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THE CONTRIBUTION OF ORBITAL FORCING TO THE PROGRESSIVE INTENSIFICATION OF NORTHERN HEMISPHERE GLACIATION

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Abstract—In this study, we reconstruct the timing of the onset of Northern Hemisphere glaciation. This began in the late Miocene with a significant build-up of ice on Southern Greenland. However, progressive intensification of glaciation did not begin until 3.5–3 Ma, when the Greenland ice sheet expanded to include Northern Greenland. Following this stage we suggest that the Eurasian Arctic and Northeast Asia were glaciated at approximately 2.74 Ma, 40 ka before the glaciation of Alaska (2.70 Ma) and about 200 ka before significant glaciation of the North East American continent (2.54 Ma). We also review the suggested causes of Northern Hemisphere glaciation. Tectonic changes, such as the uplift of the Himalayan and Tibetan Plateau, the deepening of the Bering Strait and the emergence of the Panama Isthmus, are too gradual to account entirely for the speed of Northern Hemisphere glaciation. We, therefore, postulate that tectonic changes may have brought global climate to a critical threshold, but the relatively rapid variations in the Earth's orbital parameters and thus insolation, triggered the intensification of Northern Hemisphere glaciation. This theory is supported by computer simulations, which despite the relative simplicity of the model and the approximation of some factors (e.g. using a linear carbon dioxide scenario, neglecting the geographical difference between the Pliocene and the present) suggest that it is possible to build-up Northern Hemisphere ice sheets, between 2.75 and 2.55 Ma, by varying only the insolation controlled by the orbital parameters. © 1998 Elsevier Science Ltd. All rights reserved

THEORIES ON THE CAUSES OF NORTHERN HEMISPHERE GLACIATION

Many explanations have been put forward to explain the initiation of Northern Hemisphere glaciation. One group of theories suggests changes in atmospheric composition or a change in total solar radiation. Theories involving changes in solar radiation are not testable (Opik, 1959), whereas changes in atmospheric CO_2 content could be detected in the geological record (Sarnthein and Fenner, 1988). Increased volcanism during the latest Cenozoic (Kennett and Thunell, 1975) has also been suggested as a possible cause of glaciation. It is now believed by some that the onset of glaciation may have caused the observed increase in Northern Hemisphere volcanism (e.g. Rea *et al.*, 1995). Other theories include virtual polar wandering (Ewing and Donn, 1956; Schneider and Kent, 1986); uplift of the high-lands of northern Canada (Flint, 1957; Emiliani and Geiss, 1958; Birchfield *et al.*, 1982); and changes in land–sea distribution by sea floor spreading (North *et al.*, 1983). These theories are either too negligible in effect or too long-term to have caused the sudden initiation of Northern Hemisphere glaciation.

Tectonic explanations have also been suggested (Hay, 1992; Raymo, 1994a), such as the emergence of the Panama Isthmus (Keigwin, 1978, 1982; Keller *et al.*, 1989; Mann and Corrigan, 1990) and the deepening of the Bering Straits (Einarsson *et al.*, 1967) and/or the Greenland–Scotland ridge (Wright and Miller, 1996) (Fig. 1). A recent dating of the closure of the Pacific–Caribbean gateway (Keller *et al.*, 1989) suggests that the Panama Isthmus began to emerge gradually at 6.2 Ma and finally closed at 1.8 Ma. Keller *et al.* (1989) also documented four major events in the progressive closure of the Pacific–Caribbean gateway dated at 6.2 Ma, 4.2 Ma, 2.4 Ma (2.55 Ma with new time scale of Shackleton *et al.*, 1995) and 1.8 Ma respectively. Keller *et al.* (1989) showed that there was an increasing abundance of salinity-tolerant planktonic foraminifera in the Caribbean from 2.55 Ma onwards, suggesting that the restriction of water flow between the Pacific and Caribbean started at 2.55 Ma and finally ceased at 1.8 Ma. Subsequent work by McDougall (1996), Collins (1996) and Geary *et al.* (1996) seem however to disagree with Keller *et al.* (1989) timing of the emergence of the Panama Isthmus. For example, McDougall (1996) study of the relative abundances of benthic foraminifera suggests the major changes due to the emergence of the Panama Isthmus occurred at 6.7–6.2 Ma, 3.4 Ma, 2.0 Ma and 1.1 Ma in the Caribbean and 6.7–6.4 Ma, 4.0–3.2 Ma, 2.1 Ma, 1.4 Ma and 0.7 Ma in Pacific. Detailed results from recent ODP Legs, particularly Leg 165, are eagerly awaited,

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as they will provide new evidence for the timing of the closing of the Panama gateway and its affect on the palaeoceanography of the Caribbean, Pacific and the Atlantic. The very latest work by Hang and Tredemann (in press) suggest the closure began at 4.6 Ma and continued until 2 Ma.

None of the present dating suggests that any key event in the closure of the Panama gateway was coincident with the timing of the intensification of North Hemisphere glaciation. Keller *et al.* (1989) suggested progressive and gradual closure from 2.55 Ma onwards is too late to have been an initiating cause. There is also a debate whether the closure of the Panama gateway would have helped or hindered the intensification of North Hemisphere glaciation. The reduced inflow of Pacific surface water to the Caribbean increased the salinity of the Caribbean. This would have both increased the salinity and strength of the Gulf Stream, thus enhancing deep water formation (Mikolajewicz *et al.*, 1993). Increased deep-water formation could have worked against the initiation of Northern Hemisphere glaciation as it enhances the heat transport of heat to the high latitudes and would have tended to prevent ice sheet formation. A contrasting argument is that the enhanced Gulf Stream could have pumped moisture north, stimulating the formation of ice sheets (Mikolajewicz *et al.*, 1993). If the closure of the Panama gateway did increase the strength of the Gulf Stream, this in turn should have increased deep water ventilation. However, the benthic foraminifera $\delta^{13}C$ records from Site 552, 607, 610, 659, 704, 846 (Tiedemann, 1991; Raymo *et al.*, 1992, 1996; Dwyer *et al.*, 1995; Shackleton *et al.*, 1995), all show a long-term decrease in $\delta^{13}C$ between 3.5 Ma and 2.0 Ma which is attributed to increased suppression of NADW formation. The records also do not contain a step-like increase at 2.55 Ma, which might have been expected to occur with an abrupt closure of the Panama gateway. Therefore, at present, the evidence seems ambiguous as to whether or not the formation of the Panama Isthmus caused or even enhanced the glaciation of the Northern Hemisphere.

The timing of tectonic changes in the Bering Sea was initially dated at between 3.5 Ma and 3 Ma (Einarsson *et al.*, 1967), and, more recently, the submergence of the Bering Strait has been dated at 3.2 Ma (Fyles *et al.*, 1991), which is too early to have caused the dramatic changes near 2.7 Ma, but they may indeed have contributed to the long-term global cooling that started at about 3.2 Ma. More recently it has been suggested that changes in the Greenland–Scotland Ridge bathymetry in the Neogene may have affected the production of Northern Component Water (NCW) (Wright and Miller, 1996). Wright and Miller (1996) suggest that reduction of the NCW on its own is unlikely to have caused the long-term Cenozoic climatic cooling. However, it may have triggered the development of a permanent ice sheet on Antarctica and the onset of Northern Hemisphere glaciation.

Ruddiman and Raymo (1988), Ruddiman *et al.* (1989a,b), and Ruddiman and Kutzbach (1991) discussed how the initiation of Northern Hemisphere glaciation could have been caused by progressive uplift of the Tibetan–Himalayan and Sierran–Coloradan regions. They suggested that this uplift altered the circulation of atmospheric planetary waves such that summer ablation was decreased, which allowed snow and ice to build-up in the Northern Hemisphere. Discussion is on-going as to whether orography (Charney and Eliassen, 1949; Bolin, 1950), differential heating of land and sea surfaces (Sutcliffe, 1951; Smagorinsky, 1953), or a combination of both (Trenberth, 1983), control the structure and direction of the planetary waves. In contrast, Copeland *et al.* (1987) and Molnar and England (1990) have suggested that the majority of the Himalayan uplift occurred much earlier between 20 Ma and 17 Ma; while Harrison *et al.* (1992), from thermochronologic results, have suggested the Tibetan Plateau reached its maximum elevation during the late Miocene at about 7–8 Ma and not during the mid-Pliocene as suggested by palaeobotanical studies (Mercier *et al.*, 1987). Quade *et al.* (1987) also showed, from carbon and oxygen isotopes of a pedogenic sediment profile in Pakistan, that the Asian monsoon underwent a strong intensification at 7.4 Ma, which they present as further evidence of a late Miocene uplift, thus suggesting the Ruddiman and Raymo (1988) hypothesis may be invalid.

