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The evolution of oceanic oxygen-isotope variability in the North Atlantic over the past three million years

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Throughout the past three million years, variability in the oxygen-isotopic composition of the ocean, caused by changing ice-sheet mass on the continents, has been concentrated at the frequencies associated with changes in the earth's orbital geometry. The amplitude of variability has increased towards the present. An increase in variability associated with changes in the obliquity of the Earth's rotational axis (period 41 ka) during the early Pleistocene was followed by an increase in power related to the precession cycle (23 ka) and associated ellipticity cycle (*ca.* 100 ka) during the past million years. Although deep-sea sediments are the best place to observe this evolution in climatic variability, we will not be able to understand it without more data from other geological sources.

Introduction

The oxygen-isotope record in benthic Foraminifera from deep-sea sediments reflects both the history of ocean isotopic composition (which varies with the quantity of isotopically light ice stored on the continents) and small changes in deep-water temperature. In principal, this record provides the clearest single indicator of the evolution of Plio-Pleistocene climatic variability.

The first Pleistocene oxygen-isotope records from the North Atlantic were obtained by Emiliani (1955), who analysed planktic Foraminifera. In particular, Swedish Deep-Sea Expedition core 280, from which Emiliani analysed the species *Globorotalia inflata*, revealed in detail the now-familiar succession of the past two glacial–interglacial cycles. It is generally agreed that, in showing his findings as surface-temperature records, Emiliani underestimated the contribution that changing ice-volume makes to the overall record (Olausson 1965; Shackleton 1967). This is clear from the very similar character of oxygen-isotope records derived from analyses of benthic Foraminifera, which inhabit the cold abyssal depths where the possibility of further refrigeration during glacial times is severely restricted (Shackleton & Opdyke 1973; Ninkovich & Shackleton 1975). On the other hand, it is not recognized that, in the North Atlantic Deep Water, where the present-day temperature is around 3 °C, changes in deep-water temperature have not been negligible (Duplessy *et al.* 1980; Shackleton *et al.* 1983; Chappell & Shackleton 1986).

Probably the only area where oxygen-isotope analyses of benthic Foraminifera might yield a pure ice-volume record is the Norwegian Sea, where deep water forms at about −1 °C. Labeyrie *et al.* (1988) have obtained an important record by stacking several fragmentary sequences, but it seems unlikely that a truly detailed record will ever be available from that region. Thus a record from the open North Atlantic is at present the best means we have for examining the development of glacial–interglacial variability.

Deep-Sea Drilling Project site 552A

DSDP site 552A was cored on the flank of Rockall Bank by using the then newly developed hydraulic piston-corer (HPC) during DSDP leg 81 (Roberts *et al.* 1985). The site was located at 56° 03′ N, 23° 14′ W in 2311 m depth of water. A good magnetostratigraphic record extending through the Brunhes, Matuyama and Gauss and into the Gilbert magnetochrons was obtained by D. Kent; this, together with J. Backman's nanofossil biostratigraphy, provides a very reliable basic chronology for the lithostratigraphic and stable-isotope records (Shackleton *et al.* 1984; Shackleton & Hall 1985; Zimmerman *et al.* 1985; Backman 1985).

Lithologically the sediments from the past 2.5 Ma consist, in alternating units, of white nanofossil–foraminiferal ooze (deposited during interglacials) and greyish sandy sediment containing Foraminifera and ice-rafted detritus (deposited during glacials (Zimmerman *et al.* 1985)). Before 2.5 Ma BP, only white ooze was deposited. It is clear from the virtual absence of nanofossils in the glacial sediments that the polar front moved south of the site during glacial episodes back to 2.4 Ma BP. Apart from a very brief and minor pulse of ice-rafting at *ca.* 2.50 Ma BP, there were no ice-rafted grains reported from sediments older than this transition. It seems certain that, at least at the latitude of Britain, this transition just after 2.5 Ma BP constituted a major climatic break.

