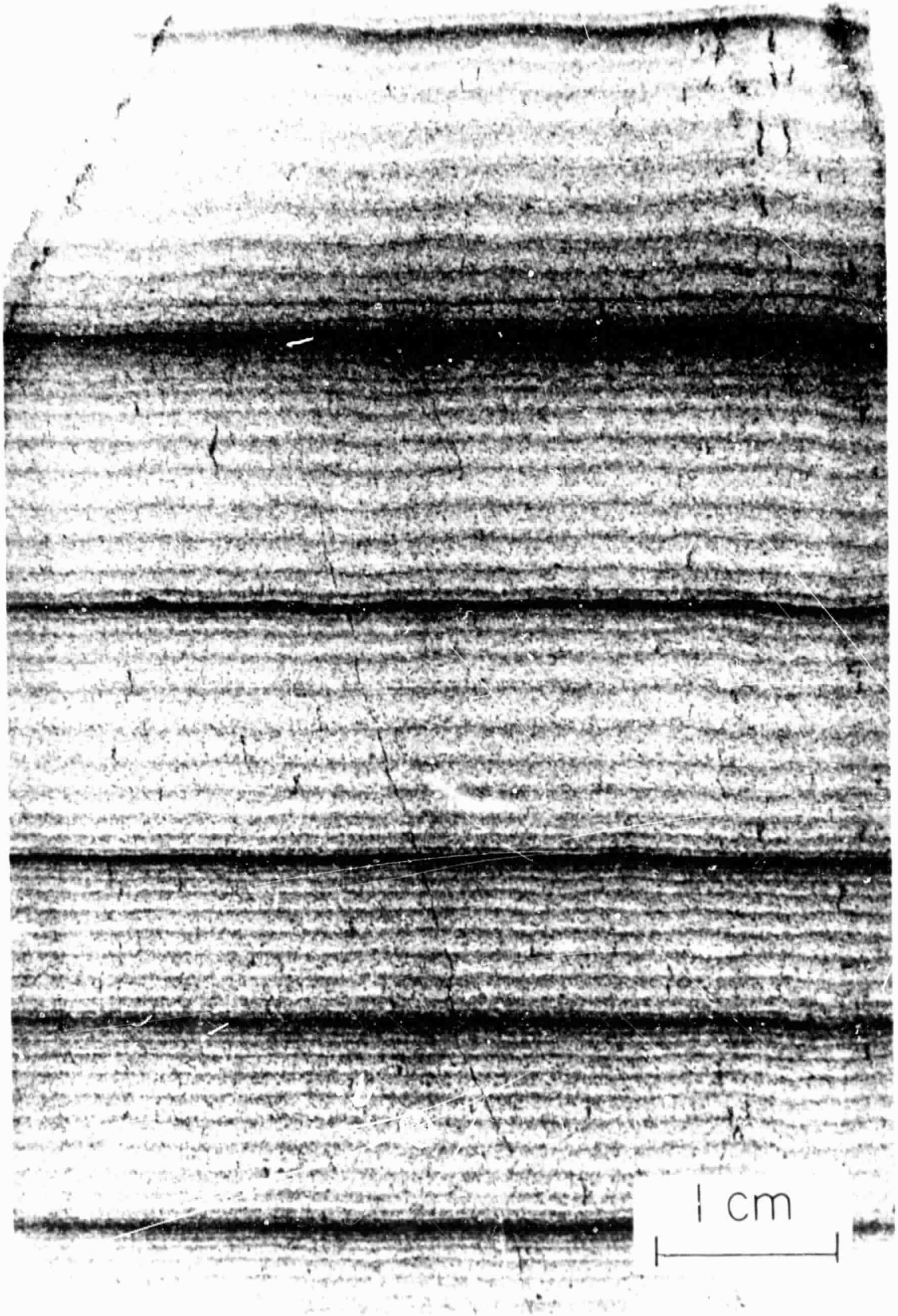


Fig. 49 Evidence for a 22-yr. drought rhythm in tree-ring records from the western U.S. Confidence limits based on assumption of first order Markhov process as a generating function. Abbreviations: PDSI (Palmer Drought Severity Index), SSC (sunspot cycle), DSSC (double sunsopt cycle). (Data by Mitchell et al, 1979; figure from Roberts, 1979. Reproduced with the permission of D. Reidel Publishing Company).