Raymo *et al.* (1988), Raymo (1991, 1994b), and Raymo and Ruddiman (1992) have taken the Himalayan debate one stage further suggesting that the uplift caused a massive increase in tectonically driven chemical weathering in the late Cenozoic. They argue that carbonation of rainwater removes CO_2 from the atmosphere and forms a weak solution of carbonic acid. Dissociated H^+ ions in the acidified rainwater by hydrolysis chemically weathers the rocks. However, only weathering of silicate minerals makes a difference to atmospheric CO_2 levels, as weathering of carbonate rocks by carbonic acid returns CO_2 to the atmosphere. Bi-products of the hydrolysis reaction affecting silicate minerals are bicarbonate (HCO_3^-) anions and calcium cations. These, when washed into the oceans, become metabolised by marine plankton and are converted to calcium carbonate. The calcite skeletal remains of the marine plants and animals are ultimately deposited as deep-sea sediments and hence lost from the global biogeochemical carbon cycle for the duration of the life cycle of the oceanic crust on which they were deposited (i.e. at least 30–40 Ma). Consequently, atmospheric CO_2 is depleted causing a cooling of the global climate and thus the onset of Northern Hemisphere Glaciation.

This theory suffers from one major problem, there is no negative feedback mechanism to prevent a complete depletion of the relatively small atmospheric CO_2 reservoir (e.g. Berner, 1994; Compton and Mallinson, 1996; Raymo, in press). There have been a number of proposals which have been suggested to deal with this problem. There are three main possibilities: the geological evidence for increased chemical weathering over the last 40 Ma may have been misinterpreted

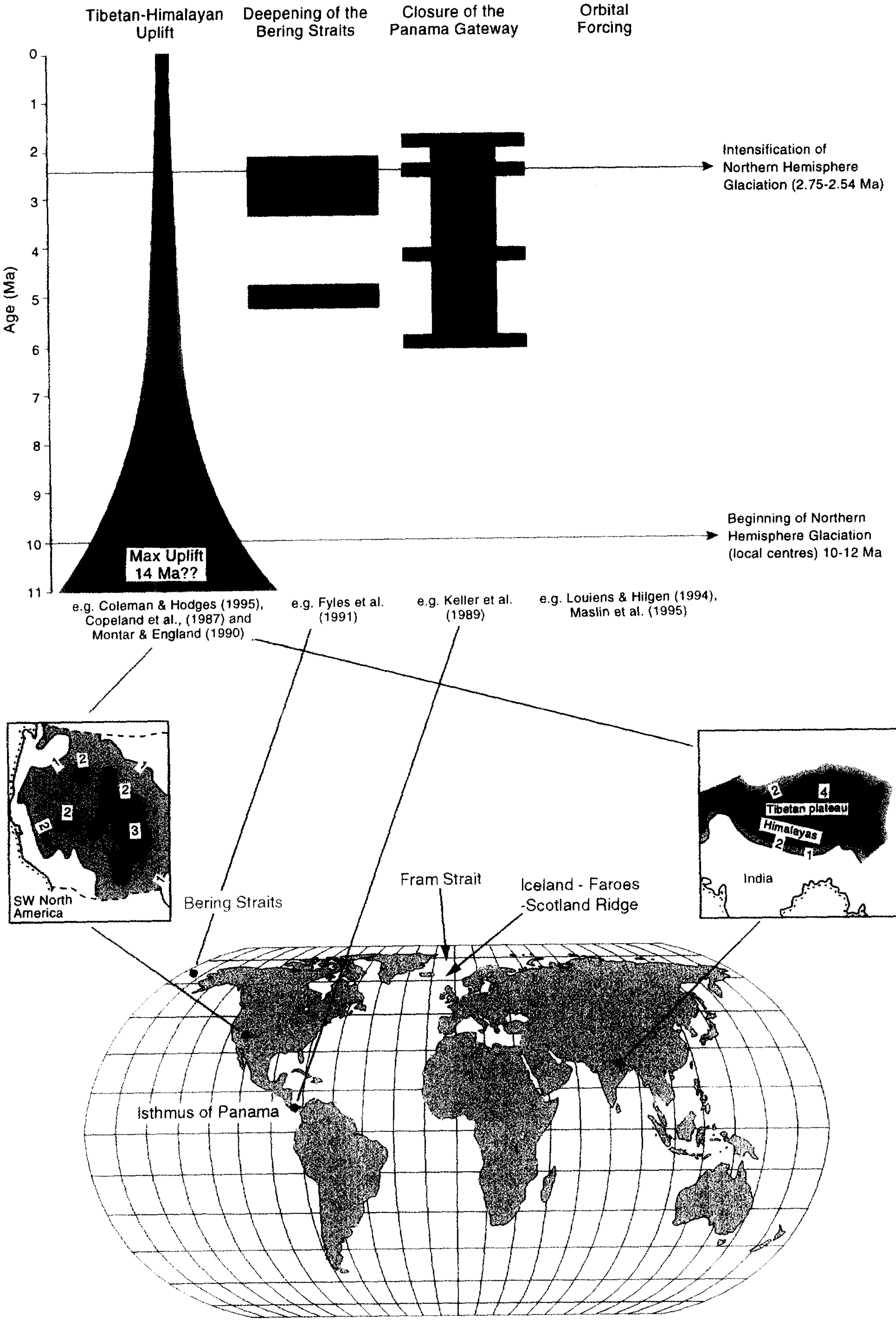


Fig. 1. Summary cartoon of the suggested causes of Northern Hemisphere Glaciation and map of the location of the major tectonic changes associated with the onset of Northern Hemisphere Glaciation.

(Kump and Arthur, in press), the assumption of constant mantle CO_2 input over the Neogene is incorrect, or another significant flux of C into the atmosphere, which is sensitive to global climate or ocean/atmosphere CO_2 levels must exist (Raymo, in press). Unfortunately, as Raymo (in press) very succinctly suggests, we do not at present have the geological data to fully understand global carbon cycles in the Cenozoic, thus our present geochemical mass balance models are only reflecting our ignorance.

In general it has been accepted that the Raymo model is a good candidate for the cause of long-term cooling of the late Cenozoic and therefore a precursor to the onset of Northern Hemisphere Glaciation. As the dates for the Himalayan uplift lie between 21 and 8 Ma, the model cannot explain the sudden step-like nature of the intensification of Northern Hemisphere Glaciation at about 2.7–2.5 Ma, unless we invoke a long-term nonlinear amplification.

A recent suggestion is that changes in orbital forcing may have been an important mechanism contributing to the gradual global cooling and the subsequent rapid intensification of Northern Hemisphere glaciation (Lourens and Hilgen, 1994; Maslin *et al.*, 1995). This theory expands on the ideas of Berger *et al.* (1993) of characterising different time periods, in the Pleistocene and late Pliocene, by the relative strength of the different orbital parameters identified. Maslin *et al.* (1995) suggested that the observed increase in the amplitude of orbital obliquity cycles, from 3.2 Ma onwards, may have increased the seasonality of the Northern Hemisphere, thus initiating the long-term global cooling trend. The subsequent sharp rise in the amplitude of precession and thus insolation between 2.8 Ma and 2.55 Ma may have forced the rapid glaciation of the Northern Hemisphere. Essentially, however, the causes of the long-term global cooling trend which started at 3.2 Ma and the significant increase in Northern Hemisphere glaciation near 2.75 Ma, are still unresolved.