Oxygen-isotope measurements were made on benthic Foraminifera by using standard methods (Shackleton & Opdyke 1973; Shackleton *et al.* 1983). Raw data are listed in Shackleton & Hall (1984). In the present paper, as in Shackleton *et al.* (1983), they are shown after applying adjustment factors for those species that exhibit a systematic departure from isotopic equilibrium.

Oxygen and carbon isotope measurements were made at 10 cm intervals in the top 12 cores, covering the past 3.6 Ma, excluding core 7 which was badly disturbed in the coring process and was not considered worth sampling in detail. Although this provides a favourable average sampling interval of 2.5 ka, the reality is that the sampling interval represents widely varying time intervals owing to the large variations in the sediment accumulation rate. Obtaining a detailed timescale for the past 0.9 Ma was straightforward, because the record can easily be correlated with other published isotope records. In figure 1 this part is shown on the SPECMAP timescale (Imbrie *et al.* 1984).

For the section older than 0.9 Ma BP, little has been done so far to develop a detailed chronology because there have been few high-resolution data sets available to work with. Pisias & Moore (1981) examined the distribution of variance in the oxygen-isotope record of core V28-239 (Shackleton & Opdyke 1976) and concluded that, in the portion older than 0.9 Ma, variance was concentrated at a frequency of *ca.* 2.5×10^{-4} cycles ka^{-1}, corresponding to the 40 ka period associated with variations in the tilt of the Earth's rotational axis (obliquity). They proposed a detailed timescale for core V28-239 by preserving a constant phase relation between ^{18}O and tilt (as calculated by Berger (1978)). However, the degree of detail preserved in core V28-239 (accumulated at only about 1 cm ka^{-1}) is not sufficient to permit DSDP site 552A to be correlated accurately to it, especially in view of the gap in 552A resulting from the disturbance of core 7. More recently, Ruddiman *et al.* (1986) published oxygen-isotope data from DSDP site 607 and again showed that variance is concentrated at the tilt period. By using their data, it is possible to achieve a satisfactory solution for the detailed chronology of site 552A back to the Olduvai chron at 1.6 Ma BP.

In the lower Matuyama we have developed a provisional timescale based only on tuning the 552A record to orbital variations. The difficulty in doing this with satisfactory accuracy is that the accumulation rate has been so variable at this site. The records from sites such as DSDP 607, which is south of the southernmost excursions of the polar front, will ultimately provide a more accurate chronology for this interval also.

Time-series analysis

Table 1 lists the age control points that we have used to construct the time series in figure 1. Ages of samples between these were estimated by linear interpolation. To obtain estimates at uniform time intervals of 3 ka for spectral analysis, we used a Gaussian weighting method which has the effect of providing some smoothing as well as interpolation. The smoothing used is equivalent to that of a Gaussian window with a halfwidth of 3 ka. Interpolation points for which data are not available are not used in the spectral estimation.

Table 1. Age control points used for DSDP site 552A

depth/m	age/Ma	depth/m	age/Ma	depth/m	age/Ma
00.0	0.0	13.20	0.689	35.50	1.953
00.67	0.024	13.60	0.711	35.98	1.996
01.62	0.065	14.13	0.726	37.70	2.102
01.63	0.078	14.50	0.736	37.99	2.134
02.00	0.107	15.89	0.797	38.43	2.170
02.40	0.128	17.60	0.874	39.01	2.201
02.80	0.157	19.90	0.980	39.60	2.235
03.50	0.194	21.00	1.045	40.40	2.287
04.50	0.249	21.88	1.081	41.10	2.334
05.00	0.269	22.69	1.120	41.40	2.358
05.30	0.297	23.95	1.200	43.14	2.482
05.55	0.329	29.10	1.476	43.90	2.569
06.25	0.371	29.33	1.494	45.40	2.720
06.46	0.405	29.80	1.529	46.30	2.800
06.90	0.423	31.00	1.598	47.50	2.906
08.40	0.471	31.80	1.652	48.30	2.955
09.15	0.513	32.50	1.706	50.50	3.075
09.60	0.540	33.10	1.748	53.40	3.232
10.80	0.620	34.10	1.823	55.60	3.364
12.00	0.656	34.60	1.874	56.10	3.401
12.60	0.668	34.92	1.896	57.20	3.470

Spectra were calculated by using a modification of standard Fourier analysis methods (Jenkins & Watts 1968). The principal modification is that we have calculated variance with a constant bandwidth on a logarithmic frequency scale. (In most studies the bandwidth is specified as a constant frequency interval, a practice that yields very different degrees of resolution across the spectrum and makes it difficult to draw quantitative inferences).