complete discussion of this section). Results yield conflicting evidence of solar conditions. For example, Crozaz (1980) has examined solar flare tracks and concluded that the energy spectrum of solar flare particles has not changed drastically during the last four billion years. Conclusions from cosmogenic nuclide studies are of a more variable nature. Abundances of ^{26}Al and ^{53}Mn imply little variations during the last ten million years (Reedy, 1980). Alternatively, preliminary analyses of ^{14}C and ^{81}Kr suggest rather large variations during the same time span (Reedy, 1980). The latter conclusion is supported by analyses of $^{15}N/^{14}N$ ratios in the lunar regolith, where there is now a well-documented (and unexplained) variation of 30% in the ratio from samples spanning the last four billion years (Clayton and Thiemens, 1980).

Terrestrial Evidence from Annual Records Some of the best evidence for a climatic response to solar variability is derived from tree ring records (see, e.g., Fritts, 1979). Mitchell *et al* (1979) have succeeded in documenting a 22-yr. drought rhythm in the tree-ring record of the western U.S. (Fig. 49). Epstein and Yapp (1976) have also identified the same cycle (i.e., the double sunspot cycle) in temperature-dependent deuterium-hydrogen ratios in wood cellulose. Hameed and Wyant (1982) have shown that the 22-yr. cycle can even be detected in records from the Maunder Sunspot Minimum (1660-1710); in fact, the cycle seems slightly stronger at that time than during later periods of increased solar activity.

In addition to the above records that cover the last 400 years, there are special types of geological deposits that occasionally record annual fluctuations. The records are from units that contain rhythmic bedding - i.e., alternate bands of different sediment types. In many cases the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alternations represent annual cycles in sedimentation, in which case the units are called "varves". The reasons for the annual variations are usually of climatic origin. For example, in many lakes, dark clayey bands accumulate during the winter months. In summer, increased plant production deposits lighter-colored shells in the sediment. The ability of the "sedimentary couplets" to faithfully record an annual cycle has been demonstrated in a number of studies. For example, Nipkow (1928; see Duff et al, 1967) was able to correlate annual successions in Lake Zurich with historical events -- thicker layers of sediment formed after "bank collapses" whose date of origin were known. Likewise, Flint (1971, p. 404) and Schove (1978) discuss ^{14}C-varve comparisons in longer sequences, and demonstrate less than a 10% error in age estimates based on varves.

The geologic record contains examples of different types of varved sediments at strategic intervals of the last two billion years (Fig. 1). Figure 50 illustrates what these varves look like in a 680 M.y.-old record from Australia. Results from the varved records at 50-, 680-, and 2000-M.y. ind' te cycles of 11-and 23-yrs., i.e., the sunspot and double-sunspot cycles (P ,30; Trendall, 1973; Williams, 1981). The cycles in the two older

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Fig. 50 Geological evidence for the presence of 11-yr. sunspot cyclicity in the Late Precambrian (680 M.y.). The cyclic record is from a record at Pichi Richi Pass in the Flinders Range of South Australia. Dark clayey bands separate groups of 11-14 pale sand/silt laminae interpreted as annual deposits or "varves". There are additional indications of strong climatic periods at about 22, 145, and 290 years, and a weaker period near 90 year. (Figure provided by G. E. Williams, The Broken Hill Proprietary Company, Ltd., Camberwell, Victoria, Australia. Reprinted with permission from G. E. Williams, NATURE, 291:624-628, 1981).

records indicate a degree of solar stability not always anticipated by models of solar evolution. In addition to the above cycles, some of the longer sedimentary sequences in the geologic record also show a cycle of approximately 20,000-yr. duration, i.e., the precession period (Bradley, 1930; Fischer, 1980; Anderson, 1982).