REVIEW OF THE NEW EVIDENCE FROM THE NORTH PACIFIC

Ocean Drilling Program (ODP) Site 882 in the North West Pacific Ocean is located at the northern end of the Emperor Seamount Chain, on the western flank of the Detroit Seamount (50°22′N, 167 36′E) in a water depth of 3244 m (Rea *et al.*, 1993; ODP Leg 145, 1993) (Fig. 2). Site 882 provides the first

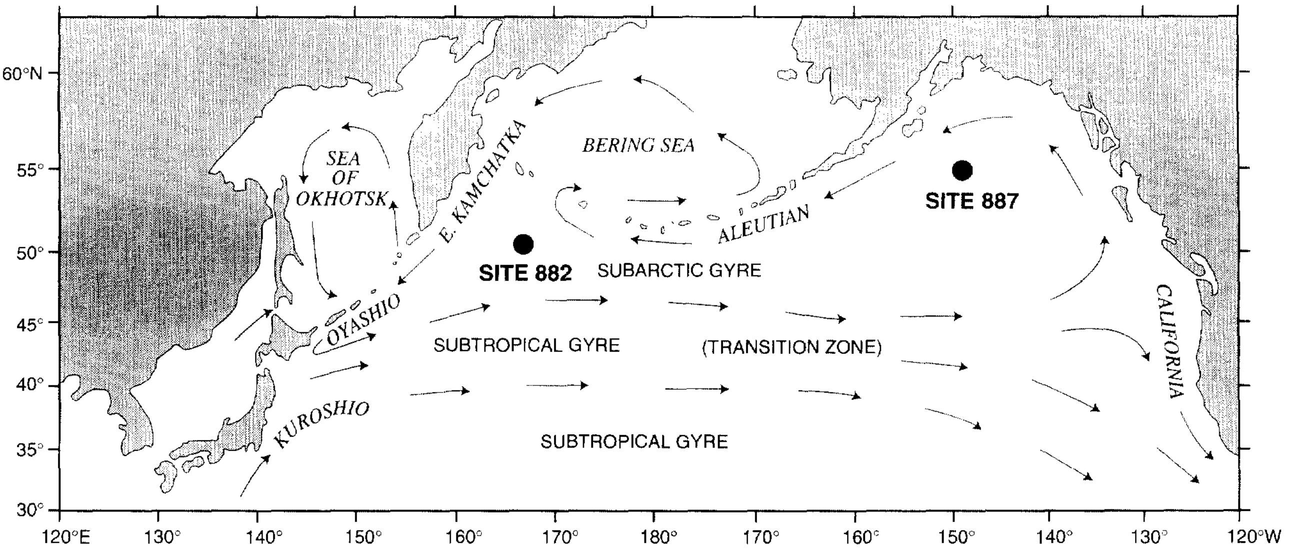
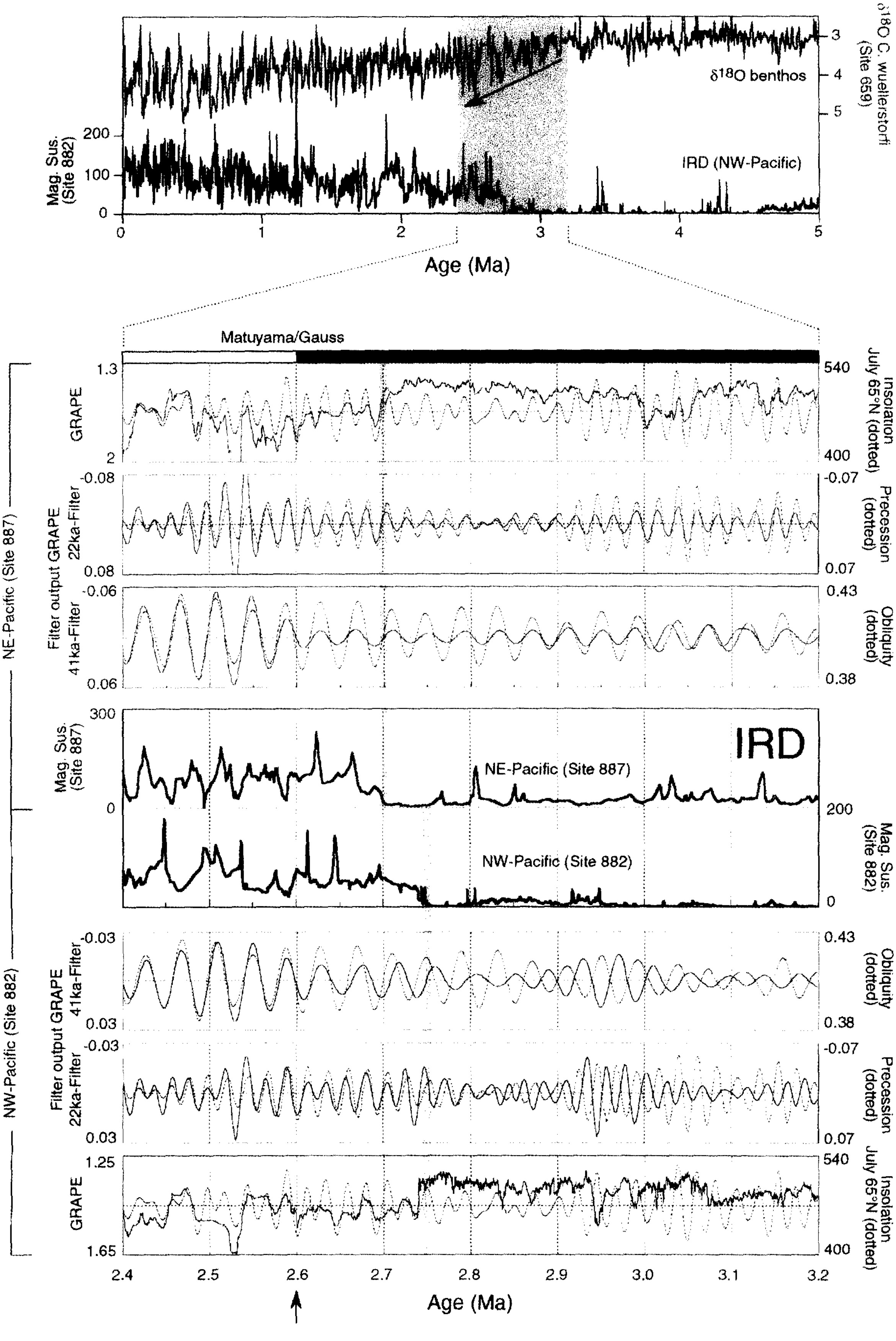


Fig. 2. Major surface water currents of the North Pacific and the location of Site 882 (50 21.79′N, 167 35.99′E) and the other ODP Leg 145 Sites.

Fig. 3. Comparison of age models of ODP Sites 882 and 887 for the time interval 3.2 Ma to 2.4 Ma with the benthic isotopes of Site 659 (Tiedemann *et al.*, 1994). Based on the Matuyama/Gauss magnetic reversal boundary at 2.6 Ma for initial age control, the GRAPE density record was fine-tuned to precession-related oscillations of the summer (July) insolation record for 65 N, assuming no phase differences. After tuning, the precession (22-ka filter output) and obliquity (41-ka filter output) components were isolated from the GRAPE-density records and compared with orbital precession and obliquity records to test the tuning strategy. The magnetic susceptibility records indicate the input of ice-rafted debris (IRD) see text. Based on the finely-tuned stratigraphy, the major onset of ice rafting occurred at 2.74 Ma in the North West Pacific (Site 882), whereas the first onset of ice rafting in the North East Pacific (Site 887) occurred at 2.7 Ma, about one obliquity cycle later. Note: At Site 882, the sediment composition changes abruptly at approximately 2.74 Ma as a consequence of the major onset of IRD deposition. Prior to 2.74 Ma, a time interval with no significant IRD input, high GRAPE-density values reflect minima in opal and maxima in carbonate accumulation (Tiedemann and Haug, 1995; Haug *et al.*, 1995a and b). Benthic and planktonic isotope records indicate that GRAPE density maxima correlate with warm stages (Maslin *et al.*, 1995). Between 2.74 and 2.4 Ma, GRAPE density minima are related to IRD input minima and biogenic opal maxima (Haug *et al.*, 1995a and b). The result is a negative correlation between GRAPE density values and the summer (July) insolation for 65 N before and after 2.74 Ma. This inverse correlation was not noted in the North East Pacific at Site 887.