In figure 1, spectra are shown for overlapping time intervals of 0–1, 0.5–1.2, 1.5–2.5, 2.0–3.0

BIOLOGICAL SCIENCES THE ROYAL SOCIETY PHILOSOPHICAL TRANSACTIONS OF

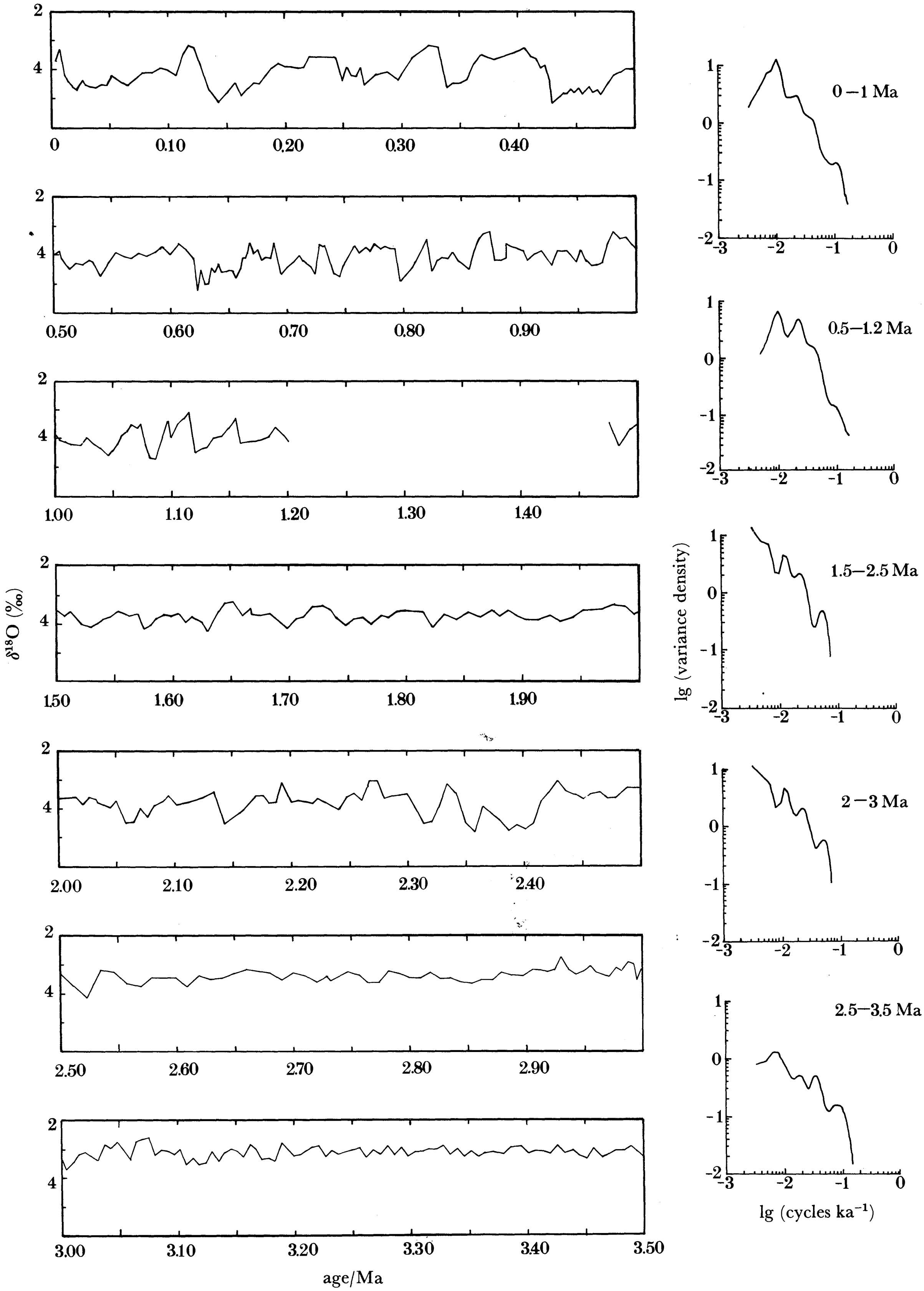

FIGURE 1. Oxygen-isotope record of DSDP site 552A; timescale from table 1. Spectra to the right are calculated with a constant bandwidth in log F for the interval; variance-density scale is $(‰)^2$ $(\text{cycles/ka})^{-1}$.

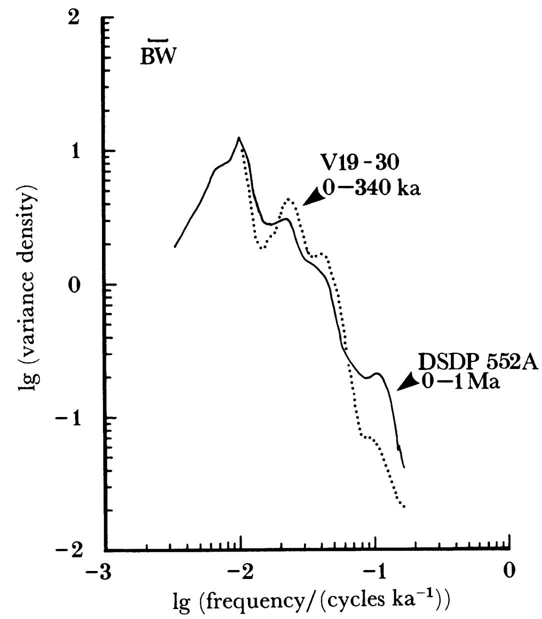

FIGURE 2. Spectrum for 0–1 Ma BP in DSDP 552A compared with that of 0–340 ka BP in piston core V19-30. Both are records of ^{18}O variations in benthic Foraminifera (proxy for global ice volume).