Some recent work on new varve locations offers the prospect of further detailed information concerning annual variability. Although results are not yet in from these locations, they are included here for future reference. There is a 250,000 year record from the Gulf of California (Schrader et al, 1980) that may contain information about fluctuations of such Pacific-wide phenomena as the Southern Oscillation and the El Niño. A recent project focusing on lakes in the East African Rift Valleys indicates the possibility of a varved record extending back several million years (Cromie, 1982). The latter region is of special interest because it comprises the time interval over which man evolved in the same region.

Terrestrial Evidence of Longer-Term Variability As mentioned previously, Stuiver (1980) has documented 120- and 240-yr. cycles in ^{14}C production rates. But the correlations of these fluctuations with climatic fluctuations during the last millennium has yielded mixed results. For example, in the North Atlantic sector, Stuiver (1980) has shown a statistically significant correlation between ^{14}C, the ^{18}O records from Greenland, and the severity of winter in European Russia (e.g., Fig. 47). In the northeast Atlantic, the most severe interval of the Little Ice Age (Lamb, 1979) occurred during the Maunder Sunspot Minimum (1660-1710). However, some other records do not show a statistically significant relationship between ^{14}C

and climate. Stuiver (1980) suggests that perhaps jet-stream oscillations, which also seem to relate to the 11-yr. solar cycle (Nastrom and Belmont, 1980), could provide some of the explanations for the irregularity of the spatial response.

The geologic record may also supply some intriguing evidence for another type of solar variability -- changes in the solar radius. Recently, Gilliland (1981) has analyzed a variety of solar observations to propose a 76-yr. modulation of the solar radius. This period is remarkably close to an 80-yr. cycle (Fig. 46) identified from a core on the Greenland Ice Sheet (Johnsen et al, 1970).

A new monitor of solar variability is presently coming on line and should provide additional valuable information in the future (Raisback and Yiou, 1980). ^{10}Be is another radionuclide produced in the upper atmosphere by cosmic rays. It has two advantages over ^{14}C. The isotope is not involved in the complex carbon cycle; rather, beryllium attaches itself to aerosols and is deposited with rain and snow. Ice cores are therefore an ideal repository. Additionally, ^{10}Be has a half-life of 1.5×10^6 yrs. (compared to 5.6×10^3 years for ^{14}C), and is therefore capable of providing access to earlier periods of earth history.

Mechanisms With the increasing documentation of solar periods in the geologic record, there is the inevitable question as to how solar variations might affect the climate. This question cannot be answered at the present time, although there have been significant advances in delineating possible pathways (see, e.g., McCormac and Seliga, 1979). In this section I will briefly discuss two models so that the reader may develop some sense of the types of explanations being advanced. A feature of both examples involves the non-reliance on total solar irradiance changes, which over short periods of time do not appear to exceed 0.2% (Willson et al, 1981).

One solution involves possible stratosphere-troposphere interactions (see, e.g., Holton, 1979, Chap. 11). For example, ozone production is affected by ultraviolet radiation. Measurements in the UV-band indicate significant historical changes in solar UV-output (Heath and Thekaekara, 1977; Cook et al, 1980). Increases in the UV-band are presumably compensated by slight decreases at other wavelengths, so that the total irradiance is still approximately constant. Toon et al (1980) have calculated that the UV fluctuations are sufficient to explain fluctuations in albedo of the atmospheres of Titan and Neptune (see Lockwood and Thompson, 1980). The connection presumably involves some photosensitive chemical reactions in the planetary atmospheres. The impact of UV fluctuations on the earth's climate is more problematical; however, Borucki et al (1980) have calculated that in some cases the effect could be significant.

An alternate approach to the sun-weather connection involves ionosphere-troposphere coupling (Markson, 1979; Markson and Muir, 1980). This mechanism involves variations in ionospheric charge that correlate with solar wind fluctuations. Since the ionospheric electric field is part of a global circuit, tied into the troposphere via thunderstorms, variations in the ionosphere electric field may modulate the electric field intensity in developing cumulus clouds. Markson and Muir (1980) note that cloud electrification may play an important role in coalescence of droplets and condensation of water vapor. The resultant release of the latent heat of condensation might then plan an important role in tropospheric dynamics. The unique feature of the above approach is that, as a consequence of increased latent heat release, climatic changes are effected by a more efficient conversion of potential energy to kinetic energy.