δ18O C. wuellerstorfi (Site 659)
3
4
5
δ18O benthos
Mag. Sus. (Site 882)
200
100
0
IRD (NW-Pacific)
0
1
2
3
4
5
Age (Ma)
Matuyama/Gauss
NE-Pacific (Site 887)
GRAPE
1.3
2
Insolation July 65°N (dotted)
540
400
Filter output GRAPE
22ka-Filter
-0.08
0.08
Precession (dotted)
-0.07
0.07
41ka-Filter
-0.06
0.06
Obliquity (dotted)
0.43
0.38
Mag. Sus. (Site 887)
300
0
IRD
NE-Pacific (Site 887)
NW-Pacific (Site 882)
Mag. Sus. (Site 882)
200
0
NW-Pacific (Site 882)
Filter output GRAPE
41ka-Filter
-0.03
0.03
Obliquity (dotted)
0.43
0.38
22ka-Filter
-0.03
0.03
Precession (dotted)
-0.07
0.07
GRAPE
1.25
1.65
Insolation July 65°N (dotted)
540
400
2.4
2.5
2.6
2.7
2.8
2.9
3.0
3.1
3.2
Age (Ma)

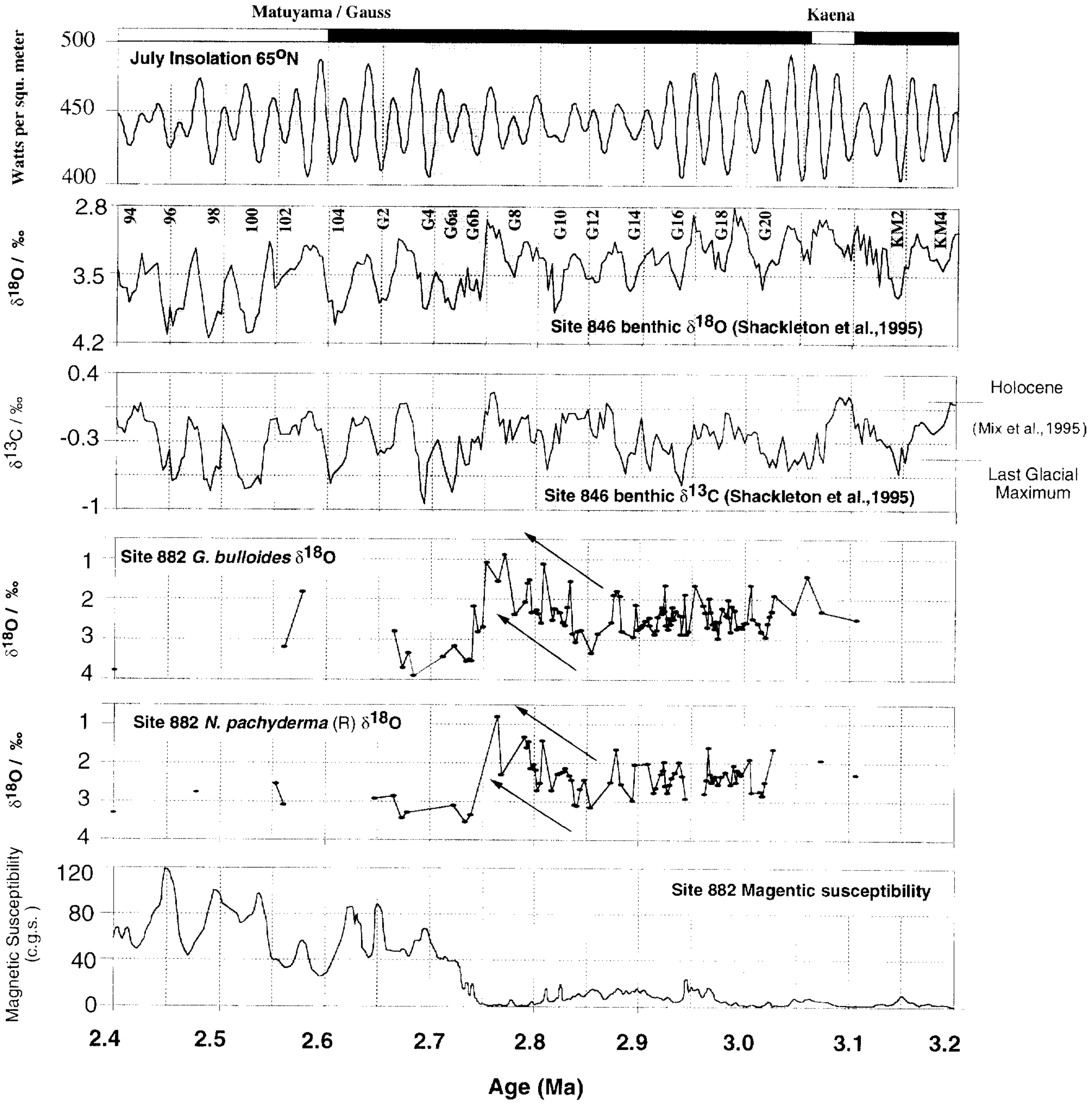


Fig. 4. Comparison between the calculated insolation for 65 N (Loutre and Berger, 1993), benthic $\delta^{18}O$ and $\delta^{13}C$ isotope records from the equatorial Pacific at Site 846 (Shackleton *et al.*, 1995), the $\delta^{18}O$ records from *N. pachyderma* (r) and *G. bulloides* and magnetic susceptibility record from Site 882 for the time interval 3.2 Ma to 2.4 Ma. The shaded regions represent global cold periods and are labelled according to the nomenclature of Shackleton *et al.* (1995) and Tiedemann *et al.* (1994).

high-resolution carbonate, opal, and foraminifera stable isotope records for the interval between 3.2 and 2.4 Ma in the North Pacific (Haug, 1995; Haug *et al.*, 1995a; Maslin *et al.*, 1995). The sedimentation rate at this site varied between 12 to 4 cm/1000 years between 3.2 Ma and 2.4 Ma, with a major drop in the sedimentation rate occurring near 2.75 Ma (Tiedemann and Haug, 1995). ODP Site 887 is also used in this study and provides the high resolution records of climate changes in the far North East Pacific. The site is located on the Patton–Murray Seamount (54 22′N, 148 27′W) in a water depth of 3645 m (Rea *et al.*, 1993; ODP Leg 145, 1993) (Fig. 2).

The time scale for Sites 882 and 887 were based initially on magnetostratigraphy. The ages for the magnetic reversal boundaries were derived from the orbitally tuned time scale of Shackleton *et al.* (1990, 1995) and Tiedemann *et al.* (1994) which modified the ages published by Hilgen (1991). Using the astronomically dated magnetic reversals for initial age control, we found that the fluctuations in the gamma ray attenuation porosity evaluator (GRAPE) density and magnetic susceptibility records were linked to variations in the Earth's orbit, as expected. GRAPE-density provides an analog data of the sediment wet-bulk density, while magnetic susceptibility at this location provides a proxy for ice rafted debris (Haug *et al.*, 1995a; Maslin *et al.*, 1995).

Tiedemann and Haug (1995) generated an astronomically calibrated stratigraphy for Site 882 for the last 4 Ma based on fine-tuning the GRAPE-density oscillations in the precession band to the summer insolation at 65°N (Berger and Loutre, 1991). This calibration was based on the assumption that GRAPE-density maxima are linked to minima in biogenic opal and maxima in ice-rafted debris, and hence to glacial stages (insolation minima) during the last 2.75 Ma, confirmed by benthic and planktonic stable isotopes for the last 750 ka (Haug, 1995; Haug *et al.*, 1995a,b). Prior to 2.75 Ma, the relationship between GRAPE-density and climate was the reverse, which is confirmed by the benthic $\delta^{18}O$ results from Site 882 (Maslin *et al.*, 1995). A detailed description of the astronomically calibrated age model is given by Tiedemann and Haug (1995). The age model for Site 887 (Fig. 3) between 2.4 Ma and 3.2 Ma was also based on the assumption that GRAPE-density maxima are linked to minima in biogenic opal and maxima in ice-rafted debris, and hence to glacial stages. There is no evidence of an inverse correlation which was found at Site 882 by Tiedemann and Haug (1995).