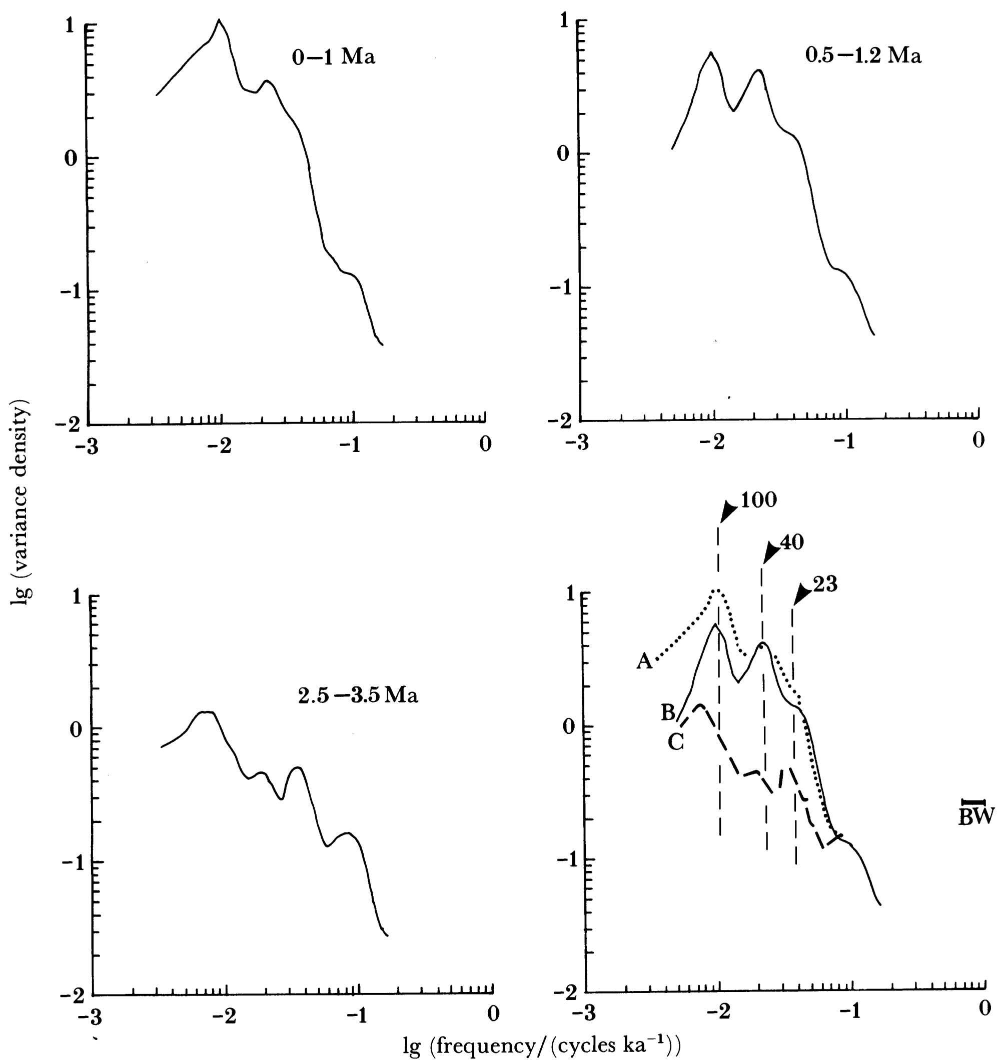

FIGURE 3. Spectra for the 0–1 Ma BP, 0.5–1.2 Ma BP and 2.5–3.5 Ma BP sections of DSDP 552A (from figure 1) compared, to demonstrate the growth in variance associated with earth-orbital variations. In the fourth sector these are superimposed with identical scales for variance density and labelled A (0–1 Ma), B (0.5–1.2 Ma) and C (2.5–3.5 Ma).

and 2.5–3.5 Ma BP. These spectra demonstrate the progressive increase towards the present in the variance associated with the Milankovitch frequencies.

For the interval 0–1 Ma BP the spectrum is rather similar to others for the same interval. Figure 2 compares it with that for core V19-30 (0–340 ka BP (Shackleton & Pisias 1985)) the same variance-density scale. The spectrum for V19-30 is 'cleaner', with a more pronounced minimum between the 100 ka and 40 ka peaks, and it also shows somewhat more variance at 23 ka. These differences almost certainly arise mainly from the fact that the sampling interval is three times closer in V19-30.

Figure 3 compares three of the spectra from figure 1 (again on the same variance-density scale). If we compare the 0–1 Ma BP and 0.5–1.2 Ma BP sections, the most striking changes is in the concentration of variance at the 100 ka period. This change has been noted before (see, for example, Ruddiman *et al.