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Northern Hemisphere Climate Regimes During the Past 3 Ma: Possible Tectonic Connections [and Discussion]

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Northern Hemisphere climate regimes during the past 3 Ma: possible tectonic connections

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The 100 ka rhythm of orbital eccentricity has dominated large-amplitude climatic variations in the high-latitude North Atlantic during the Brunhes magnetic chron (0–0.735 Ma BP). Earlier, during the Matuyama chron (0.735–2.47 Ma BP), climatic variations in this region were lower in amplitude and concentrated mainly at the 41 ka rhythm of orbital obliquity. These rhythmic climatic responses to orbital forcing are evident both in stable isotopic ($\delta^{18}O$) indicators of ice volume or temperature and in biotic and lithologic indicators of local North Atlantic surface-ocean variability. The synchronous responses of these indicators are consistent with results from atmospheric general circulation models showing that the North American ice sheet directly controls North Atlantic surface-ocean responses via strong cold winds that are generated on the northern ice-sheet flanks and blow out across the ocean, chilling its surface. Before 2.47 Ma BP, smaller-scale quasiperiodic oscillations of the planktonic fauna and flora occurred, but the cause of these variations in the absence of significant ice sheets is unclear.

Introduction

Analyses of $\delta^{18}O$ (a rough approximation of global ice volume) and of indicators of local climate near Northern Hemisphere ice sheets (sea-surface temperature and ice-rafted debris in the North Atlantic ocean) have demonstrated that Plio-Pleistocene ice sheets were primarily driven by orbitally controlled variations in insolation (Hays *et al.* 1976; Imbrie *et al.* 1984). To first order, the Brunhes chron (0–0.735 Ma BP) was a time of large-amplitude variation at the 100 ka period, the Matuyama chron (0.735–2.47 Ma BP) was a time of moderate-amplitude variation at the 41 ka cycle, and the Gauss chron (2.47–3.40 Ma BP) was a time of very small or non-existent Northern Hemisphere ice sheets. We review here some of the evidence for these distinct climatic régimes, with greatest focus on non-isotopic indicators.

Not yet explained are the changes between these régimes: the initiation of Northern Hemisphere glaciation at or near the Gauss–Matuyama transition; and the intensification of glaciation in the upper Matuyama and into the lower Brunhes. We review possible explanations for these changes in prevailing climatic régime, and further develop a recently proposed tectonic theory.

Evidence from the North Atlantic

Oceanic records contain three independent monitors of the amount of ice on land. Oxygen-isotope ratios ($\delta^{18}O$) from late Pleistocene cores in open-ocean areas are largely (*ca.* 70%) controlled by changes in global ice volume (Shackleton & Opdyke 1973), with North

American ice representing one half or more of the ice-volume component (Flint 1957). The amount of continental detritus ice-rafted into the high and middle latitudes of the North Atlantic is also a first-order measure of the size of ice sheets on surrounding continents, with North American ice dominating this signal because of geographic proximity and ocean circulation patterns (Ruddiman 1977). Estimates of sea-surface temperature (SST) in the high-latitude North Atlantic are a third indicator of ice volume; both geological evidence and climate modelling experiments isolate North American ice as the key control on this response as well (Ruddiman & McIntyre 1984; Manabe & Broccoli 1985). A full discussion of the rationale behind the use of percentages of $CaCO_3$, and estimated SST, as indicators of Northern Hemisphere (and particularly Laurentide) ice-sheet size is given by Ruddiman *et al.* (1986*b*).

The $\delta^{18}O$ record by no means entirely reflects changes in North American ice volume; it also reflects volumetric changes in other ice sheets, as well as variations in local temperature signals. Still, the great resemblance of the local % $CaCO_3$ and SST records in high-latitude North Atlantic cores to the $\delta^{18}O$ signal is strong confirmation that all three signals are good first-order measures of changes in North American ice volume during the Pleistocene and late Pliocene.

Initiation of Northern Hemisphere glaciation

The initiation of moderate-sized ice sheets in the Northern Hemisphere occurred at 2.40 Ma BP, although it was preceded by small increases in ice volume at *ca.* 2.55 Ma BP (Backman 1979; Shackleton *et al.* 1984; Zimmerman *et al.* 1985). This finding is based primarily on % $CaCO_3$ records from Deep-Sea Drilling Project (DSDP) Site 552 in the North Atlantic. The uniformly high carbonate values of the early Pliocene suddenly yielded to periodically much lower values as a result of influxes of ice-rafted continental debris (figure 1). In addition, the abrupt increase in $\delta^{18}O$ values at 2.40 Ma BP (figure 1) partly reflects the growth of ice sheets on land. Subsequent coring by the Deep-Sea Drilling Project and Ocean Drilling Project across the subpolar North Atlantic and in the Labrador and Norwegian Seas has confirmed 2.40 Ma BP as the age of onset of major ice rafting (Ruddiman, Kidd, Thomas *et al.* 1986; Eldholm, Thiede, Taylor *et al.* 1987; Srivastava, Arthur, Clement *et al.* 1987).

Other evidence suggests that the initiation of ice rafting to the North Atlantic Ocean at 2.40 Ma BP represents the culmination of a longer-term high-latitude cooling that began about 750 ka earlier. Progressive (but oscillatory) enrichment of $\delta^{18}O$ values in benthic foraminifera after 3.15 Ma BP (figure 1) suggests significant cooling of deep waters (Prell 1984; Weissert *et al.* 1984; Hodell *et al.* 1985; Keigwin 1986); depth-dependent decreases in $CaCO_3$ percentages in the North Atlantic during this interval (figure 2) have been attributed to increasingly corrosive deep waters (Ruddiman *et al.* 1986*c*), possibly due to the suppression of North Atlantic Deep water formation.

Progressive high-latitude cooling is also suggested by increasing percentages of cooler planktonic foraminifera (Loubere & Moss 1986; Raymo *et al.* 1986) and by decreasing discoaster abundances (Backman & Pestiaux 1986). Sea-surface temperature trends during this time interval are not easily reconstructed, however, because the assemblages are not analogous to modern faunas. A significant evolutionary turnover in planktonic foraminifera occurs in the late Pliocene near 2.40 Ma BP, probably in response to the effect of the earliest glaciations in cooling the surface North Atlantic and otherwise altering its structure (Raymo *et al.* 1986).

The history of ice-volume changes from 3.15 to 2.40 Ma BP is still controversial. Small-scale

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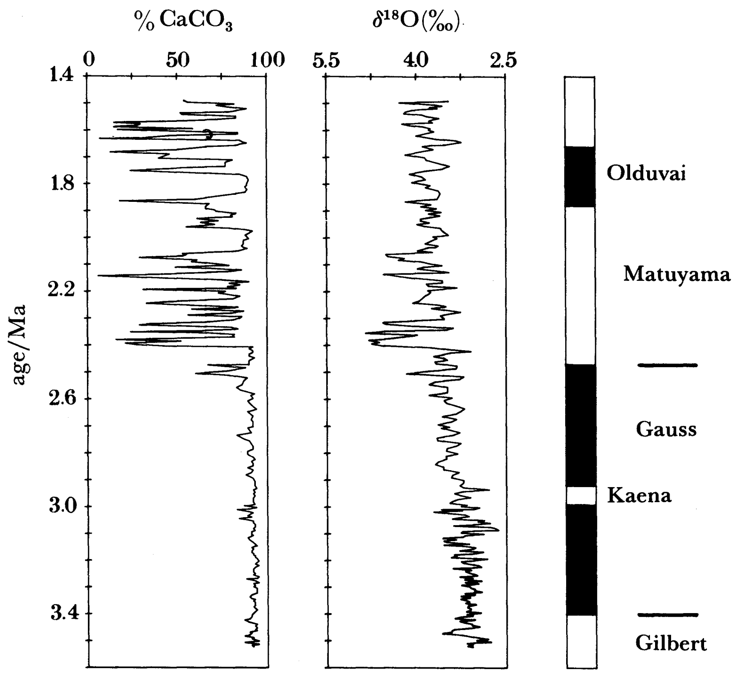

FIGURE 1. Late Pliocene and early Pleistocene % $CaCO_3$ and stable-isotope ($\delta^{18}O$) records at North Atlantic DSDP site 552 (after Shackleton *et al.* 1984). The abrupt decrease of % $CaCO_3$ values at 2.40 Ma BP marks initiation of deposition of ice-rafted sand. A correlative late-Pliocene increase in $\delta^{18}O$ values is superimposed on progressive long-term increase after 3.15 Ma BP.

glaciation began early in this interval on Iceland (McDougall & Wensink 1966) and in the Sierras (Curry 1966). Covariance of benthic and planktonic $\delta^{18}O$ values beginning at 2.90 Ma BP permits (but does not require) somewhat larger storage of ^{16}O-rich ice on land (Prell 1984; Keigwin 1986). It is plausible that Northern Hemisphere ice sheets periodically existed before 2.40 Ma BP but were too small to send ice-rafted detritus to the Atlantic. In any case, all the evidence is consistent with a progressive, but oscillatory, deterioration of Northern Hemisphere climate from 3.15 Ma BP (or earlier) until glaciations of substantial scale abruptly began at 2.40 Ma BP.