The $\delta^{18}O$ and $\delta^{13}C$ of benthic and planktic foraminifera species were measured according to the standard techniques at the University of Kiel (Maslin *et al.*, 1995). Three planktonic foraminifera species were analyzed, *Globigerina bulloides*, *Neogloboquadrina pachyderma* (r), and *Neogloboquadrina pachyderma* (l). *G. bulloides* and *N. pachyderma* (r) occurred with a sufficient time resolution to have near continuous records between 3.4 and 2.4 Ma, whereas *N. pachyderma* (l) occurred sporadically and was used only to confirm the pattern of the other two species. Using the Site 882 age model, it is possible to compare Site 882 planktonic $\delta^{18}O$ with the benthic *Cibicidoides wuellerstorfi* $\delta^{18}O$ and $\delta^{13}C$ records from the Equatorial Pacific Site 846 (Shackleton *et al.*, 1995). Fig. 4 indicates a high degree of similarity between the two records, the most significant of which occurs between 2.7 Ma and 2.9 Ma (Stage G14 to G4, using the nomenclature of Tiedemann *et al.*, 1994). Between 2.85 Ma and 2.75 Ma, the planktonic foraminifera $\delta^{18}O$ values show a trend towards lighter values ($> 1‰$), overlying the warm–cold period variations, the lightest values are at 2.76 Ma, indicating either a very high surface water temperature or low salinity or a combination of both. Subsequently, at 2.75 Ma, coeval with the onset of the G6 glaciation, significant ice rafting occurred in the Northwest Pacific Ocean and the planktonic foraminiferal $\delta^{18}O$ drops dramatically by 2.6‰.

If ice volume and sea level effects are taken into consideration and sea surface temperature (SST) is assumed to be the only other variable, then, as Maslin *et al.* (1996) suggests the SSTs change between cold to warm periods was at least 5°C before 2.75 Ma with an overall warming trend of 4°C and a dramatic drop of no less than 7.5°C at 2.75 Ma (using the SST-$\delta^{18}O$ equation of Shackleton (1974)). This led Sarnthein *et al.* (1995) to suggest that these very high Northwest Pacific SSTs and the possible warming trend prior to 2.75 Ma would have produced enhanced evaporation and thus a mechanism to transport additional moisture to the growing Northern Hemisphere continental ice sheets. They stressed, however, that surface water salinity changes could not be ruled out.

Biomarkers, such as the $U^{K_{37}}*$ index (Brassell *et al.*, 1986a,b) and its simplification, $U^{K'_{37}}$ (Brassell *et al.*, 1986b), have been developed as an independent estimate of sea surface temperatures. The $U^{K_{37}}*$ index is based on the relative abundances of long-chain alkenones with 37 atoms, which are ubiquitous in the World's oceans, both in the water column and in the sediment (Brassell *et al.*, 1986a; Rosell i Mele, 1994). Haug *et al.* (1995b), Haug (1995) and Haug *et al.* (*in prep.*) were able to measure the $U^{K'_{37}}$ index for the last 6 Ma at ODP Site 882. The preliminary $U^{K'_{37}}$ index results indicate that before 2.75 Ma at Site 882, there was a moderate cooling, indicating that the very light planktonic $\delta^{18}O$ was probably not due to warming, but to a strong freshening of the North Pacific, about 1–2‰ over 80 ka. The $U^{K'_{37}}$ index also indicates a warming of the surface waters at 2.75 Ma (Haug *et al.*, 1995b and Haug, 1995), though this could be due to contamination from older material brought in by the initiation of ice rafting, as indicated by the magnetic susceptibility record in Fig. 3 (Haug, 1995). Therefore at present the true cause of the extremely light planktonic foraminifera $\delta^{18}O$ prior to 2.75 Ma, and the large increase at 2.75 Ma, is unclear, but the evidence at present points to strong salinity changes in the Northwest Pacific during the mid to late Pliocene.

DISCUSSION

Timing of the initiation and intensification of Northern Hemisphere glaciation

The earliest recorded onset of significant global glaciation during the last 100 Ma was the continent-wide glaciation in Antarctica at about 34 Ma (e.g. Hambrey *et al.*, 1991; Breza and Wise, 1992; Miller *et al.*, 1991; Zachos *et al.*, 1992, 1996). In contrast the earliest recorded glaciation in the Northern Hemisphere is between 10 and 6 Ma (e.g. ODP Leg 151, 1994; Jansen *et al.*, 1990; Wolf and Thiede, 1991; Jansen and Sjøholm, 1991; Wolf-Welling *et al.*, 1995). This Miocene initiation of the Northern Hemisphere glaciation has been recorded in the Greenland Sea (Wolf and Thiede, 1991; Wolf-Welling *et al.*, 1996; Fronval and Jansen (1996); Thiede and Myhre (1996)), the Norwegian Sea (Jansen *et al.*, 1990; Jansen and Sjøholm, 1991), the Arctic (ODP Leg 151, 1994) and in the Northwest Pacific (Haug *et al.*, 1995a; Haug, 1995).

The first occurrence of significant ice sheets in the Northern Hemisphere was thought not to occur until about 2.55 Ma (Shackleton *et al.* (1984), a date which is adjusted to the new time scale of Shackleton *et al.* (1990) and Hilgen (1991)), although these were preceded by smaller increases in ice volume from 2.7 Ma onwards (Backman, 1979; Shackleton *et al.*, 1984; Zimmerman *et al.*, 1985). These conclusions were based primarily on relatively low-resolution records of %$CaCO_3$ and $\delta^{18}O$ from Deep Sea Drilling Project (DSDP) Site 552 in the Atlantic Ocean (Shackleton *et al.*, 1984). The sudden decrease in %$CaCO_3$ values was explained by an increase in ice-rafted debris coeval with a drop in $\delta^{18}O$ which indicated an increase in global ice volume (Shackleton *et al.*, 1984).

Recent evidence, though, suggests that the initiation of major Northern Hemisphere glaciation was the culmination of a longer term, high latitude cooling, which began in the late Miocene with the glaciation of Greenland and the Arctic (Wolf and Thiede, 1991; ODP Leg 151, 1994; Wolf-Welling *et al.*, 1995; Wolf-Welling *et al.*, 1996; Fronval and Jansen, 1996; Thiede and Myhre, 1996). This long-term cooling intensified at 3.2 Ma (Ruddiman *et al.*, 1986b; Tiedemann *et al.*, 1994) with the progressive and oscillatory enrichment of benthic foraminiferal $\delta^{18}O$ and suggests that there was significant deep water cooling and an increase in global ice volume after that time (Prell, 1984; Keigwin, 1986; Ruddiman *et al.*, 1986a,b; Sarnthein and Tiedemann, 1989; Tiedemann *et al.*, 1994).

There is strong evidence for this progressive cooling of the Northern Hemisphere from the increase in the percentage of cold water-dwelling planktonic foraminifera (Loubere and Moss, 1986; Raymo *et al.*, 1986) and by decreasing abundances of discoaster (Backman and Pestiaux, 1986). Sea-surface temperatures (SSTs), for this time interval, however, are not easily reconstructed, as the assemblages of planktonic foraminifera are not analogous to modern faunas (Ruddiman and Raymo, 1988). Chapman (1992, 1995), however, has demonstrated using evolutionary-equivalent species that it is possible to reconstruct Pliocene subtropical SSTs. His work on ODP Site 659 shows a major cooling of the tropical Eastern Atlantic at 2.55 Ma.

The most recent benthic foraminiferal $\delta^{18}O$ results from ODP Site 659 in the tropical East Atlantic (Tiedemann *et al.*, 1994), and Site 846 in the equatorial East Pacific (Shackleton *et al.*, 1995), show a gradual enrichment of ^{18}O in the ocean between 3.2 Ma and 2.75 Ma. From 2.75 to 2.68 Ma, there are three successive intervals with very heavy $\delta^{18}O$ values, but with minimal climatic recovery between them. Subsequently, there were still further increases in the average $\delta^{18}O$ value, and a marked increase in the amplitude of variation between warm and colder periods. The major controversy at present is how much of these $\delta^{18}O$ shifts represent cooling of the deep waters and how much represent the first build up of ice in the Northern Hemisphere. It is believed that the three major increases in benthic $\delta^{18}O$ between 2.75 and 2.68 Ma, were primarily due to a significant increase in global ice volume (Shackleton *et al.*, 1995; Tiedemann *et al.*, 1994); which were associated with an increased suppression of the formation of North Atlantic Deep Water (NADW) (Sarnthein and Tiedemann, 1989; Tiedemann, 1991; Raymo *et al.*, 1992).