* 1986) but is particularly clear here. The section from 1.5 to 2.5 Ma BP has similar variance density but the spectrum is less clear, probably because of uncertainties in the timescale. The section from 2.5 to 3.5 Ma BP displays significantly less variance (which is certainly not an artefact, because the sampling interval is still adequate to reveal variance in this band). In fact the main impediment to developing a good timescale for this interval is the less favourable signal: noise ratio.

Discussion

Climatic variance covers all frequencies; in this paper we are concerned only with variance in a restricted bandwidth between about 5×10^{-3} and 2×10^{-1} cycles ka^{-1}. This is of course an interesting bandwidth for a quaternary geologist because it encompasses ice-age cycles. From a more theoretical standpoint, it is interesting because there is a clearly defined external forcing to the Earth's climate system in this bandwidth. In the geological record, cyclic sedimentation that may well arise as a result of this external forcing is widely known (see, for example, papers in *Paleoceanography*, vol. 1, no. 4 (1986). However, from a climatologist's point of view, the earlier geological records are at present disappointing because it is difficult to assess the scale of the climatic variations that have given rise to the sedimentological cycles observed. For the past few glacial cycles we have a growing understanding of the scale and spatial pattern of climatic response to orbital variations as an outcome of the CLIMAP project (CLIMAP Project members 1976, 1981) and following studies. The analysis displayed in figure 3 gives an impression of the way in which the amplitude of this climatic response to a rather uniform forcing has built up.