Recent evidence from North Atlantic DSDP sites 607 and 609 shows that Northern Hemisphere ice sheets fluctuated mainly at a period of 41 ka during the Matuyama (Ruddiman *et al.* 1986*b*). Variations in % $CaCO_3$, estimated sea-surface temperature, and benthic foraminiferal $\delta^{18}O$ were all dominated by the 41 ka rhythm of orbital obliquity (tilt) during the Matuyama magnetic chron (figures 3 and 4). Strong variance near 41 ka is evident from timescales based solely on linear interpolation between paleomagnetic datums (figure 4).

For the post-Olduvai (after 1.66 Ma BP) portion of the Matuyama chron at sites 607 and 609, we also developed an obliquity timescale by tuning the climatic indicators directly to the orbital obliquity signal. The obliquity timescale imposed age changes (relative to the magnetic time scale) no larger than 20 ka at any point in the records, and it predicted ages for the magnetic reversals (upper and lower Jaramillo and upper Olduvai) that agreed with the radiometric ages to within less than 1% of the absolute values. For the obliquity timescale, more than one half of the variance in these late-Matuyama records was narrowly centred at the 41 ka period (figure 4).

The obliquity rhythm is also prominent in the lower Matuyama (pre-Olduvai) portion of the

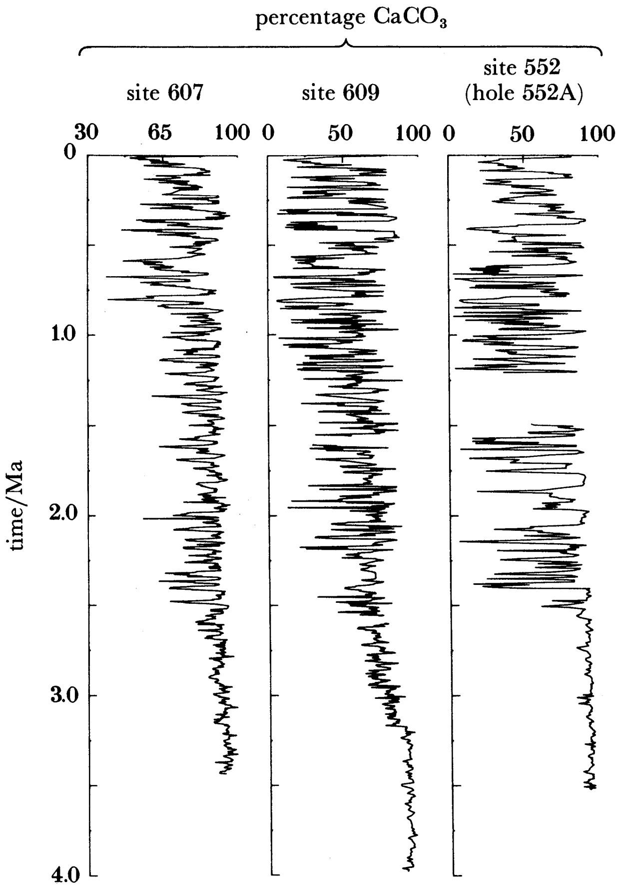

FIGURE 2. Plio-Pleistocene % $CaCO_3$ trends at North Atlantic DSDP sites 552, 607 and 609 (from Ruddiman *et al.* 1986*c*). All records show initiation of major glaciation by carbonate decreases at 2.40 Ma BP. Deeper cores (site 607, 3500 m; site 609, 3800 m) show progressive decrease of % $CaCO_3$ values between 3.15 and 2.40 Ma BP, indicating increasing dissolution with depth.

records from sites 607 and 609 (Ruddiman *et al.* 1986*b*). We did not, however, attempt to tune these records because of complications in assembling continuous records owing to coring problems. Our recent (unpublished) work confirms these early indications that the 41 ka rhythm was dominant in the early Matuyama as well.

Before 0.9 Ma BP, the dominance of 41 ka power in all signals is unequivocal and overwhelming (figure 3). After 0.9 Ma BP, the signals become more complex, with higher-amplitude fluctuations, and probably lower-frequency components as well. These changes mark the beginning of the shift into a different climatic régime during the Brunhes (Shackleton & Opdyke 1973; Pisias & Moore 1981; Prell 1982).

Because the climatic signals after 0.9 Ma BP differ from those earlier in the Matuyama, they complicate the spectral analysis shown in figure 4 by introducing effects of non-stationarity (changes in signal mean and variance). To assess this effect, we re-ran the spectral analyses of all four climatic indicators for the interval from the upper Olduvai boundary to 0.9 Ma BP. The only significant changes in relative concentration of spectral power were at very long periods (greater than 250 ka); reductions were minor for site 609 % $CaCO_3$ and site 607 $\delta^{18}O$, but larger (30–40 %) for site 607 % $CaCO_3$ and SST. Most of the low-frequency variance that

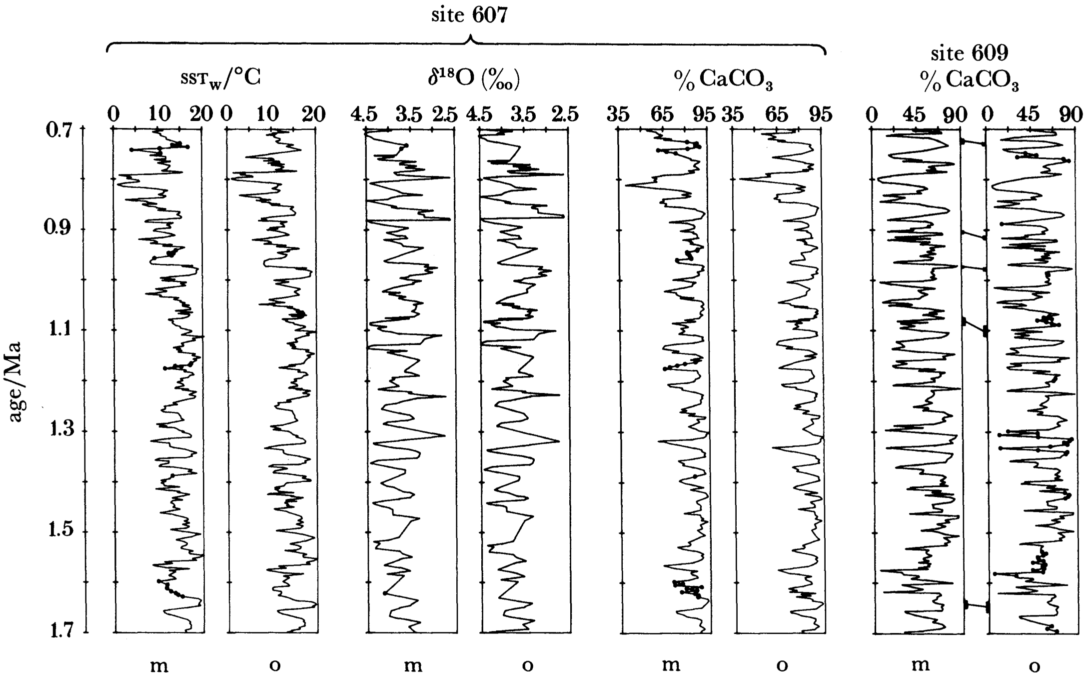

FIGURE 3. Early Pleistocene climatic trends at North Atlantic DSDP sites 607 (estimated winter sea-surface temperature (SST_w), $\delta^{18}O$, and % $CaCO_3$) and 609 (% $CaCO_3$); after Ruddiman *et al.* (1986*b*). Magnetic timescale based on linear interpolation between magnetic datums; obliquity timescale derived by tuning records to obliquity. Large dots are values spliced into main sequence from offset holes.

remained was concentrated at periods of 150 ka to 300 ka, apparently intermediate between the orbital eccentricity bands at 413 ka and 95–124 ka and thus of uncertain origin. For periods of 100 ka years or less, the late-Matuyama spectra were essentially unchanged, although the prominent 41 ka obliquity signal contains an even larger fraction of the total variance because of reduced power at the very low frequencies.

Intensification of Northern Hemisphere glaciation

After 0.9 Ma BP, changes in $\delta^{18}O$, % $CaCO_3$, and estimated sea-surface temperature increased in amplitude by *ca.* 50%; this increase suggests that ice sheets in the Northern Hemisphere during the late Matuyama and early Brunhes chrons grew to maximum volumes considerably larger than those attained during the early Matuyama. The first prominent $\delta^{18}O$ maximum occurred in isotopic stage 22 at 0.85–0.8 Ma BP (Shackleton & Opdyke 1973), an age correlating with the 'Nebraskan' glaciation in the magnetically reversed interval between 0.91 and 0.735 Ma BP (Boellstorff 1978; Rogers *et al.* 1985).