To provide more detail of the intensification of Northern Hemisphere glaciation during the Pliocene, a number of ODP deep-sea sediment proxy climate records have been tuned to the orbital 'Milankovitch' variations, with a resolution of between 3–10 ka. Taking into account the criteria and the pitfalls suggested by Shackleton *et al.* (1995), the Pliocene sections from the North Pacific have been age modelled (ODP Sites 882 and 887; Tiedemann and Haug (1995)) while other key ODP records (Sites, 609, 610, 642, 644, 552) have been evaluated and adjusted to the revised astronomically calibrated time scale (Shackleton *et al.*, 1990; Hilgen, 1991; Tiedemann *et al.*, 1994; Shackleton *et al.*, 1995).

Evidence from the Norwegian Sea and Iceland suggests that the first minor increase in Arctic, Iceland and Scandinavian glaciation occurred at 3 Ma, while a more pronounced and sustained increase occurred at about 2.74 Ma (Jansen *et al.*, 1990; Jansen and Sjøholm, 1991; Einarsson and Albertsson, 1988; Geirsdottir and Eiriksson, 1994; Fronval and Jansen 1996). This increase is coeval with the increase in the magnetic susceptibility record of Site 882 which demonstrates a minor increase in ice rafting at 3 Ma (Fig. 4), and a major increase at 2.74 Ma. McKelvey *et al.* (1995) has traced the source of late Pliocene ice rafting at Site 882 to the Kamchatka peninsula and the northern coast of the Sea of Okhotsk. This suggests that the ice rafted debris observed at Site 882 was predominantly derived from the Eurasian Arctic (via the Chukchi and Bering Sea) and Northeast Asia. Evidence from the Norwegian Sea and Site 882 confirms that there was ice cover in the circum-Arctic continents from about 3 Ma, and that this significantly expanded at about 2.74 Ma.

Evidence from Northern Alaska (including pollen, plant macrofossil and marine vertebrate records) suggests that there were three major marine transgressions in the late Pliocene (Brigham-Greete and Carter, 1992; Kaufman and Brighamn-Grette, 1993). These were dated using amino acid geochemistry, palaeomagnetic studies, vertebrate and invertebrate paleontology and strontium isotopes to between 2.48 Ma and 2.7 Ma (2.62 Ma and 2.86 Ma using the new time scale), which is equivalent to the three warm periods observed in the deep-sea sediment records between Oxygen Isotope Stage 104 and G12. They also found that these periods were too warm for permafrost and even seasonal sea ice. During the waning stages of these transgressions the terrestrial conditions were cool enough to support a herbaceous tundra vegeta tion with scattered larch trees. They found no evidence for any glaciation of Northern Alaska between 2.62 Ma and 2.86 Ma. However, the magnetic susceptibility and GRAPE density records from Site 887 in the Gulf of Alaska (Figs. 1 and 3) indicates that there was a dramatic increase in the supply of ice-rafted debris at 2.70 Ma. This suggests that Southern Alaska was glaciated 40 ka later i.e. one

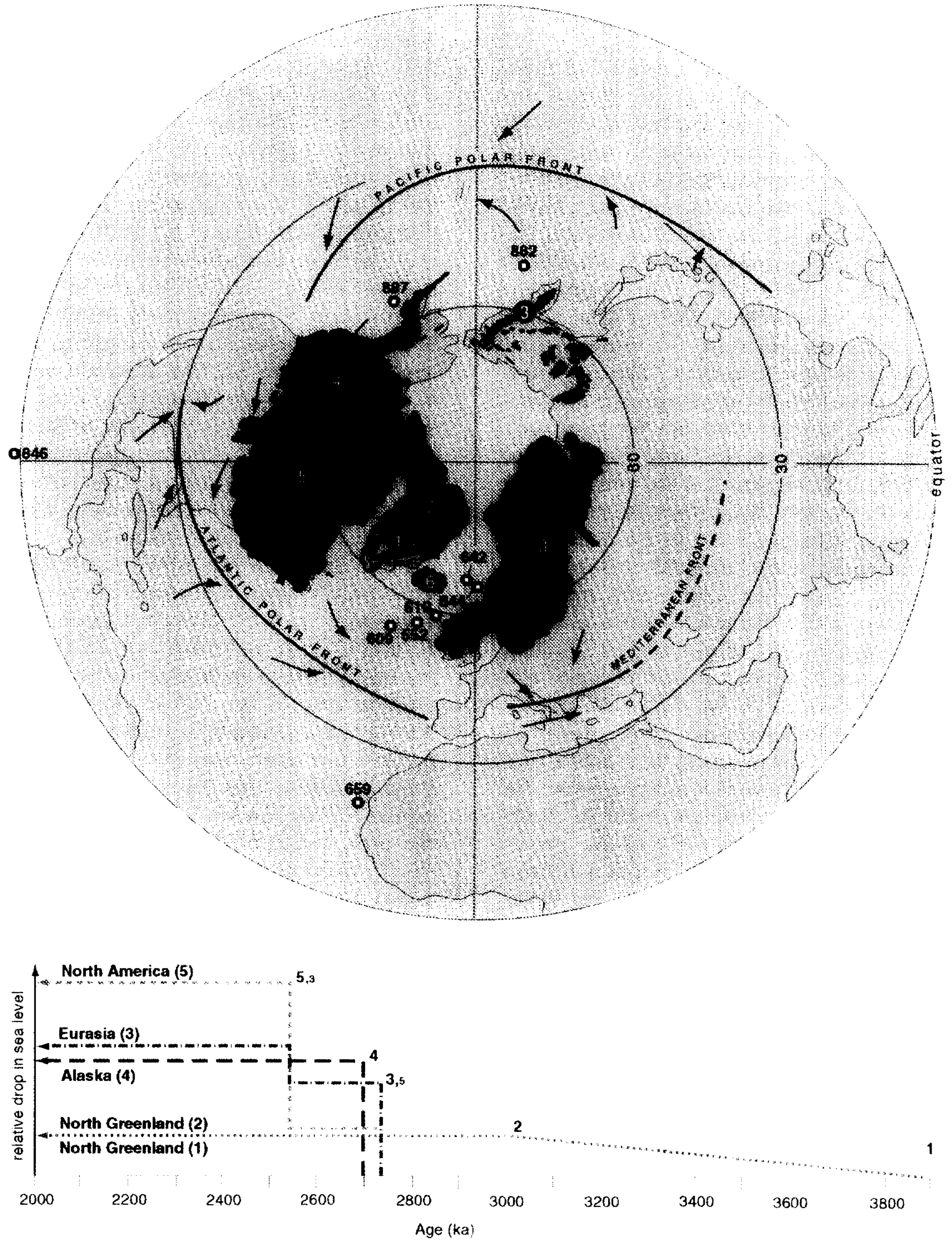


Fig. 5. Timing of the intensification of Northern Hemisphere Glaciation. Map shows the extent of the Northern Hemisphere ice sheets during the Last Glacial Maximum (CLIMAP, 1976, 1981). The separate ice sheets are identified by different colours. The **modern** climatic fronts are also shown to represent the possible atmospheric circulation of the pre-glacial Pliocene. The graph shows the estimated time at which each ice sheet was large enough to reach the edge of the continents and thus release icebergs, compared with its relative effect on the sea-level, estimated from the change in the global $\delta^{18}O$ signal (Tiedemann *et al.*, 1994; Shackleton *et al.*, 1995). Iceland seems to have had repeated glacial episodes from the mid-Pliocene onwards, but no clear dates are yet available for when this became sustained (Einarsson and Albertsson, 1988; Geirsdottir and Eiriksson, 1994). The data for the onset of ice rafting of the different ice sheets come from the following sources: Southern Greenland — coarse fraction analysis (e.g. Wolf and Thiede, 1991; ODP Leg 151, 1994; Wolf-Welling *et al.*, 1995; Wolf-Welling *et al.*, 1996; Fronval and Jansen 1996 and Thiede and Myhre 1996). Northern Greenland — coarse fraction analysis (e.g. Wolf and Thiede, 1991; ODP Leg 151, 1994; Wolf-Welling *et al.*, 1995; Wolf-Welling *et al.*, 1996; Fronval and Jansen 1996 and Thiede and Myhre 1996). Eurasian Arctic — lithic fragment counts (Jansen *et al.*, 1990; Jansen and Sjøholm, 1991) and Northeast Asia — magnetic susceptibility and coarse fraction analysis (Haug, 1995; Haug *et al.*, 1995a,b; Maslin *et al.*, 1995 and this study), Alaska and the Northwest coast of America — magnetic susceptibility (Rea *et al.*, 1993; Maslin *et al.*, 1995 and this study) and Northeast America — calcium carbonate, magnetic susceptibility, stable isotopes, ostracods and coarse fraction analysis (e.g. Shackleton *et al.*, 1984; Ruddiman and Raymo, 1988; Raymo *et al.*, 1989, 1992; Cronin *et al.*, 1996).

obliquity cycle after the Eurasian Arctic and Northeast Asia (Fig. 5).