Summary Although there is geological evidence for some solar variability, there are also indications of an overall long-term stability in the sun. Some advances have been made in mapping possible features of the sun-weather connection; however, many details of the connection are still obscure.

X. SUMMARY AND CONCLUSIONS

At the beginning of this review, it was noted that erosion has been responsible for removing most of the rocks formed during earth history. Nevertheless, if we assume that the available evidence is a reasonably faithful reflection of the real earth history, then the following statements can be made concerning the geologic record of climatic change:

1. 4.6 - 2.3 B.y — the earth was apparently ice-free, despite a substantially lower solar luminosity. An enhanced atmospheric greenhouse effect may have compensated for the decreased insolation receipt.

2. 2.3 B.y. — evidence for the first glaciation seems to mark a threshold temperature through which the atmosphere passed.

3. 2.3 - 0.9 B.y. — the earth was apparently ice-free, despite low luminosity <u>and</u> a presumably depleted greenhouse effect. A suitable explanation for this phenomenon remains an important unattended problem in paleoclimatology.

4. 0.9 -0.6 B.y. — three major phases of glaciations occurred, with paleomagnetic data suggesting ice in low latitudes. This phenomenon has also not received much attention. It should be noted that because there is no evidence bearing on the presence of low-latitude sea-ice, low-latitude continental glaciation should not be construed to imply an ice-covered earth.

5. 600 - 100 M.y. — climates were generally mild, but punctuated by two major phases of ice growth. Paleomagnetic data indicate a strong correlation between the occurrence of continents in high latitudes and the formation of continental ice sheets.

6. 100 - 50 M.y. — mild, generally non-glacial climates prevailed. The Late Cretaceous (100 M.y.) interval has served as a testing ground for atmospheric and oceanic circulation models of a non-glacial world. Simple atmospheric circulation models cannot adequately account for relatively high polar temperatures. Additional factors (cloudiness, oceanic circulation changes) may be important for balancing the heat budget.

7. 50 - 3.0 M.y. — sequential cooling of the globe occurred. Two models of glacial history apply to the time period 50-3.0 M.y.: 1) formation of the Antarctic Ice Sheet about 14 M.y. and initiation of northern hemisphere glaciation about 3 M.y.; 2) formation of the Antarctic Ice Sheet as early as 30 M.y. The picture of Cenozoic climatic patterns may be substantially revised in the coming years, with the advent of high-quality hydraulic piston core records being the principal agent for the revisions.

8. 3.0 - 0.0 M.y. — numerous oscillations of northern hemisphere ice sheets occurred, with intervals as warm as today occupying only 10% of the Late Quaternary record. During the last glacial maximum (18,000 YBP), sea ice and ice sheets migrated into mid-latitudes. SST changes were generally small in regions not directly affected by sea ice. Equatorward of the continental ice margins, the surface of the planet dried out considerably. Late Quaternary time series indicate a strong correlation between ice volume fluctuations and orbital perturbations, but the nature of the interaction is presently not well understood. As much as 65% of the variance of Pleistocene climatic fluctuations appears to be due to internal processes operating within the land-sea-air-ice system.

9. A re-interpretation of the pre-Pleistocene oxygen isotope record implies that the frequency of glaciation through the last 600 M.y. may be significantly greater than the 15-20% figure suggested by direct glaciological evidence.

10. Terrestrial and extraterrestrial materials provide some evidence of variations in solar activity. However, the envelope of variability seems to be relatively small. For example, the energy spectrum of solar flare tracks in 4 B.y. lunar samples is similar to the present spectrum. Likewise, 11-and 22-yr. cycles have been detected in sediments as old as 680 and 2000 M.y.

The perspective afforded by an overview of the entire geologic record allows for some additional syntheses. The pattern of temperature change during the last four and a half billion years may be characterized as having both secular and fluctuating components. The secular variation in solar luminosity has caused changes in global insolation receipt on a time scale of 10^9 yrs. An early greenhouse effect may have offset the lower luminosity. Fluctuations of temperature involve mechanisms with characteristic time scales of 10^0 - 10^8 yrs. On time scales of 10^7 - 10^8 years, paleogeographic factors (e.g., continental drift, sea level changes) appeared to have played an important role in controlling fluctuations of global temperature. On a time scale of 10^6 yrs., changes in global volcanism may be important (Kennett and Thunell, 1975, 1977; Bray, 1979). Additional significant changes in the boundary conditions of the solid earth can occur in about 10^6 years. For example, the emergence of the Central American isthmus at about 3 M.y. may have had a profound impact on circulation patterns in the equatorial oceans (Luyendyk <u>et al</u>, 1972). The vertical uplift of the Andes by as much as

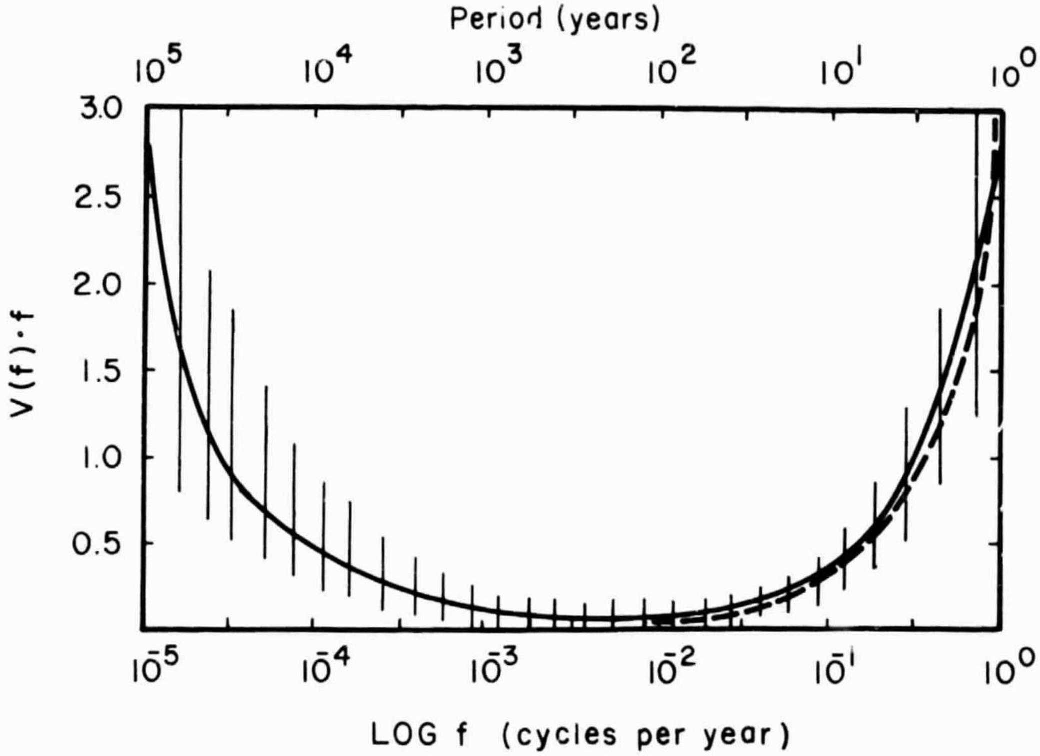

Fig. 51 Schematic version of the variance spectrum of temperature fluctuations in the North Atlantic sector on time scales of 10^0 to 10^5 years (after data from Kutzbach and Bryson, 1974; Hays et al, 1976a). The ordinate is V(f), variance spectral density, times f, frequency, with units of $(°K)^2$; the abscissa is a logarithmic frequency scale. These coordinates are chosen such that equal area under the curve represents equal variance. Vertical lines are used to indicate the relative degree of uncertainty in the shape of the curve. The dashed line indicates a white noise spectrum fitted approximately to the high-frequency portion of the spectrum. (From Kutzbach, 1976; reproduced with the permission of QUATERNARY RESEARCH).

1000-3000 m also occurred within a time span of about 1-2 million years (Simpson, 1975). Patterson and Pascual (1972) have suggested that the cutoff of moisture-laden Pacific winds to Patagonia probably accounted for the increased aridity observed in that area during the Pleistocene.

On a time scale of 10^3 - 10^5 years, the earth's climate appears to be sensitive to both external forcing by orbital perturbations and internal feedback interactions within the land-sea-air-ice system. Finally, on a time-scale of 10^0 - 10^2 yrs., the earth's climate appears to be sensitive to solar variability, volcanism (e.g., Mass and Schneider, 1977), and internal feedback interactions within the terrestrial system.

The above results can be summarized with a plot of climatic fluctuations across a broad range of frequencies (Fig. 51). Kutzbach (1976) has shown that a log-log plot of the variance spectrum in the North Atlantic sector indicates three main trends: a generally white-noise pattern at high frequencies that is characterized by relatively high variance; a mid-frequency pattern with distinctly less variance; and finally a lower-frequency record of increasing variance. The latter red-noise shape suggests an important stochastic contribution to factors responsible for climatic change.

A final analysis of the major results involves a comparison of the estimated changes in average global temperatures between non-glacial and full-glacial worlds. Values of + 6°C for the Cretaceous (Barron et al, 1981) and -4°C for the last glacial maximum (Gates, 1981) yield a total range of 10°C for the last 100 million years. This figure provides a useful frame of reference when evaluating the climatic impact of a CO_2-induced greenhouse warming. A doubling of the atmospheric CO_2 content is estimated to cause an increase of global temperatures by 2-3°C (e.g., Manabe and Stouffer, 1979, 1980; Hansen et al, 1981). This value is therefore about one-fourth of the total range estimated for the last 100 million years. The CO_2 effect is also about one-half of the range estimated for a glacial cycle. Furthermore, the expected rate of temperature change exceeds by an order of magnitude that which occurred during the catastrophic deglaciations of the Pleistocene.

The geologic record provides additional valuable information concerning the possible consequences of a CO_2-induced warming. Based on the climatic scenario reconstructed for the warmest period of the present Holocene Interglacial, the anticipated consequences of a CO_2-warming may include increased precipitation in North Africa and a drought on the North American plains. Another CO_2 - induced change in climate may involve an increase of about 5 m in world sea level, due to the melting of the West Antarctic Ice Sheet (Mercer, 1978). The slightly higher sea level during the last interglacial maximum at 125,000 YBP (cf. Figs. 24 & 26, pp. 46 & 48) indicates that such a response may be attainable with a relatively small change in temperature. Finally, atmospheric circulation models (Parkinson and Kellogg, 1979; Manabe and Stouffer, 1980) also predict changes in Arctic Ocean ice cover. The geological data appears to set an upper limit on the magnitude of this response — the Arctic Ocean has apparently been covered by at least seasonal pack ice for the last five million years (Clark et al, 1980).

In conclusion, direct evidence for major glaciations exists for perhaps 5 - 10% of earth history. However, indirect evidence (e.g., ^{18}O records) suggests that this figure could be significantly larger. The correlations of glacial events with paleogeographic changes implicates plate tectonics as an important underlying factor responsible for climatic change. However, Barron et al (1981) have shown that additional, unspecified, atmospheric-oceanic circulation processes are required to explain the occurrence of non-glacial climate during much of earth history. The onset of glaciations may be viewed as due to orbitally-induced fluctuations superimposed on long-term global coolings of terrestrial origin. Additional feedback processes are required to amplify the relatively small orbital signals into climatic responses of glacial magnitude. Stochastic processes appear to play an important role in the evolution of climatic fluctuations over the same time scale.

As for the future, research in paleoclimatology will continue to investigate the interactions between the internal terrestrial processes and external forcing due to orbital perturbations. In addition to the normal advances gained from acquisition of new data, climate models will also be tested against two types of very high-quality data: a geographic array of high-resolution Quaternary time-series (the objective of the SPECMAP Project), and very long Cenozoic time series (accessible via the hydraulic piston core). The data evaluation and model refinements will probably carry paleoclimatology through the present decade.

APPENDIX: ANNOTATED BIBLIOGRAPHY OF SELECTED REFERENCE WORKS

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An important book that contains a number of classic papers, which are still frequently referenced. In addition to providing a summary of past accomplishments in paleoclimatology, the book also set the tone for much of the research done in the subsequent decade.

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