The change observed between the 0.5–1.2 Ma BP interval and the 0–1.0 Ma BP interval is interesting because it displays the association between the 100 ka cycle (eccentricity) and the 23 ka cycle (precession). There has been much discussion regarding the origin of the 100 ka cycle because it is not intuitively obvious that there should be any marked response at this period whereas visually it is the major feature of ice-age cycles. Some workers have suggested that the origin is unrelated to orbital variations. The present analysis cannot exclude the possibility that there is some resonance or instability in the climate system near 100 ka (Saltzman & Sutera 1987). However, the fact that response at this period increased over the same interval during which response to the precessional cycle increased supports the notion that the strong 100 ka cycle is a result of the long time-constant and nonlinear response associated with massive continental ice sheets. As these extended to lower latitudes, the precession effect became more and more important in controlling the distribution of solar insolation and hence in controlling their mass balance.

Passing now to the older part of the record, it is clear from figure 3 that it will be necessary to improve the signal: noise ratio of stable-isotope records before the magnitude of any response to orbital forcing in this type of record can be accurately described. Considerable subtlety will be needed if we are to build up a quantitative picture of the climatic response to orbital forcing in preglacial times.

DSDP site 552A has recovered a record of changes in globally integrated ice volume; of heavy deposition of iceberg-transported material at the latitude of Britain; and of oscillations in the position of the polar front. The location of the site is ideal for recording these aspects of climatic variability in the past, and as correlations are improved we may expect an ever-increasing similarity between this record and those reconstructed from northwest Europe. At present it is perhaps only the vegetational records of The Netherlands that offer the promise of comparability in detail, despite the fact that all the areas surrounding the North Atlantic Ocean must have experienced local variants of the same pattern of climatic change. It is appropriate to consider the question: in areas where the prospect of reconstructing long records that are suitable for comparison with the deep-sea records is remote, is there any palaeoclimatic justification for continued Quaternary research?

The answer to this question must surely be in the affirmative; there are many questions which can be answered better by terrestrial than by deep-sea records. One example concerns the interglacials. A cursory examination of the oxygen-isotope record suggests a general trend towards more positive values over the past three million years, but a more careful examination shows that this is only a result of the trend towards more positive values at the glacial extremes. If the Earth's climate system has become gradually more sensitive to glacial tendencies over this interval, one would surely expect some trend in the climate of the interglacials, unless the mechanism involved in this gradual change is very tightly tied to the actual ice sheets. Perhaps vegetational records would give us crucial information regarding the trend in interglacial climates through the past three million years if examined with this question in mind. Soil profiles, and the floras and faunas preserved in interglacial marine highstand deposits, are other sources of information which may be very valuable in this respect, despite the fact that they provide necessarily incomplete records.

At the opposite climatic extreme, the oxygen-isotope record portrays only the gross ice volume recorded at successive glacial maxima; the spatial distribution of ice masses can only be reconstructed by field observations. The question of the location of the ice sheets that came and went between 2.5 Ma BP and 1 Ma BP with a 40 ka periodicity must be answered if we are to understand why their response was different from that of the more recent ice sheets. It is to be hoped that the discussions at this meeting will help workers in different fields to exploit the full potential of their contributions from the point of view of their significance for the understanding of climatic variability. This should provide the scientific rewards that arise from interacting with a wider research community and especially with climate modellers.

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BIOLOGICAL SCIENCES THE ROYAL SOCIETY PHILOSOPHICAL TRANSACTIONS OF

Discussion

J. Rose (*Department of Geography*, *Birkbeck College*, *University of London*, *U.K.*). It would be difficult to leave a meeting such as this without an assessment of relation of the main elements of the terrestrial record of northwest Europe with main elements the oceanic record. The forces driving the global atmospheric system are now well known and the dominant components of these forces during different parts of the Quaternary have been identified (Ruddiman & Raymo, this symposium). I wish to draw attention to the parallel between the main elements of the ocean record and the main elements of the terrestrial and offshore record in northwestern Europe when considered at the scale of geomagnetic chrons rather than conventional stages.