Evidence from ODP Sites 607, 609 and 610 (Ruddiman and Raymo, 1988; Raymo *et al.*, 1989, 1992; Cronin *et al.*, 1996) suggests that the first incidence of ice-rafting in the North Atlantic occurred at about 2.74 Ma coeval with the expansion and increased production of icebergs from the Eurasian ice sheet. However evidence from these sites and DSDP Site 552 (Shackleton *et al.*, 1984) suggests that the North American ice sheets did not expand until 2.54 Ma, when there was a significant increase in the deposition of ice rafted debris as recorded in the sediments of the North Atlantic Ocean (Oxygen Isotope Stage 100). There may have been, therefore, a delay of 200 ka between the intensification of glaciation in the Arctic and Northeast Asia and the major build-up ice on the North East America continent (Fig. 5). This suggests that the initiation and intensification of Northern Hemisphere glaciation was driven by climate changes in the Arctic and North Pacific regions, opposed to the Laurentide ice sheet dominated scenario of the Last Glacial Maximum, and much of the late Pleistocene (Ruddiman and McIntyre, 1981).

Sea-level changes inferred from the benthic foraminiferal oxygen isotope records of Sites 659 (Tiedemann *et al.*, 1994) and 846 (Shackleton *et al.*, 1995; see Fig. 4), corroborate the timing described above and suggest that the two most important stages in the glaciation of the Northern Hemisphere were the maturing of the Eurasian–Northeast Asian ice sheets (Oxygen Isotope Stage 110 or G6) and the proto-Laurentide ice sheet on the eastern North American continent (Oxygen Isotope Stage 100). We stress though that the step-like nature of the ice rafting records may conceal a more gradual process of ice build-up, indicated by the progressive ^{18}O enrichment of benthic isotope records (Tiedemann *et al.*, 1994; Shackleton *et al.*, 1995). This is because the ice-rafting records indicate only when the continental ice sheets were mature enough to impinge on the adjacent oceans. Dramatic changes, however, are observed in each ocean basin when ice-rafting first occurs.

Was orbital forcing a contributing factor in the intensification of Northern Hemisphere glaciation?

According to the astronomical theory of palaeoclimates (Milankovitch, 1949; Berger, 1988), the long-term variations in the geometry of the Earth's orbit and rotation are the fundamental causes of the comings and goings of Pleistocene ice ages (Berger, 1989). It has been suggested (Berger, 1976, 1979) that it is changes in the latitudinal distribution and seasonal pattern of insolation which are the key factors driving the behaviour of the climate system. The causes of the initiation and intensification of Northern Hemisphere glaciation essentially remain unresolved. We suggest that the long-term cooling, and the dramatic intensification of Northern Hemisphere glaciation, can be in part explained by both tectonics and changes in orbital forcing between 4 Ma and 2 Ma (Figs. 1 and 6). Changes in the climate of the Pliocene are dominated by the 41 ka periodicity of orbital obliquity (Ruddiman *et al.*, 1986a; Tiedemann *et al.*, 1994; Shackleton *et al.*, 1995). This climate response to orbital forcing does not change until approximately 900–800 ka ago when the 41 ka cyclicity is progressively superseded by the 100 ka period of orbital eccentricity (Shackleton and Opdyke, 1973; Imbrie *et al.*, 1992 and 1993). Berger *et al.* (1993) has demonstrated that longer-term climate fluctuations of the last 1.8 Ma can be subdivided into three distinct response periods to orbital forcing: (1) 0–610 ka, where the eccentricity-related signal is dominant, (2) 610 ka to 1.22 Ma, where the obliquity signal is dominant and (3) 1.8 Ma to 1.22 Ma, where both signals are equally important. DeMenncal (1995) documented that African climate was dominated by a precession-related signal before 2.8 Ma, by an obliquity signal between 2.8 and 1 Ma, and by an eccentricity-related signal after 1 Ma.

Fig. 6 shows the key orbital parameters of the late Pliocene, obliquity and precession and the resultant July insolation at 65°N, as calculated by Loutre and Berger (1993). From 3.5 Ma until 2.5 Ma the amplitude of obliquity cycles gradually increased until it reach its peak between 2.5 Ma and 2.1 Ma. Between 2 Ma and 3.5 Ma, the eccentricity-modulated amplitude of the precessional signal was dominated by a period of roughly 400 ka, with the largest amplitudes centred around 2.2 Ma, 2.6 Ma, 3.05 Ma and 3.45 Ma, though the peak in amplitude at around 2.6 Ma is slightly smaller than the three other peaks. This pattern of the orbital elements resulted in the amplitude of the high latitude summer insolation cycles being dominated by an approximate 400-ka period. As July 65°N insolation is dominated by the precessional signal, four stages with large amplitudes of the signal can be identified, roughly between 3.45 Ma and 3.3 Ma, between 3.2 Ma and 2.9 Ma, between 2.75 Ma and 2.45 Ma, and between 2.35 Ma and 2.1 Ma.

For a glaciation to develop, the astronomical theory requires that summers in the northern high latitudes must be cool enough to prevent winter snows from melting. This allows there to be a positive annual budget of snow and ice, thus initiating a positive feedback via increased surface albedo. In the late Pliocene, lower summer insolation during a period of large amplitude in the insolation signal, should favour the development of Northern Hemisphere ice sheets (especially if the tectonically-induced cooling had brought the hemisphere to a critical threshold), as the high-latitude winter snow would not be fully removed by the cooler summers.

The observed rapid intensification of Northern Hemisphere glaciation between 3.2 Ma and 2.4 Ma occurred within one of these periods of high-amplitude in insolation. Tectonic forcing alone cannot explain the pronounced transition in both the intensity of glacial–interglacial cycles and mean global ice volume. We suggest that the gradual long-term tectonic forcing

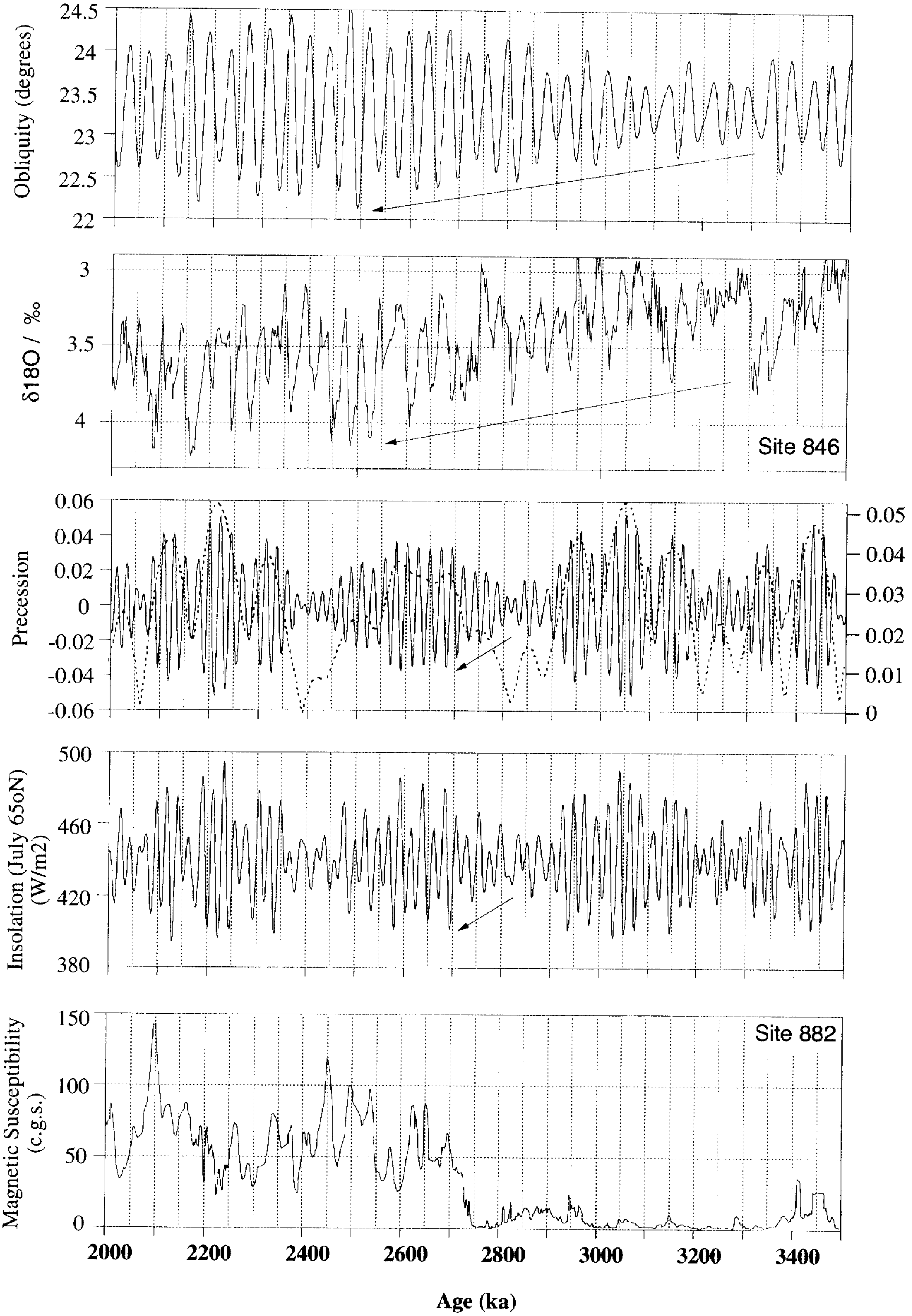


Fig. 6. Comparison between the key orbital parameters of the Pliocene, obliquity and precession and the resultant insolation at 65°N, as calculated by Loutre and Berger (1993) with the benthic $\delta^{18}O$ record from the equatorial Pacific at Site 846 (Shackleton *et al.*, 1995) and the magnetic susceptibility record from Site 882. Arrows indicate general tends in the data.

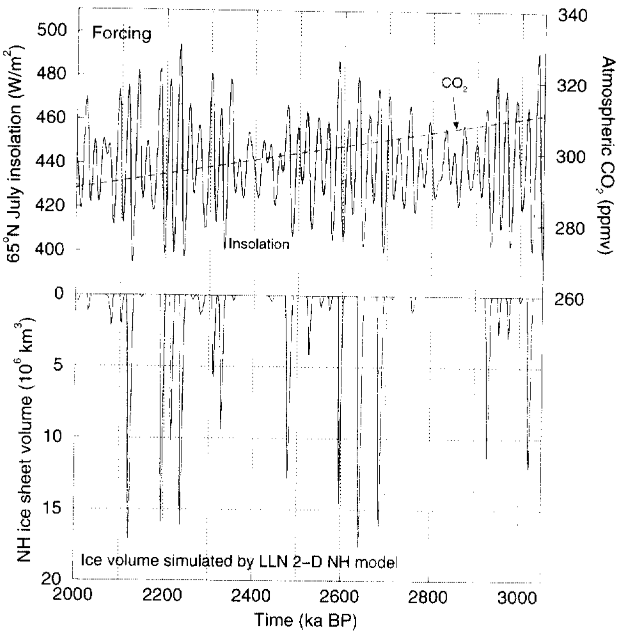


Fig. 7. July 65°N insolation variation (Loutre and Berger, 1993) (solid line, upper panel), the tectonically-induced linear CO$_2$ concentrations (Saltzman *et al.*, 1993) (dashed line, upper panel), and the simulated Northern Hemisphere ice sheet volume with the LLN 2-D model (solid line, lower panel), from 3.05 Ma to 2 Ma BP (Li *et al.*, submitted).

brought the Northern Hemisphere to critical conditions for glaciation after 3 Ma, and then orbital forcing induced a rapid transition to the glacial–interglacial climate regime in the Northern Hemisphere between 2.75 Ma and 2.55 Ma. This theory is supported by the recent simulation of the Northern Hemisphere ice-sheet volume variation made by Li *et al.* (in press) with the LLN 2-D model (Galleé *et al.*, 1991, 1992). In this experiment, ice volume fluctuations were forced by insolation variations (Loutre and Berger, 1993) and the assumption of a linearly decreasing atmospheric CO$_2$ concentration (Saltzman *et al.*, 1993), and ignoring other plausible differences in boundary conditions (e.g. meridional heat flux, geography). Three major periods of glaciation were simulated between 3 Ma and 2 Ma (Fig. 7): one before 2.9 Ma, one between 2.75 Ma and 2.45 Ma, and one between 2.35 and 2.1 Ma, each one corresponding to the periods of large amplitude in the insolation signal. Among them, the glaciations before 2.9 Ma had the smallest amplitude, coeval with the minor increase in the magnetic susceptibility (IRD proxy) record of Site 882 and with geological evidence from the Norwegian Sea and Iceland which suggests a minor increase in glaciation in the Arctic, Iceland and Scandinavia at 3 Ma. A significant increase in ice rafting occurred in the Northwest Pacific Ocean at 2.75 Ma, as revealed by a dramatic drop of the planktonic foraminifera $\delta^{18}O$ record of Site 882. This is reinforced by a sudden increase in simulated Northern Hemisphere ice sheet volume at around that time. As the ice volume simulation is mainly forced by insolation variations, the coincidence between the geological observations and the numerical modelling confirms the role of orbital forcing in triggering the intensification of Northern Hemisphere glaciation at around 2.75 Ma.

CONCLUSIONS

(1) Minor increases in ice rafting have been detected in the Norwegian Sea and the Northwest Pacific from 3 Ma onwards.

(2) A dramatic increase in ice rafting occurred in the Northwest Pacific and the Norwegian Sea at 2.75 Ma, suggesting that the Eurasian Arctic and Northeast Asia was significantly glaciated after 2.75 Ma. In the Northwest Pacific, this major change was accompanied by a dramatic drop in sea-surface temperatures (> 7.5°C), or an increase in salinity (1–2‰). This dramatic change was preceded by 80 ka of unusually warm or fresh Northwest Pacific surface waters.

(3) Data from Site 887 in the Northeast Pacific suggest that Alaska became glaciated at 2.70 Ma, 40 ka (one obliquity cycle) after the glaciation of Eurasian Arctic and Northeast Asia. Sites 609, 610 and 552 in the Atlantic Ocean indicate that the North East American continent was first glaciated at about 2.74 Ma and its ice volume increased dramatically at 2.54 Ma, about 200 ka after the glaciation of Eurasian Arctic and Northeast Asia. Hence, the initiation and intensification of Northern Hemisphere Glaciation was driven by increases in ice volume in the Arctic and North Pacific, as opposed to the Laurentide ice sheet dominated scenario of the late Pleistocene.

(4) Tectonic forcing, and the resultant lowering of atmospheric CO$_2$ might have gradually cooled the Northern Hemisphere, but it is the orbital forcing that triggered the intensification of Northern Hemisphere glaciation between 2.74 Ma and 2.54 Ma.

(5) Simulation of Northern Hemisphere ice sheet volume with the LLN 2-D model shown in this study, supports the theory that orbital forcing was the triggering mechanism of the intensification of Northern Hemisphere glaciation.

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