There is some regional variation in the SST and % $CaCO_3$ expression of the shift from the Matuyama to Brunhes climatic régimes. At mid-latitude site 607 (41° N), the largest SST and % $CaCO_3$ changes occurred at *ca.* 0.9–0.8 Ma BP (figure 3). At high-latitude site 552 (56° N), glacial SST minima also deepen at and after 0.9–0.8 Ma BP, but there is a further doubling in amplitude of SST fluctuations at a period of 95 ka in the middle of the Brunhes chron at 0.45 Ma BP (figure 5). $\delta^{18}O$ spectra show a similar doubling of variance at a period near 100 ka after 0.4 Ma BP (Imbrie 1985). Despite regional differences in SST response, all three climatic indicators (SST, % $CaCO_3$, and $\delta^{18}O$) show increasing variance after 0.9 Ma BP at

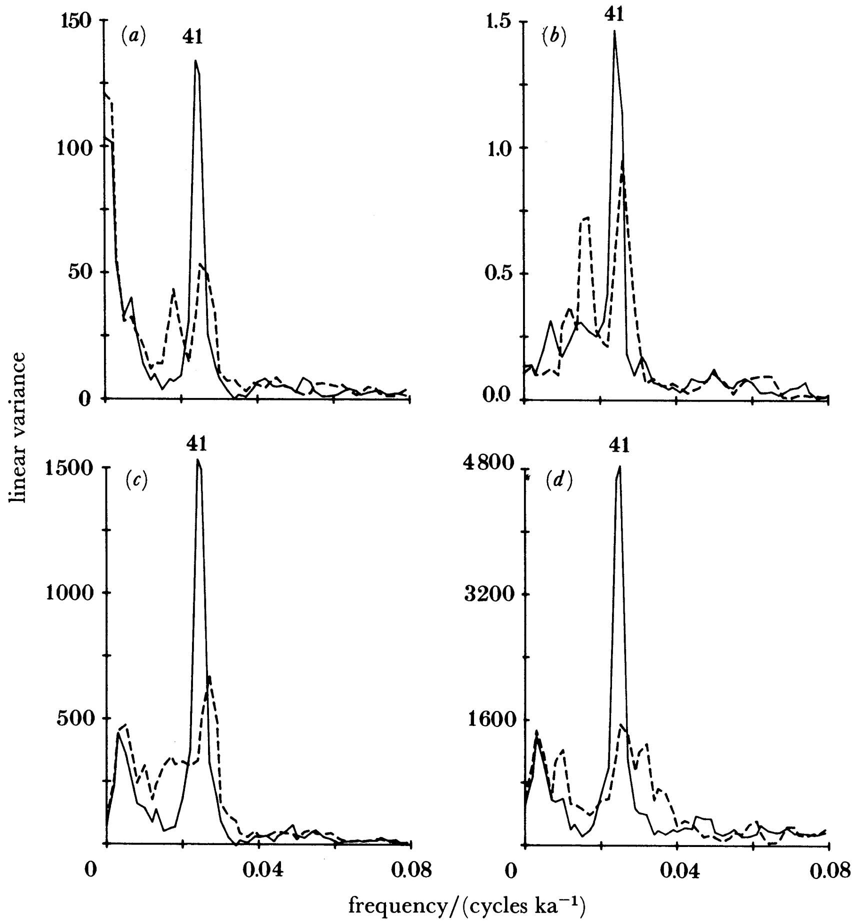

FIGURE 4. Spectral analysis of early Pleistocene records shown in figure 3 (from Ruddiman *et al.* 1986*b*). Spectra from both magnetic (broken lines) and obliquity timescales (solid lines) show prevalent 41 ka periodicity during the late Matuyama. (*a*) Site 607, SST_w; (*b*) site 607, $\delta^{18}O$; (*c*) site 607, % $CaCO_3$; (*d*) site 609, % $CaCO_3$; age range for all sites, 0.735–1.64 Ma BP.

periods of orbital eccentricity (near 100 ka) and precession (23 ka and 19 ka), and all indicate that the 100 ka rhythm attained strongest dominance in the late Pleistocene after 0.45 Ma BP.

Some uncertainty remains as to the form of the climatic response near the Brunhes–Matuyama boundary. SST records from site 552 suggest a brief interval (0.775–0.625 Ma BP) of prominent fluctuations at a period of 54 ka and a smaller coincident increase in 54 ka power also occurs in the SPECMAP $\delta^{18}O$ signal (Ruddiman *et al.* 1986*a*). It is not yet clear whether this 54 ka signal is real. It could have resulted from gaps in the single-HPC (hydraulic piston core) record cored at site 552, from incorrect correlation of the site 552 $\delta^{18}O$ record to the SPECMAP timescale, or from an error in the pre-Brunhes part of the SPECMAP timescale. More work is needed to resolve this important transition between the 41 ka (Matuyama) and 100 ka (Brunhes) régimes.

In summary, ice sheets large enough to send large amounts of ice-rafted debris to the open North Atlantic ocean did not exist before 2.40 Ma BP. There is evidence of Northern Hemisphere cooling after about 3.15 Ma BP, but major ice sheets only appeared by 2.40 Ma BP. The first 1.6–1.7 Ma of the Northern Hemisphere ice age (roughly the Matuyama

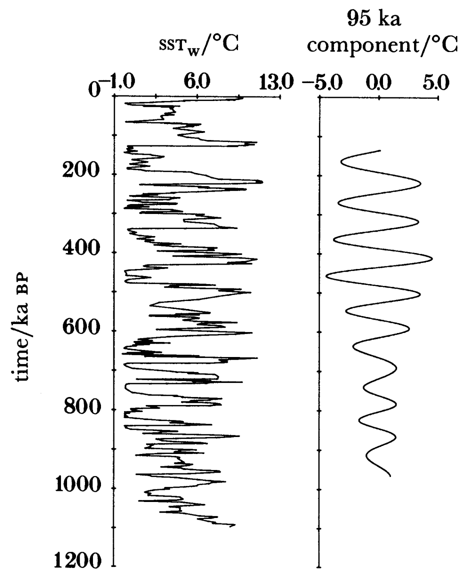

FIGURE 5. Late-Pleistocene record of estimated winter sea-surface temperature (SST_w) from spliced record of piston core K708-7 and DSDP site 552 (after Ruddiman *et al.* 1986*a*). Filtered record of dominant 95 ka SST component shows major increase in variance at 0.45 Ma BP.

chron) was largely a sequence of some 40 climatic cycles at the 41 ka rhythm of orbital obliquity. During the mid-Pleistocene (0.9–0.4 Ma BP), the obliquity rhythm was progressively superseded by an increased response at or near the 100 ka period of orbital eccentricity, and this rhythm dominated climatic responses throughout the Brunhes chron, particularly after 0.45 Ma BP. In addition, numerous glacial advances at periods of 41 ka and 23 ka were superimposed on the basic 100 ka cycle during the Brunhes chron.

This long-term perspective raises four fundamental questions: (1) Why did the Northern Hemisphere ice age begin? (2) Why did the Matuyama ice-sheet variations occur at a rhythm of 41 ka? (3) Why did the mid-Pleistocene shift in rhythmic response occur? (4) What is the origin of the 100 ka rhythm of the late Pleistocene? In this paper, we shall address a possible reason for the major changes in prevailing climatic régime during the late Pliocene and mid-Pleistocene (questions 1 and 3).

THEORIES OF ICE-AGE INITIATION AND INTENSIFICATION

Several explanations have been advanced to explain why glaciation in the northern hemisphere was restricted to the late Neogene. One group of theories looks to changes in atmospheric composition or in solar radiation. Some involve variations in parameters that are untestable, such as changes in solar strength (Opik 1958). Other plausible theories, such as changes in atmospheric CO_2, ozone or trace gases (Plass 1956), although also not yet tested, should be at least partly verifiable from geological data. Variations in orbitally driven insolation (Milankovitch 1941) are obviously critical on time scales of 20 to 100 (or 400) ka, but the mean long-term form and strength of this forcing (Berger 1984) did not change sufficiently to explain the glacial initiation or intensification discussed in the last section. Increased volcanism during the latest Cainozoic (Kennett & Thunell 1975) has also been suggested as a possible cause of glaciation, but vulcanism does not appear to provide persistent enough forcing (i.e. decadal repetition) to maintain the climate system in one 'régime' for intervals lasting a million years.

Other theories call on tectonic changes to explain glaciation. Ewing & Donn (1956) suggested that global albedo increased during the late Cainozoic, owing to relative movement of the poles and continents. Virtual polar wander is, however, thought to have been negligible (Schneider & Kent 1986), leaving this theory dependent on the unlikely possibility of a decoupling of the spin and magnetic poles. Changes in land–sea distribution by sea-floor spreading are of the order of 10–100 km during the last 3 Ma and are mostly latitude-parallel. This is probably too small a change to have had a major impact on climate, although North *et al.* (1983) suggest a possible link in which the ocean increasingly moderates summer radiative heating (and thus ablation) in the mid-continents, until some critical threshold or discontinuity is passed and glaciation begins.

Changes in susceptibility to glaciation could also be caused by epeirogenic uplift of highland regions in northern Canada (Flint 1957; Emiliani & Geiss 1958; Birchfield *et al.* 1982), although such changes in elevation are not well enough dated to relate unambiguously to glacial initiation and intensification. The consensus viewpoint today is probably that continental rearrangements and uplift have contributed to long-term cooling since the Cretaceous (Barron 1981) but are not very convincing as explanations for the more sudden initiation and intensification of glaciation in the Plio-Pleistocene, unless there are extremely sensitive thresholds of response in the climate system.

Another proposed tectonic explanation involves changes in oceanic heat and moisture fluxes (Emiliani *et al.* 1972) induced by shoaling or deepening of critical oceanic gateways such as the Panama Isthmus (Keigwin 1978, 1982) or the Bering Straits (Einarsson *et al.* 1967). Both of these tectonic changes occurred at 3.5–3.0 Ma BP, somewhat before the late-Pliocene initiation of significant glaciation at 2.40 Ma BP (figure 6), but very close to the beginning of the long-term late Pliocene climatic deterioration at 3.15 Ma BP. Neither of these tectonic changes could be causally involved in the mid-Pleistocene glacial intensification from 0.9–0.45 Ma BP.

We also believe that the changes in response of the climate system during the Pliocene and Pleistocene could not have been generated spontaneously within the internal physical variability of the ice–ocean–atmosphere system, but must be sought in tectonic changes that altered the solid boundary conditions within which that system operates. We discuss below two theories that ascribe Plio-Pleistocene climatic change to large-scale uplift in two key regions: Tibet and western North America. In both of these regions, there is abundant evidence of major late Cainozoic uplift that has continued through the Pleistocene (figure 6) and even into the last century. This evidence is summarized below.

Evidence for Plio-Pleistocene uplift in Tibet and North America

Although one geophysical interpretation holds that the Himalayas and Tibet have been high-standing since the Oligocene–Miocene orogeny because of continuous ('steady-state') underthrusting of the Indian continent (see, for example, Seeber & Armbruster 1983), other evidence favours the view that *ca.* 3 km of net uplift of the land surface has occurred during the Plio-Pleistocene (Liu & Menglin 1984). Early Pliocene pollen assemblages in both Tibet and the Himalayas contain mostly subtropical and warm deciduous flora indicative of elevations of 1.5 km or less (Hsu 1978); these deposits occur today at elevations of 4 km or more. Mammalian distributions (the *Hipparion* fauna) indicate that the Himalayas and Tibet

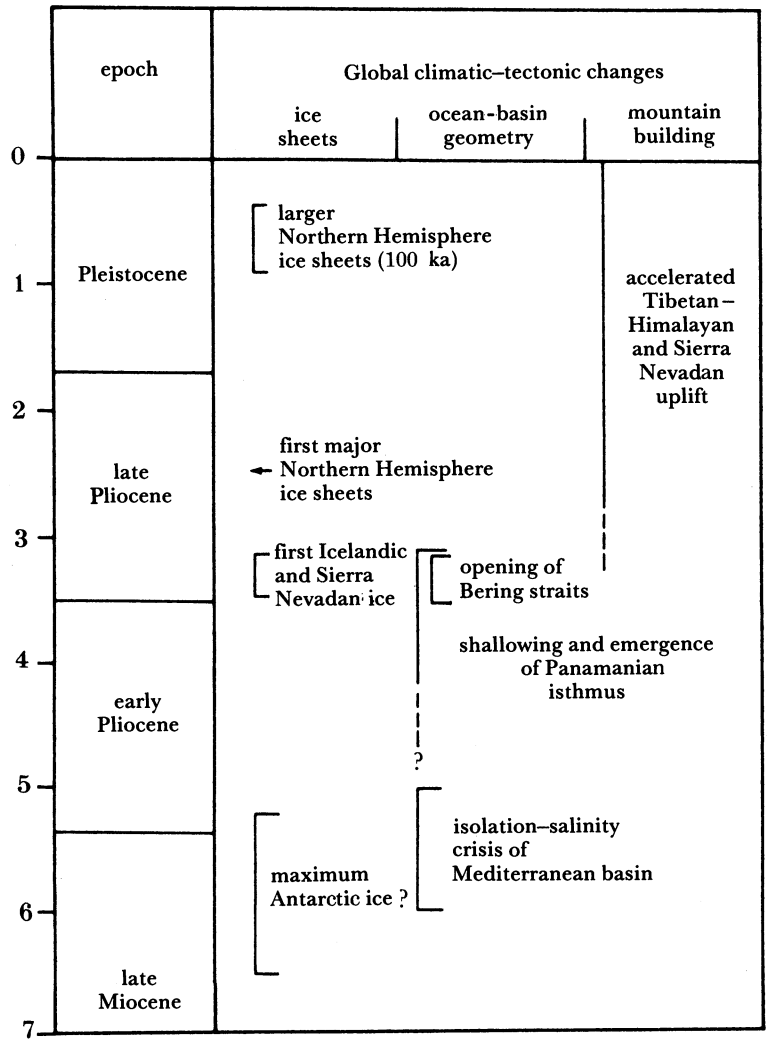

FIGURE 6. Timing of large-scale tectonic changes potentially important to Northern Hemisphere glaciation.

were not an effective barrier to north–south faunal exchange until the late Pliocene (West 1984). The Siwalik Formation at the southern foot of the Himalayas coarsens progressively upwards from early Pliocene silts to middle and late Pleistocene conglomerates, consistent with increasing uplift through the past 3 Ma.

The pollen data suggest mean rates of net uplift of just under 1 mm per year over the last 3 Ma. This agrees with the average rates of uplift in this century, determined from geodetic measurements in the Tibetan–Himalayan region (Mehta 1980). Radiometric data (Rb–Sr, K–Ar, and fission-track ages) also suggest mean rates of Himalayan uplift of 0.7–0.8 mm per year during the Miocene, Pliocene and Pleistocene (Mehta 1980), with some data suggesting progressively larger uplift rates during the Plio-Pleistocene (Saini *et al.* 1979) and, in some regions of the High Himalayas, rates as high as 3.5–10 mm per year during the late Pleistocene (Burbank & Johnson 1983).

Mineralogical data cannot be used to specify absolute elevations of the earth's surface in the past, because the competing effect of erosion in lowering the rising surface is not accurately

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known. Still, the mean Plio-Pleistocene uplift rates derived from the radiometric data support the rates estimated from pollen and faunal data if erosion is negligible, and the highest uplift rates estimated radiometrically would permit both rapid net uplift and substantial erosion of the High Himalayas.

In western North America, the time of most active tectonism and uplift substantially predates the Plio-Pleistocene glaciations. However, evidence from several regions in the Southwestern United States (particularly Colorado and Nevada) suggests that substantial net uplift has occurred in the Pliocene and Pleistocene as well.

Winograd *et al.* (1985) found a progressive 60‰ Pleistocene depletion of deuterium values in calcitic veins dated by U-series methods in the Sierra Nevada and Transverse Ranges of the Great Basin area. If this depletion is entirely attributed to increased elevation during the Pleistocene, and if modern lapse rates and deuterium profiles in this region are used as a basis for calibration, the 60‰ depletion would require 1 km of net uplift during the past 2 Ma.

Other evidence suggests major Plio-Pleistocene uplift of the southern Colorado Plateau. As recently as the late Miocene, the southern part of the modern Colorado Plateau was a relative topographic low, receiving debris from higher topography in Arizona (McKee & McKee 1972). Lucchitta (1979) used uplift of sediments, inferred to have been deposited in estuaries at sea level and in lakes near sea level, to suggest that the southern Colorado Plateau has attained some 800–900 m of its 2000 m mean height since 5.5 Ma BP. Of this total amount, 500 m is attributed to broad regional uplift and the rest to more localized fault-related movement along the plateau margin. Thus, somewhere between 25 and 45% of the present height of the southern Colorado Plateau was attained during the Pliocene and Pleistocene.

Significant Pliocene uplift is also indicated in northern Colorado (Izett 1975), but absolute values are not known. In general, it is easier to detect relative uplift on a local scale (by vertical offsets of strata) than to detect broader regional uplift. As a result, absolute late-Cainozoic uplift is not well constrained in the western U.S.A.

In summary, pollen and faunal data argue for a net uplift of 2–3 km in Tibet and the High Himalayan Mountains during the past 3 millions years, and at an average rate consistent with those measured geodetically over the past 70 years. Thus, as much as 75% of the net elevation of Tibet (an area approximately one third the size of the contiguous United States) may have been attained during the time of Northern Hemisphere glacial inception and intensification. Although significant net Plio-Pleistocene uplift (several hundred metres to 1 km) has also occurred in several regions of the southwest United States, these represent lesser alterations of a region with a long earlier history of high-standing mountains.

Taken together, these Plio-Pleistocene tectonic alterations may have had two effects on Northern Hemisphere climate that could be relevant to glacial inception and intensification. One involves major increases in zonal mean albedo (Birchfield & Weertman 1983); the other, large-scale rearrangement of the planetary wave structure (Ruddiman *et al.* 1986*b*).

Effects of uplift on zonal mean albedo

Using numerical experiments with a zonally averaged energy-balance model, Birchfield & Weertman (1983) proposed that uplift in Asia and North America would enhance albedo–temperature feedback on a globally significant scale. The cold air over the large, newly created region of high-standing topography at middle latitudes provides a temperature discontinuity that accelerates the normal southward movement of the snowline in autumn–winter. This enlarged high-albedo surface in turn leads to global cooling during the transitional and winter seasons.

Birchfield & Weertman did not attempt to resolve fully the sensitivity of the climate system to this albedo–temperature feedback; their experiments used numerous simplifying assumptions and could not simulate important geography-dependent features of the atmospheric circulation, such as the Asian monsoon. Their results thus pertain only to zonal mean climate, and it is not clear what specific effects these changes would have in the regions of North American and Eurasian glaciation. Nevertheless, the proposed albedo–temperature feedback is a plausible link between mountain uplift in Asia and North America and glaciation in the Northern Hemisphere.

Effects of uplift on planetary waves

Ruddiman *et al.* (1986*b*) suggested a second explanation for the initiation and intensification of Northern Hemisphere glaciation. This explanation specifically links changes in mid-latitude orography to climate in the glaciated regions of North America and Eurasia. They proposed that increased elevation in the Tibetan–Himalayan and Sierran–Coloradan regions have altered the planetary wave structure in such a way as to cool the North American and European land masses and increase their sensitivity to orbitally driven insolation changes. The concept is further developed and evaluated here.

At the simplest possible level, both the late-Pliocene initiation of glaciation and the mid-Pleistocene intensification require the same thing: some means of permitting larger ice sheets to grow, during favourable orbital configurations, than had previously been possible. We avoid here detailed discussion of the complex mechanisms involved in explaining the dominance of specific orbital rhythms during specific intervals. Instead, we focus on possible tectonic alterations of the planetary wave pattern that would permit faster (and thus greater) ice growth during glacial cycles.

Orographic control of the planetary waves

Maps of geopotential height or pressure distribution in the upper troposphere show that the middle-latitude westerlies are concentrated in narrow bands (including the jet stream) that are perturbed into large, wave-like meanders (the 'planetary waves'). The mean positions of the modern planetary waves in winter are shown in figure 7 (top left). Prominent southward meanders occur over east-central North America, western Europe, and the western North Pacific in winter (Bolin 1950); these waves continue in similar positions into the spring season. Because the southward meanders over North America and Europe match regions of Northern Hemisphere mid-latitude glaciation, they obviously have the potential to be relevant to the glaciation process. A weaker southward meander of the planetary waves occurs over eastern North America in summer and continues into autumn.

There has long been interest in what factors control the planetary long-wave structure. One view is that the waves are largely created by thermal contrasts at the earth's surface (Sutcliffe 1951; Smagorinsky 1953). Differential heating of land and sea surfaces sets up, via different vertical energy exchanges, longitudinal contrasts in the distribution of upper-tropospheric mass and pressure. Relative warmth over the heated land surface in summer and over the heat-

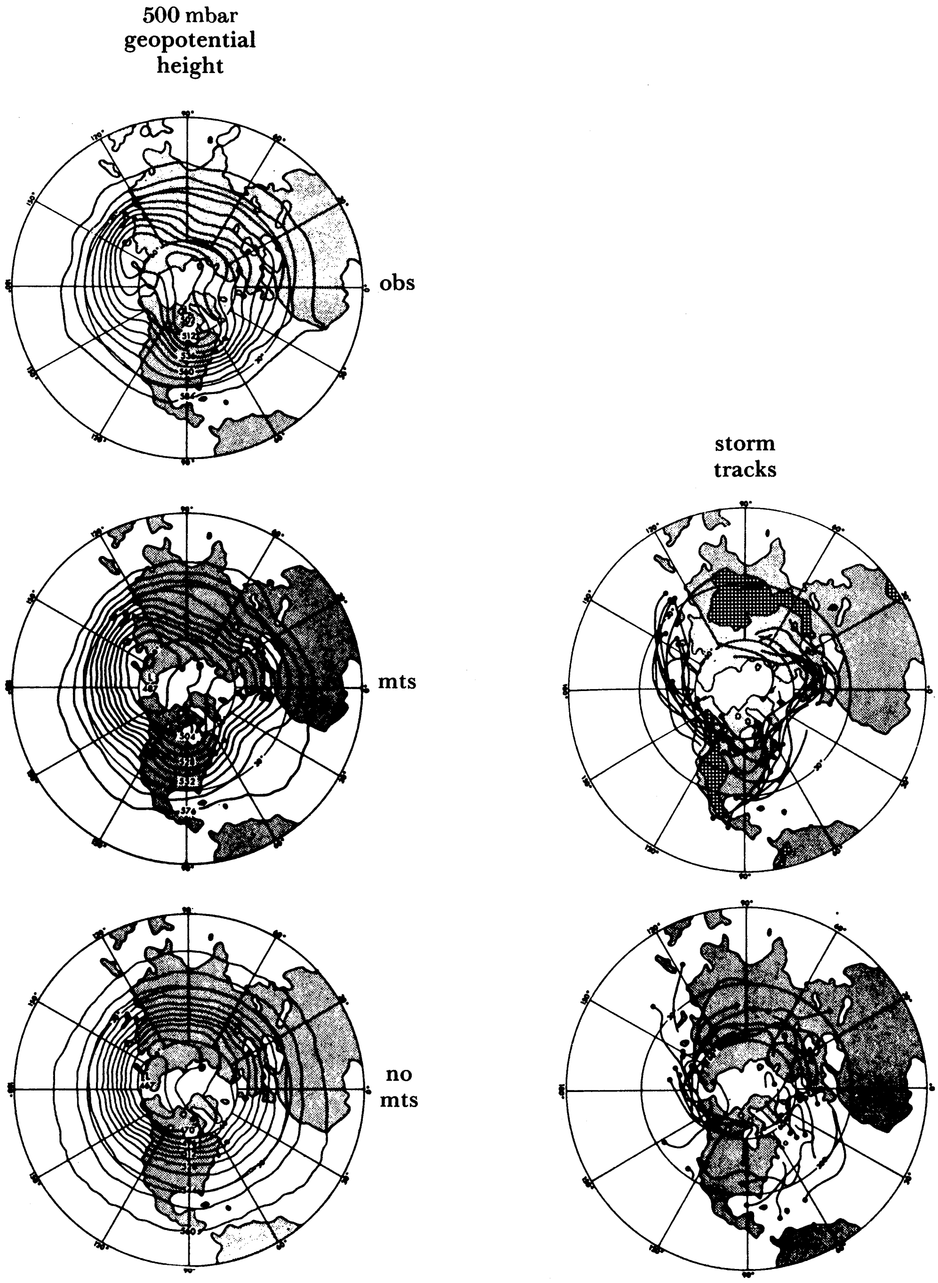

FIGURE 7. (*a*) Geopotential height (in decametres) at 500 mb for modern circulation and for 'mountain' and 'no-mountain' experiments; (*b*) storm tracks for 'mountain' and 'no-mountain' experiments (from Manabe & Terpstra 1974).

retaining ocean in winter promotes strong vertical fluxes from the lower to the upper troposphere compared with the weaker fluxes from the cooler summer ocean and colder winter land surface. These longitudinal contrasts then perturb the otherwise zonal upper westerlies into long-wave meanders.

The other view holds that orography controls the planetary wave meanders (Charney & Eliassen 1949; Bolin 1950). In some areas, the wave meanders tend to maintain similar longitudinal positions in both winter and summer, even though the land–sea temperature difference reverses sign between the seasons. This suggests that orography, which is constant between the seasons, is a stronger control on the waves than thermal contrasts. From a dynamical analysis of two-dimensional models, Bolin proposed that only two regions in the Northern Hemisphere are capable of controlling the planetary waves: mountains in the interior of Asia (Tibetan Plateau and Himalayas) and mountains in western North America (primarily the Rockies).

Control of the planetary waves is probably a combination of orographic and thermal factors (Trenberth 1983). Whereas thermal contrasts at the earth's surface are important determinants of many surface features, such as sea-level pressure, orography increasingly determines the distribution of pressure and mass (and thus the planetary waves) with altitude. There is also basic agreement that winter circulation is relatively dominated by dynamics and thus orography, whereas in summer the thermal control becomes stronger.

Bolin's analysis indicated that orography is particularly effective in positioning the planetary wave meanders over North America and the North Atlantic. In both the winter and summer seasons, the axis of the middle-tropospheric westerlies intercepts the long, linear Rocky Mountain chain and forms a downstream meander, despite shifting 10° in latitude. Over Asia, the orographic effect is strongest in winter when the westerly axis directly intercepts the Tibetan Plateau at 30°–40° N. In summer, the axis retreats far northward to latitudes where the mountains are lower and have less influence on the wave structure. Thus, thermal effects presumably are larger in Asia in summer.

The strongest confirmation of the orographic influence on the planetary waves in winter comes from general circulation model (GCM) modelling experiments by Manabe & Terpstra (1974). In one model run, present-day mountains were included in the smoothed form required by model grid-point constraints (figure 8); in the other run, mountains were omitted. The difference between the two runs showed that removing the mountains almost completely suppressed winter planetary waves in the Northern Hemisphere (figure 7). In contrast, GCM experiments by Kasahara & Washington (1971) indicated a smaller role for orography and a larger role for thermal effects.

In summary, orography appears to be the major factor determining the mean location and amplitude of the winter planetary waves. Land–sea thermal contrasts at the earth's surface are a secondary factor in winter but may be primary in summer.

Proposed link between planetary waves and glaciation

The prevailing climatological view since Hays *et al.* (1976) decisively confirmed the orbital-climate theory of Milankovitch (1941) is that the most critical factor favouring accumulation of ice sheets is decreased ablation over land, particularly in summer. This allows larger fractions of winter snowfall to survive as permanent ice. The present configuration of the planetary waves (figure 7) favours diminished ablation over North America in both seasons (Bolin 1950) because the southward meanders of the upper-level waves mark regions of increased outbreaks of lower-tropospheric polar air masses that chill the continental interiors. The wave configuration thus promotes glaciation via reduced ablation.

Other, partly conflicting, views suggest that the ocean is also critical to glaciation. One view holds that increased (or at least undiminished) snowfall in the cold season is essential to glaciation, particularly during intervals of rapid ice growth indicated by $\delta^{18}O$ data (Ruddiman & McIntyre 1981). This view calls on a warm, or at least ice-free, winter ocean as a moisture source during ice growth.

More recently, it has been suggested that the ocean may be critical to ablation, by

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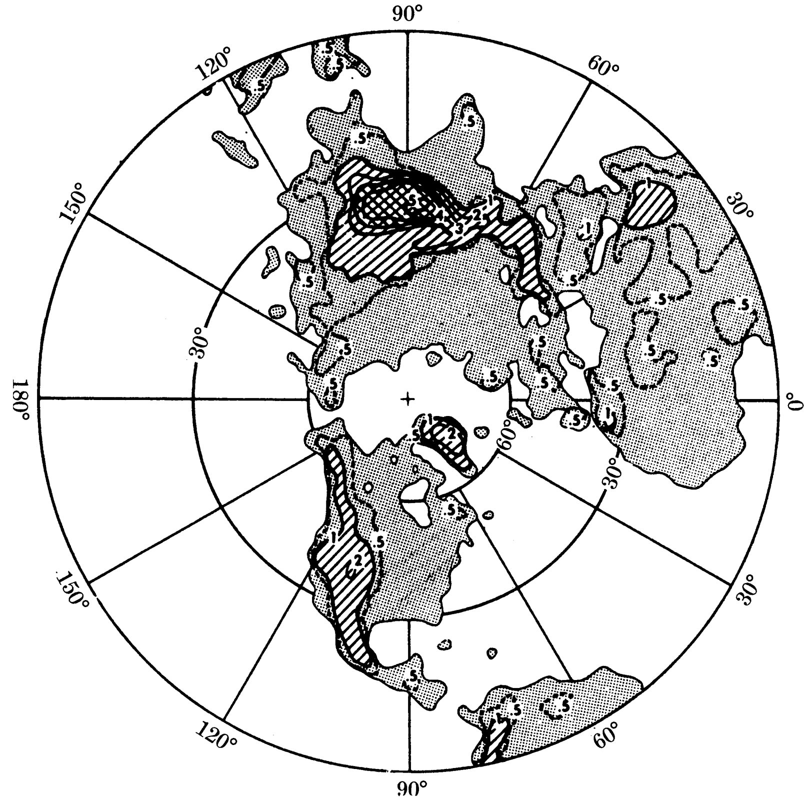

FIGURE 8. Topography entered into 'mountain' experiment of Manabe & Terpstra (1974); mountains of Tibe. and western U.S.A. are main features in spectrally smoothed representation. Elevations are contoured in kilometres. No-mountain' experiment dropped all topography to sea level.

moderating the radiative summer heating of the continental interiors (North *et al.* 1983). In this view, a cooler summer ocean promotes glaciation.

At present, it is not possible to say which of these two effects dominates long-term climatic changes. The northward meander of the modern planetary waves over the North Atlantic (figure 7) is linked to prevailing southwesterly winds that drive warm, ice-free waters far to the north; this configuration promotes glaciation via the cold-season moisture flux but opposes it via the warm-season heat flux.

In any case, diminished ablation over the continental interiors, owing to colder temperatures, is widely considered the key factor in glaciation. Thus the modern planetary-wave pattern, with southward meanders over both eastern North America and western Europe in winter and over eastern North America in summer, favours glaciation by promoting influxes of colder air from the north.

Given that the modern waves in the upper westerlies affect the lower-tropospheric circulation in ways favourable to glaciation over North America and Europe, could the form of the planetary waves have been different from that of today during the unglaciated early Pliocene and smaller glaciations of the late Pliocene and early Pleistocene? Although this question cannot be answered without GCM modelling experiments, we suggest that the planetary-wave positions shown in figure 7 may indicate at least the direction of change from the Gauss magnetic chron (before Northern Hemisphere glaciation) to the large glaciations of the Brunhes chron.

Before late-Pliocene mountain uplift, early-Pliocene cold air masses were probably more confined to the northernmost parts of North America and Europe, as in the no-mountains experiment (figure 7). Because mountains in western North America were already high-standing, presumably some kind of southwards meander of the planetary wave was present. We suggest that the degree of southward penetration of the waves over North America and Europe gradually deepened during the late Pliocene and Pleistocene, possibly changing more rapidly during the major transitions in Northern Hemisphere climatic régime at 3.15–2.40 Ma BP and at 0.9–0.45 Ma BP. The full development of the modern planetary-wave pattern may only have occurred during the 0.9–0.45 Ma BP interval of the late Pleistocene.

These changes in the winter position of the planetary waves may also have occurred for much of the late-autumn and early-spring transitional seasons, which today are characterized by the winter wave patterns. If so, the Gauss-to-Matuyama-to-Brunhes trend in circulation should be one of decreasing air temperatures and decreasing ablation of snow and ice through the winter and large parts of the transitional seasons.

The modern southward wave over eastern North America is dynamically linked to the region of increased cyclogenesis in the Atlantic off the coast of North America, as indicated by the model results (figure 7) of Manabe & Terpstra (1974). Gradual deepening of this wave through the late Pliocene and Pleistocene would have increased the presence of cyclonic storms moving to the north and northeast along the east coast of North America. One likely consequence would have been increased precipitation in maritime regions of eastern Canada, although the extent of inland penetration of these storms was probably limited.

Finally, we add the caveat that the planetary-wave theory is mainly applicable when ice sheets are relatively small in size, particularly in the early parts of ice-growth phases. When ice sheets become large, their own orography has a large effect on the planetary-wave structure (Manabe & Broccoli 1985) that persists until the ice again shrinks under orbital forcing. The proposed contribution of mountain orography is to permit faster growth (or slower decay) during intervals when ice sheets are relatively small.

Discussion

The tectonic–climate connections proposed both by Birchfield & Weertman (1983) and Ruddiman *et al.* (1986*b*) have the appeal of calling on tectonic changes that are (1) capable of affecting large-scale climate and (2) known to be occurring during the times of largest climatic change. Both hypotheses, however, have key features that need testing by climate-modelling experiments, as outlined below.

Both the albedo and orographic theories have in common one potentially major weakness. There is wide acceptance of the view of Milankovitch (1941) that glacial variations at the orbital periods are most sensitive to variations in summer insolation. Both of the above theories, however, have their largest effect on climate during the winter and transitional seasons, when orographic control of the waves is strongest. Hartman & Short (1979) have argued that the climatic effects of planetary-wave asymmetries should be greatest during summer and the transitional seasons.

Although summer insolation is the critical control for climatic changes at orbital timescales, it is not necessarily the only important ablation season for climatic change. Today, over North America, ablation does not only occur in summer; spring is a particularly important time for snowmelt, although this happens at lower rates than in midsummer. Reducing spring (and autumn) ablation must make some annual contribution to ice-sheet mass balance, and shortening the length of the high-ablation season by expanding the interval of winter-like planetary waves would further diminish annual ablation.

In addition, there is a small southward meander of the planetary waves over east-central North America even in summer, suggesting that orography continues to play some role in guiding planetary waves in that season. There may thus be a link between orography, planetary waves and ablation even in summer, although probably a weaker one than in other seasons. Ultimately, it will require GCM experiments to test whether integrated annual ablation over the glaciated regions is sufficiently affected by tectonic changes to enhance preservation of snow and ice.

A second issue is specific to the hypothesis of tectonic alteration of the planetary waves (Ruddiman *et al.* 1986*b*), which relies on the Manabe & Terpstra (1974) model results (figure 7). That experiment eliminated all mountains, not just those in regions of major Plio-Pleistocene uplift. It thus remains an open question whether observed uplift of the Tibetan plateau, the Himalayas, the Sierras, and the southern Colorado Plateau was sufficient to affect the planetary waves, and to do so in the specific way discussed here. An especially critical question is whether the relatively modest net Plio-Pleistocene uplift in portions of the western U.S.A. would be sufficient to affect the downstream wave over east-central North America. Alternatively, did the much larger but geographically more remote Tibetan–Himalayan uplift amplify wave trends over eastern North America by reinforcing the orographic effects of the already elevated Rockies? The location of the Himalayas and Tibet almost directly opposite the Rockies and Colorado Plateau (figure 8) might be particularly conducive to amplification of meanders on the other side of the world (planetary wave 2 and possibly wave 3). GCM modelling experiments to test these ideas are under way.

Critical to both theories is the need for geological evidence providing more detail about the timing of uplift during the past 3 Ma. Is there evidence that Tibetan–Himalayan and southwest U.S.A. uplift rates were greatly accelerated during the times of observed transition between major climatic régimes (3.15–2.4 and 0.3–0.45 Ma BP)? Or was uplift gradual during the entire late Pliocene and Pleistocene? In the latter case, it may be that key elevation thresholds were reached at which orographic control of the planetary waves suddenly became effective. This idea could be tested by sensitivity tests with a range of heights for the Tibetan Plateau.

One possible line of inquiry on the history of Tibetan–Himalayan uplift involves the Siberian high-pressure cell. Manabe & Terpstra (1974) note that, in their model, the wintertime strength of the Siberian High is markedly increased by the presence of the high-elevation Tibetan and Himalayan topography. Sancetta & Silvestri (1986) noted that at 2.4 Ma BP a diatom flora, interpreted to represent the first strong influence of cold dry winds from Siberian high-pressure cells, appeared in North Pacific sediments. Percentage variations in these species then intensified during the middle Pleistocene (1.0–0.35 Ma BP). These changes may be linked to intervals of rapid (or threshold-crossing) uplift in Tibet and the Himalayas. Similarly, increases in grain size and mass influx rates of aeolian material at 2.4 Ma BP in the western North Pacific are thought to reflect, in part, increased aridification of Asia (Janecek 1985). This, too, may reflect Tibetan–Himalayan uplift and isolation of Asia from moist subtropical air masses. Although these changes could of course simply be manifestations of the appearance

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and growth of Northern Hemisphere ice sheets, and their subsequent impacts on the climate system, they could also be indications that tectonic uplift altered climate.

Finally, it would be helpful to have more data on the transitions between the main Northern Hemisphere climatic régimes. Although it is quite clear that the Brunhes, Matuyama, and Gauss chrons all had distinctly different climatic responses, the sequence of change during the transitions between régimes is not yet well defined.

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References

Backman, J. 1979 Pliocene biostratigraphy of DSDP Sites 111 and 116 from the North Atlantic Ocean and the age of the Northern Hemisphere Glaciation. *Stockh. Contr. Geol.* **32**, 115–137.

Backman, J. & Pestiaux, P. 1986 Pliocene Discoaster abundance variations at Deep Sea Drilling Project Site 606: Biochronology and Paleoenvironmental implications. In *Init. Repts Deep-Sea Drilling Project*, vol. 94 (Ed. W. F. Ruddiman, R. B. Kidd, E. Thomas *et al.*), pp. 903–910. Washington, D.C.: U.S. Government Printing Offic

Barron, E. J. 1981 Paleogeography as a climatic forcing factor. *Geol. Rdsch.* **70**, 737–747.

Berger, A. L. 1984 Accuracy and frequency stability of the earth's orbital elements during the Quaternary. In *Milankovitch and climate* (ed. A. L. Berger, J. Imbrie, J. Hays, G. Kukla & B. Saltzman), pp. 3–39. Boston: D. Reidel.

Birchfield, G. E. & Weertman, J. 1983 Topography, albedo-temperature feedback, and climate sensitivity. *Science, Wash.* **219**, 284–285.

Birchfield, G. E., Weertman, J. & Lunde, A. T. 1982 A model study of the role of high-latitude topography in the climatic response to orbital insolation anomalies. *J. atm. Sci.* **39**, 71–87.

Boellestorf, J. 1978 North American Pleistocene Stages reconsidered in light of probable Pliocene–Pleistocene glaciation. *Science, Wash.* **202**, 305–307.

Bolin, B. 1950 On the influence of the earth's orography on the general character of the westerlies. *Tellus* **2**, 184–195.

Burbank, D. W. & Johnson, G. D. 1983 The late Cenozoic chronologic and stratigraphic development of the Kashmir intermontane basin, Northwestern Himalaya. *Palaeogeogr. Palaeoclim. Palaeoecol.* **43**, 205–235.

Charney, J. G. & Eliassen, A. 1949 A numerical method for predicting the perturbations of the middle-latitude westerlies. *Tellus* **1**, 38–54.

Curry, R. R. 1966 Glaciation about 3,000,000 years ago in the Sierra Nevada. *Science* **154**, 770–771.

Einarsson, T., Hopkins, D. M. & Doell, R. R. 1967 The stratigraphy of Tjornes, northern Iceland, and the history of the Bering Land Bridge. In *The Bering Land Bridge* (ed. D. M. Hopkins), pp. 312–325. California: Stanford University Press.

Eldholm, O., Thiede, J., Taylor, E. *et al.* 1987 *Init. Repts Ocean Drilling Project*, vol. 104A. Washington, D.C.: U.S. Government Printing Office.

Emiliani, C. & Geiss, J. 1958 On glaciations and their causes. *Geol. Rdsch.* **46**, 576–601.

Emiliani, C., Gartner, S. & Lidz, B. 1972 Neogene sedimentation on the Blake Plateau and the emergence of the Central American Isthmus. *Palaeogeogr. Palaeoclim. Palaeoecol.* **11**, 1–10.

Ewing, M. & Donn, W. L. 1956 A theory of Ice Ages. *Science, Wash.* **123**, 1061–1066.

Flint, R. F. 1957 *Glacial and Pleistocene geology*. New York: John Wiley.

Hartman, D. L. & Short, D. A. 1979 On the role of zonal asymmetries in climate change. *J. atm. Sci.* **36**, 519–528.

Hays, J. D., Imbrie, J. & Shackleton, N. J. 1976 Variations in the earth's orbit: pacemaker of the Ice Ages. *Science, Wash.* **194**, 1121–1132.

Hodell, D. A., Williams, D. F. & Kennett, J. P. 1985 Late Pliocene reorganization of deep vertical water mass structure in the western South Atlantic: faunal and isotopic evidence. *Bull. geol. Soc. Am.* **96**, 459–503.

Hsu, J. 1978 On the paleobotanical evidence for continental drift and Himalayan uplift. *Paleobotany* **25**, 131–142.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

Imbrie, J. 1985 A theoretical framework for the Pleistocene Ice Ages. *J. geol. Soc. Lond.* **142**, 417–432.

Imbrie, J., Hays, J. D., Martinson, D. G., McIntyre, A., Mix, A. C., Morley, J. J., Pisias, N. G., Prell, W. L. & Shackleton, N. J. 1984 The orbital theory of Pleistocene climate: support from a revised chronology of the marine $\delta^{18}O$ record. In *Milankovitch and climate* (ed. A. L. Berger, J. Imbrie, J. Hays, G. Kukla & B. Saltzman), pp. 269–305. Boston: D. Reidel.

Izett, G. A. 1975 Late Cenozoic sedimentation and deformation in Northern Colorado and adjacent areas. *Mem. geol. Soc. Am.* **144**, 179–209.

Janecek, T. R. 1985 Eolian sedimentation in the Northwest Pacific Ocean: A preliminary examination of the data from Deep Sea Drilling Project Sites 576 and 578. In *Init. Repts Deep Sea Drilling Project*, vol. 85 (ed. G. R. Heath, L. H. Burckle *et al.*), pp. 589–603. Washington, D.C.: U.S. Government Printing Office.

Kasahara, A. & Washington, W. M. 1971 General circulation experiments with a six-layer NCAR model, including orography, cloudiness, and surface temperature calculations. *J. atm. Sci.* **28**, 657–701.

Keigwin, L. D. 1978 Pliocene closing of the Isthmus of Panama, based on biostratigraphic evidence from nearby Pacific Ocean and Caribbean Sea cores. *Geology* **6**, 630–634.

Keigwin, L. D. 1982 Isotopic paleoceanography of the Caribbean and east Pacific: role of Panama uplift in late Neogene time. *Science, Wash.* **217**, 350–353.

Keigwin, L. D. 1986 Pliocene stable-isotope record of Deep Sea Drilling Project Site 606: sequential events of ^{18}O enrichment beginning at 3.1 Ma. In *Init. Repts Deep-Sea Drilling Project*, vol. 94 (ed. W. F. Ruddiman, R. B. Kidd, E. Thomas *et al.*), pp. 911–920. Washington, D.C.: U.S. Government Printing Office.

Kennett, J. P. & Thunell, R. C. 1975 Global increase in Quaternary explosive vulcanism. *Science, Wash.* **187**, 497–503.

Liu, Dongsheng & Menglin, D. 1984 The characteristics and evolution of the paleoenvironment of China since the Late Tertiary. In *The evolution of the East Asian environment* (ed. R. O. Whyte, T.-N. Chiu, C.-K. Leung & C.-L. So), pp. 11–40. Hong Kong: University of Hong Kong Centre of Asian Studies.

Loubere, P. & Moss, K. 1986 Late Pliocene climatic change and the onset of Northern Hemisphere glaciation as recorded in the northeast Atlantic Ocean. *Bull. geol. Soc. Am.* **97**, 818–828.

Lucchiatta, I. 1979 Late Cenozoic uplift of the southwestern Colorado Plateau and adjacent lower Colorado River region. *Tectonophysics* **61**, 63–95.

Manabe, S. & Broccoli, A. J. 1985 The influence of continental ice sheets on the climate of an ice age. *J. geophys. Res.* **90**, 2167–2190.

Manabe, S. & Terpstra, T. B. 1974 The effects of mountains on the general circulation of the atmosphere as identified by numerical experiments. *J. atm. Sci.* **31**, 3–42.

McDougall, I. & Wensink, H. 1966 Paleomagnetism and geochronology of the Pliocene–Pleistocene lavas in Iceland. *Earth Planet. Sci. Lett.* **1**, 232–236.

McKee, E. D. & McKee, E. H. 1972 Pliocene uplift of the Grand Canyon region – Time of drainage adjustment. *Bull. geol. Soc. Am.* **83**, 1923–1932.

Mehta, P. K. 1980 Tectonic significance of the young mineral dates and the rates of cooling and uplift in the Himalaya. *Tectonophysics* **62**, 205–217.

Milankovitch, M. M. 1941 *Canon of insolation and the ice-age problem.* Koniglich Servische Akademie, Beograd. [English translation by the Israel Program for Scientific Translations; published by the U.S. Department of Commerce and the National Science Foundation, Washington, D.C.].

North, G. R., Mengel, J. G. & Short, D. A. 1983 Simple energy balance model resolving the seasons and continents: Application to the astronomical theory of the ice ages. *J. geophys. Res.* **88**, 6576–6586.

Opik, E. 1958 Climate and the changing sun. *Scient. Am.* **198**, 85–92.

Pisias, N. G. & Moore, T. C. Jr 1981 The evolution of Pleistocene climate: a time series approach. *Earth Planet. Sci. Lett.* **52**, 450–458.

Plass, G. N. 1956 The carbon dioxide theory of climatic change. *Tellus* **8**, 140–156.

Prell, W. L. 1982 Oxygen and carbon isotope stratigraphy for the Quaternary of Hole 502B: evidence for two modes of isotopic variability. In *Init. Repts Deep-Sea Drilling Project*, vol. 68 (ed. W. L. Prell, J. V. Gardner *et al.*), pp. 455–464. Washington, D.C.: U.S. Government Printing Office.

Prell, W. L. 1984 Covariance patterns of foraminiferal $\delta^{18}O$: an evaluation of Pliocene ice-volume changes near 3.2 million years ago. *Science, Wash.* **226**, 692–694.

Raymo, M. E., Ruddiman, W. F. & Clement, B. M. 1986 Pliocene–Pleistocene paleoceanography of the North Atlantic at Deep Sea Drilling Project Site 609. In *Init. Repts Deep-Sea Drilling Project*, vol. 94 (ed. W. F. Ruddiman, R. B. Kidd, E. Thomas *et al.*), pp. 895–902. Washington, D.C.: U.S. Government Printing Office.

Rogers, K. L., Repenning, C. A., Forester, R. M., Larson, E. E., Hall, S. A., Smith, G. R., Anderson, E. & Brown, T. J. 1985 Middle Pleistocene (Late Irvingtonian: Nebraskan) climatic changes in South-central Colorado. *Natn. geogr. Res.* **1**, 535–563.

Ruddiman, W. F. 1977 Late Quaternary deposition of ice-rafted sand in the subpolar North Atlantic (lat. 40° to 65°). *Bull. geol. Soc. Am.* **88**, 1813–1827.

Ruddiman, W. F. & McIntyre, A. 1981 Oceanic mechanisms for the amplification of the 23,000-year ice-volume cycle. *Science, Wash.* **212**, 617–627.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

Ruddiman, W. F. & McIntyre, A. 1984 Ice-age thermal response and climatic role of the surface North Atlantic ocean, 40° to 63° N. *Bull. geol. Soc. Am.* **95**, 381–396.

Ruddiman, W. F., McIntyre, A. & Shackleton, N. J. 1986*a* North Atlantic sea-surface temperatures for the last 1.1 million years. In *North Atlantic paleoceanography* (ed. C. P. Summerhayes & N. J. Shackleton), (*Geol. Soc. Spec. Publ.* 21), pp. 155–173.

Ruddiman, W. F., Raymo, M. & McIntyre, A. 1986*b* Matuyama 41,000-year cycles: North Atlantic Ocean and Northern Hemisphere Ice Sheets. *Earth planet. Sci. Lett.* **80**, 117–129.

Ruddiman, W. F., McIntyre, A. & Raymo, M. 1986*c* Paleoenvironmental results from North Atlantic Sites 607 and 609. In *Init. Repts Deep-Sea Drilling Project*, vol. 94 (ed. W. F. Ruddiman, R. B. Kidd, E. Thomas *et al.*), pp. 855–878. Washington, D.C.: U.S. Government Printing Office.

Ruddiman, W. F., Kidd, R. B., Thomas, E. *et al.* 1986 *Init. Repts Deep-Sea Drilling Project*, vol. 94. Washington, D.C.: U.S. Government Printing Office.

Saini, H. S., Waraich, R. S., Malhotra, N. K., Nagpaul, K. K. & Sharma, K. K. 1979 Cooling and uplift rates and dating of thrusts from Kinnaur Himachal, Himalaya. *Himalayan Geol.* **9**, 549–567.

Sancetta, C. D. & Silvestri, S. 1986 Pliocene–Pleistocene evolution of the North Pacific ocean-atmosphere system, interpreted from fossil diatoms. *Paleoceanography* **1**, 163–180.

Schneider, D. A. & Kent, D. V. 1986 Influence of non-dipole field on determination of Plio-Pleistocene true polar wander. *Geophys. Res. Lett.* **13**, 471–474.

Seeber, L. & Armbruster, J. G. 1983 Continental subduction along the northwest and central portions of the Himilayan arc. *Boll. geof. Teoretica Appl.* **25**, 409–425.

Shackleton, N. J. & Opdyke, N. D. 1973 Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: oxygen isotope temperatures and ice volumes on a 10^5 and 10^6 year scale. *Quat. Res.* **3**, 39–55.

Shackleton, N. J. *et al.* 1984 Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic Region. *Nature, Lond.* **307**, 620–623.

Smagorinsky, J. 1953 The dynamical influence of large-scale heat sources and sinks on the quasi-stationary mean rotations of the atmosphere. *Q. Jl r. met. Soc.* **79**, 342–366.

Srivastava, S. P., Arthur, M., Clement, B. *et al.* 1987 *Init. Repts Deep-sea Drilling Project*, vol. 105A. Washington, D.C.: U.S. Government Printing Office.

Sutcliffe, R. C. 1951 Mean upper-air contour patterns of the Northern Hemisphere – the thermal-synoptic viewpoint. *Q. Jl r. met. Soc.* **77**, 435–440.

Trenberth, K. E. 1983 Interactions between orographically and thermally forced planetary waves. *J. atm. Sci.* **40**, 1126–1153.

Weissert, H. J., McKenzie, J. A., Wright, R. C., Clark, M., Oberhansi, H. & Casey, M. 1984 Paleoclimatic record of the late Pliocene at Deep Sea Drilling Project Sites 519, 521, 522, and 523 (central South Atlantic). In *Init. Repts Deep-Sea Drilling Project*, vol. 73 (ed. K. J. Hsu, J. L. LaBrecque *et al.*), pp. 701–715. Washington, D.C.: U.S. Government Printing Office.

West, R. M. 1984 Siwalik Faunas from Nepal: paleoecological and paleoclimatic interpretations. In *The evolution of the East Asian environment* (ed. R. O. Whyte, T.-N. Chiu, C.-K. Leung & C.-L. So), pp. 724–744. Hong Kong. University of Hong Kong Centre of Asian Studies.

Winograd, I. J., Szabo, B. J., Coplen, T. B., Riggs, A. C. & Holesar, P. T. 1985 Two-million year record of deuterium depletion in Great Basin ground waters. *Science, Wash.* **227**, 519–521.

Zimmerman, H. *et al.* 1985 History of Plio-Pleistocene climate in the Northeast Atlantic, Deep Sea Drilling Project Hole 552A. In *Init. Repts Deep-Sea Drilling Project*, vol. 81 (ed. D. Roberts, D. Schnitker *et al.*), pp. 861–875. Washington, D.C.: U.S. Government Printing Office.

Discussion

H. H. Lamb (*Climatic Research Unit, University of East Anglia, Norwich, U.K.*). There is surely no difficulty in supposing that a great stream of northwesterly winds from the north side of the Laurentide ice-sheet would produce a surface ocean drift and a great spread of sea ice towards mid Atlantic, if we interpret this feature of the modelled atmosphere in terms of very great persistence of the windstream (like the cold winds off the Antarctic ice-sheet today) rather than concentrating on the occasions of greatest strength, which would doubtless break up and blow away a good deal of ice.

But I am surprised that Dr Ruddiman lays so little stress on the drift of ice and polar water in glacial times from the East Greenland Current to mid ocean, which his own studies of ice-rafted material in the ocean-bed deposits first revealed. I am similarly surprised that he makes

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no mention at all of the (admittedly weaker) supply east of Iceland from the north, directly towards the British Isles. This can be substantiated as a feature of the recent Little Ice Age from the actual reports of drift ice between the late seventeenth century and the early part of this century. In the worst year, A.D. 1695, it seems (from ice and fisheries reports) that the polar water dominated the surface of the ocean across the whole width of the Norwegian Sea, extended south to the area of the Faeroe Islands and approached Shetland.

It is hard to believe that both these branches of the East Greenland cold current were not of some importance also in the major glaciation(s). Perhaps it is that they made less mark on the situation because of an inherent variability of the longitude position in which their main southward thrust occurred. This has to do with the anchoring of the east-Canadian–west-Atlantic cold trough in the upper westerlies, east of the great ice massif, and a greater variability of downstream cold troughs in the east Atlantic and European sectors.

W. F. RUDDIMAN. There is support for at least one of the drift patterns Professor Lamb mentioned. The general circulation model results of Manabe & Broccoli (1985) indeed show that the ice sheets intensified the northeasterly winds along the east coast of Greenland. As for the question of northerly winds producing a stronger southwards drift toward the British Isles east of Iceland, the evidence is more equivocal. Deposition patterns of ice-rafted sand published by Ruddiman (1977) suggest that Scandinavian icebergs did not move southwards along the European coast during glaciations; a depositional minimum located northwest of Britain argues against such a path. On the other hand, stronger north–south drift is more likely further out from the coast in the central North Atlantic, as well as in the southern Norwegian Sea near Iceland.

J. T. ANDREWS (*Department of Geological Sciences, University of Colorado, U.S.A.*). The suggested strong winds that are shown in various climatic models as moving along the northern perimeter of the ice sheet and then moving SE through Baffin Bay are a consequence of the height of the ice sheet. I still wonder whether strong, persistent offshore winds would not cause upwelling and open-pack conditions in northernmost Baffin Bay? If the wind régime were as predicted I would expect to find aeolian deposits and ventifacted surfaces in ice-free areas of Baffin Island (cf. Loken 1966) but such have not been reported.

Reference

Loken, O. H. 1966 Baffin Island refugia older than 54000 years. *Science, Wash.* **153**, 1378–1380.

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