In the ocean record the large-amplitude $\delta^{18}O$ signature which is dominant throughout most of the Brunhes chron (0–0.6 Ma BP) (Ruddiman & Raymo, this symposium) is considered to reflect the build-up of continental ice masses. In northwestern Europe this is reflected in the glacial–interglacial cycles during which ice extended into midland and southern Britain and across part of the north European plain. The period of maximum expansion of glaciers appears to have been during the Anglian in Britain and the Elsterian of northern Europe, which are equated with oxygen-isotope Stage 12 (Bowen *et al.* 1986). This same stage has been identified from oceanic evidence as the period of maximum global cooling (Shackleton 1987, this symposium). It is probable that traces of earlier continental glaciation during the Brunhes have been obscured by the effects of the Anglian and Elsterian ice sheets. Subsequent glacial advances are almost as obscure because of the extent of the late Devensian–late Weichselian ice sheets. Nevertheless, this part of Quaternary time in northwestern Europe saw the expansion of continental ice sheets during the cold stages.

During the later part of the Matuyama chron (0.6–1.7 Ma BP) the amplitude of the $\delta^{18}O$ oceanic signal is reduced relative to the Brunhes but is still distinct (Ruddiman & Raymo, this symposium). In northwestern Europe this atmospheric force appears to be represented by glaciation in the mountains and extensive fluvial deposition in the lowlands. Typically, rivers such as the Thames and the Rhine transported large bodies of fluvial sediment, which are known respectively as the Kesgrave and Sterksel Formations of East Anglia and the Low Countries. In the southern North Sea, where these bodies of sediments coalesce, they are known as the Yarmouth Roads Formation. As far as is understood at present, this pattern of environmental change in Britain occurred between pre-Pastonian A and the Beestonian, including at least four glacial events in the Welsh uplands (Bowen *et al.* 1986); in the Netherlands it encompasses the period referred to as the Menapian, 'Bavel Complex' and early part of the 'Cromer Complex' during which time there is evidence for at least five glaciations in the headwater mountains (Zagwijn 1986).

The reason for this environmental pattern in northwestern Europe at the later part of the Matuyama chron can be explained in terms of the lower magnitude in the variation of the atmospheric forcing, with climatic severity sufficient only to maintain glaciation in western uplands and high mountains, but not sufficient to cause glaciation at a continental scale. The reasons for the exceedingly large bodies of fluvial sediments, such as have not been produced in the Quaternary either before or since, can be explained by the effectiveness of glacial processes during early stages of glaciation acting on a landscape that is not adjusted to glacial flow patterns. In these circumstances erosion would be enhanced by positive feedback until relatively adjusted large-scale glacial erosional landforms were produced and a self-regulating

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

system was established in which erosional processes would be dampened down. (Professor Andrews made this same point with regard to North America in response to a question by Professor Sugden on the origin of fiords.)

Finally, in the period of the Matuyama chron before 1.7 Ma BP, the atmospheric variation is, in general, of a much lower magnitude. In northwestern Europe the response is to be seen in the relatively fine-grained sediments transferred to the North Sea Basin by rivers such as the Thames and Rhine, with catchments smaller even than at the present day. In the southern North Sea these deposits are represented by the Westkapelle Ground Formation, the IJmuiden Ground Formation and the Winterton Shoal Formation (Balson & Cameron 1985). In East Anglia they are represented by the Red Crag and Norwich Crag Formations (Bowen *et al.* 1986), and in The Netherlands by the Oosterhaut and Maassluis Formations. In The Netherlands, lower-energy fluvial equivalents exist in the form of the Scheemda Formation, Kieseloolite Formation, Tegelen Formation, Harderwijk Formation and Kedichem Formation (Zagwijn 1986).

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PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES