Chapter 2 Late Cainozoic environmental change

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Introduction

The Quaternary Period is part of the Cainozoic Era and it has been characterized by periodic, major, climatic changes that led to the growth and significant expansion of large ice-caps in the temperate mid-latitudes. The Quaternary Period is divided into the Pleistocene (from 1.6 million years ago (Ma) to 10 000 years ago (ka)) and the Holocene (the past 10 000 years) epochs (Table 2.1). Although the terms 'Quaternary' and 'Pleistocene' are often synonymous with the expression 'the Ice Ages', the onset of glaciation in the Southern Hemisphere has been dated back well beyond early Tertiary times, and a continental ice sheet first developed in Antarctica about 36 Ma. Although extensive glaciation has occurred in Alaska and Greenland since about 17 Ma, it became important in the zones bordering the Atlantic Ocean, including Britain, only from 2.9 Ma onwards (Shackleton *et al.* 1984; Ruddiman and Raymo, 1988).

The definition of the base of the Quaternary Period

The most usual date for the beginning of the Quaternary Period has been suggested as 1.64 Ma, although as defined by the International Union of Geological Sciences (IUGS) it occurs just below the top of the Olduvai magneto-subchron at Vrica in Italy (Aigurre and Passini, 1985), where its age, on an astronomically tuned timescale, is estimated at 1.905 Ma (Shackleton *et al.,* 1990). However, in The Netherlands the preference is for 2.45 Ma and north-west Europe shows a major faunal change at *c.* 2.5 Ma (Zagwijn, 1974; Gibbard *et al.,* 1991). The Stratigraphic Commission of the International Union for Quaternary Research (INQUA) is currently seeking a suitably defined boundary stratotype and if an age of 2.5 Ma is secured on a sound lithostratigraphical basis, then it will coincide with the appearance of loess in the Chinese stratigraphical record, the first evidence of substantial ice-rafting in the North Atlantic Ocean and the first appearance of *Homo habilis* in Africa (Bowen, 1999).

Divisions of the Quaternary Period

Some of the climatic phases of the Quaternary Period have been as warm as, or warmer than the present day, whereas others have been cold enough to generate extensive glacial and periglacial conditions. The evidence in British Quaternary deposits of cold and temperate alternations has formed the traditional division of the period. In the lower part of the land-based sequence the sediments are mainly of shallow-water marine origin. In the later stages, the sediments are largely terrestrial. The overall sequence therefore generally has been interpreted as indicating an initial pre-glacial phase, followed by one when glacial and periglacial environments alternated with temperate ones. Interglacials are temperate, climatic intervals between cold stages, with interstadials representing climatic ameliorations within colder stages. However, the precise number of cold and warmer phases is debatable. Traditionally seven cold stages, together with eight temperate ones, have been recognized in the British Pleistocene record (Jones and Keen, 1993). However, over the past 700 000 years at least six colder phases similar to that of the Last Glacial Maximum are suggested for a deep-sea core (Prell *et al.,* 1986) and there are as many, or more, interglacial climatic fluctuations, similar to, although not notably warmer than, the current prevailing climate (Figure 2.1). Many more such glacial–interglacial fluctuations occurred in the Quaternary Period before 700 000 years ago, and in fact had been occurring during the Tertiary Sub-era.

In the British offshore record, on the continental shelf, longer sequences of Pleistocene sediments have been located than on land. Within these longer sequences there are unc onformities and overall the number of stages represented appears to be similar to the traditional land-based record. These traditional British sequences are at variance with the primary worldwide record for climatic change, which is derived from oxygen-isotope analyses and palaeomagnetic studies of deep-ocean core sediments. The sequential variation of the oxygen isotope content in these core sediments is taken to be an indicator of variations in global ice volume.

The oxygen isotope record

Periods with different oxygen isotope ratios are identified in the sediment cores and are called 'Oxygen Isotope Stages' (OIS), and these provide the stratigraphical framework (Figure 2.2) and (Table 2.2). During glacials, the oceans are relatively enriched in ^{18}O, because the water that evaporates is ^{16}O-enriched and the resultant precipitation is trapped in the ice sheets. High ^{18}O to ^{16}O ratios in sea water, expressed as $\delta^{18}O$, the change in ratio relative to a reference standard, therefore indicate a relatively high global ice volume. Past isotope ratios can be estimated from the oxygen isotope content of calcareous marine fauna deposited on the ocean floor. The deposition process is dependent, in part, on temperature and other factors, but about two-thirds of the variance in the isotope-ratio records from calcareous sediments can be related to changes in global ice volume. Various ocean-core records collectively cover at least 2.6 Ma (Shackleton *et al.,* 1990) and a widely used composite is the SPECMAP record (Imbrie *et al.,* 1984). These records indicate multiple long-term trends of increasing ice volume, followed by abrupt terminations with a return to interglacial conditions. The analysis of the core called 'Deep Sea Drilling Project 552A', in terms of oxygen isotope content, has been extended back to 3 Ma (Shackle-ton *et al.,* 1988). The analysis indicates a major climatic break at about 2.5 Ma, with no evidence of ice-rafting before that time. This record also indicates a trend towards more severe glacial conditions over the past 3 million years but no trend towards milder interglacials. However, the long-term trends of increasing ice-volume have shorter-term fluctuations imposed on them. Based on the SPECMAP data, Thorne (1996), using data from Goodess *et al.* (1991), illustrates a global climatic index for the past 780 000 years (Figure 2.3), which shows four complete glacial–interglacial cycles over the past 400 000 years and suggests that fully interglacial conditions, with ice sheets limited to their present-day extent, existed for only 9% of the time. Fully glacial conditions similar to those in the Devensian glacial in Britain, occurred for 16% of the time, whereas the remainder of the record was shown to be Boreal (38%) or Periglacial (37%).

(Table 2.2) Relationships between the British Quaternary stratigraphical classification (after Mitchell *et al.,* 1973), selected lithostratigraphical units, oxygen isotope stratigraphy and polarity (from Bowen, 1999).

1st Edition (1973)	**Lithostratigraphy**	**Aminozone**	**D-alle/L-Ile$^{\$}$**	**Age (ka)†**	**$\delta^{18}O$ and polarity**
	Hailing Member	Hailing	0.036 ± 0.01 (3)	10.9 ± 0.12 (^{14}C)	2
Devensian	Stockport Formation Δ Upton Warren Member	Upton Warren	0.07 ± 0.007 (3)		3
	Cassington Member	Cassington	0.08 ± 0.009 (6)	~80 (OSL)	5a
Ipswichian	Trafalgar Square Member	Trafalgar Square	0.1 ± 0.001 (11)	124 ± 5.4 (U)	5e
	Ridgacre Formation Δ Kidderminster Member			159.5 ± 13 (36Cl)	6
Wolstonian	Strensham Court Bed	Strensham	0.17 ± 0.01 (4)	~200 (OSL)*	7
	Rushley Green Member				8
	Hoxne Formation	Hoxne	0.26 ± 0.02 (9)	319 ± 38 (ESR)	9
Hoxnian	Spring Hill Member				10
	Swanscombe Member	Swanscombe	0.3 ± 0.017 (34)	~400 (U)* 471 ± 15 (TL)*	11
Anglian	Lowestoft Formation A				12
	West Runton Member	West Runton	0.35 ± 0.01 (9)	~500 (ESR)	13
Cromerian	Waverley Wood Member	Waverley Wood	0.38 ± 0.026 (5)		15
	Kenn Formation Δ				16

Grace Formation ‡	Grace	0.43 ± 0.02 (4)	810 ± 140 (ESR)	21

$ Number of analyses in parentheses

† Age estimate – method in parentheses

* Age established at another locality of the aminozone

∆ Glacial formation

‡ Somme Valley, France

Ice cores and loess records

Ice cores and loess records provide information complementary to the ocean sediments relating to Quaternary climatic change. Ice cores do not produce as long a record as that from ocean cores, going back only to about 250 ka, but they have a higher temporal resolution, up to the decadal and annual scale in some cases. Climatic records can be reconstructed from such ice cores using a variety of indicators and these have provided detailed climatic reconstructions for the last glacial–interglacial cycle (Figure 2.4). Loess forms during periods of relatively cold and semi-arid climatic conditions, which are characterized by a poorly vegetated, semi-desert environment (Kukla *et al.,* 1988). The loess record can go back as far as ocean core records and the high-resolution magnetic stratigraphy for the record in north-west China extends back to 700 ka, whilst the basal loess from central China is thought to date from about 2.5 Ma. The general pattern of agreement is good between all the loess and ocean records.

Overall the various global records indicate that the glacial–interglacial cycles became more marked at about 450 ka and since then, five interglacials have occurred (at *c.* 400 ka, 320 ka, 200 ka, 120 ka and 10 ka), with four glacials separating them (at *c.* 340 ka, 250 ka, 150 ka and 20 ka).

Key long-term climatic change records for the north-east Atlantic have been illustrated from continuous deep-ocean cores, for example a composite record of North Atlantic sea-surface temperature change over the past 1.1 million years has been created by joining together data from cores (Ruddiman *et al.,* 1986). This shows an increasing amplitude of sea-surface temperature fluctuations in the more recent section. Progressively colder glacial minima are indicated from 850 ka onwards. During the period 700 ka to 400 ka the interglacial maxima becomes progressively warmer. During this latter period, the amplitude of the dominant cycle (about 95 000 years) increased by a factor of four and the amplitude of the sea-surface temperature fluctuations during the past 400 000 years indicates that temperature differences between glacial maxima and interglacial maxima were 11°C in summer and at least 8°C in winter (Goodess *et al.*, 1991).

Sea-level change

This climatic change also has affected eustatic sea levels throughout the Quaternary Period, and because present-day global temperatures are relatively high compared with those over most of the Quaternary, sea levels are too, hiding much of the direct evidence for past sea-level change. There have been advances too in the investigation and dating of earlier Quaternary sea levels. For example, Ikeda *et al.* (1991) have used the electron spin resonance method to date Middle Pleistocene corals from the Ryuku Islands in Japan. This method is considered to be useful for dating aragonitic corals and for time periods beyond the limits of the uranium-series method. Two high sea-level stands were identified at about 800 ka and 600 ka and they conclude that these are correlated with at least two of the warm marine Oxygen Isotype stages 23, 21, 19, 17 and 15. A similar approach has been used to date the coral reef sequence from Sumba Island,

Indonesia (Pira7.7oli *et al.,* 1991). Here there are six terraces broader than 500 m that represent major interglacial periods, together with many smaller terraces. The oldest interglacial stage is dated to around 1 Ma.

High sea-level events identified in the coral reef records also can be identified in the SPECMAP record, indicating a degree of correlation between the two. Indications are that there have been high sea levels at 233, 215, 125 and 100 ka. There is no evidence of a high sea level from the Bahamas record at either 195 or 80 ka (Lundberg and Ford, 1994), whereas high sea levels are indicated by the SPECMAP record (Figure 2.3). Coral reef terraces, which because of tectonic uplift lie above present-day sea level, are a major source of evidence about higher sea levels. The reefs develop when sea level rises faster than the land and are preserved when tectonic uplift continues. Most studies are based on reefs from Barbados and New Guinea, regions that are subject to ongoing and steady tectonic uplift. By their nature coral-reef records are discontinuous, recording only sea-level maxima, indicated by terraces, and minima, indicated by reef geometry and evidence of deltaic environments (Goodess *et al.,* 1992). Sea-level changes of the Late Pleistocene also have been studied using uranium-series dates of submerged speleothems from the Lucayon Caverns in the Bahamas (Lundberg and Ford, 1994).

(Table 2.3) Proposed climatostratigraphical stages in Britain (after Mitchell *et al.,* 1973).

Stage	**Stratotype**	**Notes**
Flandrian		Begins 10 ka (^{14}C); base at bottom of pollen zone IV
Devensian	Four Ashes, Staffordshire [SJ 914 082]	Late: 26–10 ka (^{14}C) Middle: 50–26 ka (^{14}C): includes Upton Warren interstadial complex Early: preceding 50 ka (^{14}C): includes Chelford interstadial ~60 ka (^{14}C)
Ipswichian	Bobbitshole, Ipswich [TM 148 414]	Base at beginning of pollen zone II
Wolstonian	Wolston, Warwickshire [SP 411 748]	Includes Baginton–Lillington gravels, Baginton sand, Wolston series, Dunsmore gravels; base at bottom of Baginton–Lillington gravels
Hoxnian	Hoxne, Suffolk [TM 543 977]	Base at beginning of pollen zone HI
Anglian	Corton Cliff, Suffolk [TM 543 977]	Lowestoft Till, Corton Sands, Norwich Brickearth/Cromer Till; base at bottom of lower till
Cromerian	West Runton, Norfolk [TG 188 432]	Upper Freshwater Bed; base at bottom of pollen zone C1
Beestonian	Beeston, Norfolk [TG 169 433]	Arctic Freshwater Bed; base at bottom of pollen zone PI
Pastonian	Paston, Norfolk [TG 341 352]	Gravels, sands and silts; base at bottom of pollen zone BeI
Baventian	Easton Bavents, Suffolk [TM 518 787]	Marine silt; base at bottom of pollen zone L4
Antian	Ludham, Norfolk (borehole at [TG 385 199])	Marine shelly sand; base at bottom of pollen zone L3 (forams: Lv)
Thurman		Marine silt: base at bottom of pollen zone L2 (forams: Lm)
Ludhamian		Shelly sand: base at bottom of pollen zone L1 (forams: Ll)
Waltonian	Walton-on-the-Naze, Essex [TM 267 237]	Older Red Crag; base at bottom of Crag at Walton

A record for global sea-level change from this type of work is available for much of the Quaternary Period in various parts of the world, but this is not the case for northern England where the record is detailed only for the Holocene Epoch.

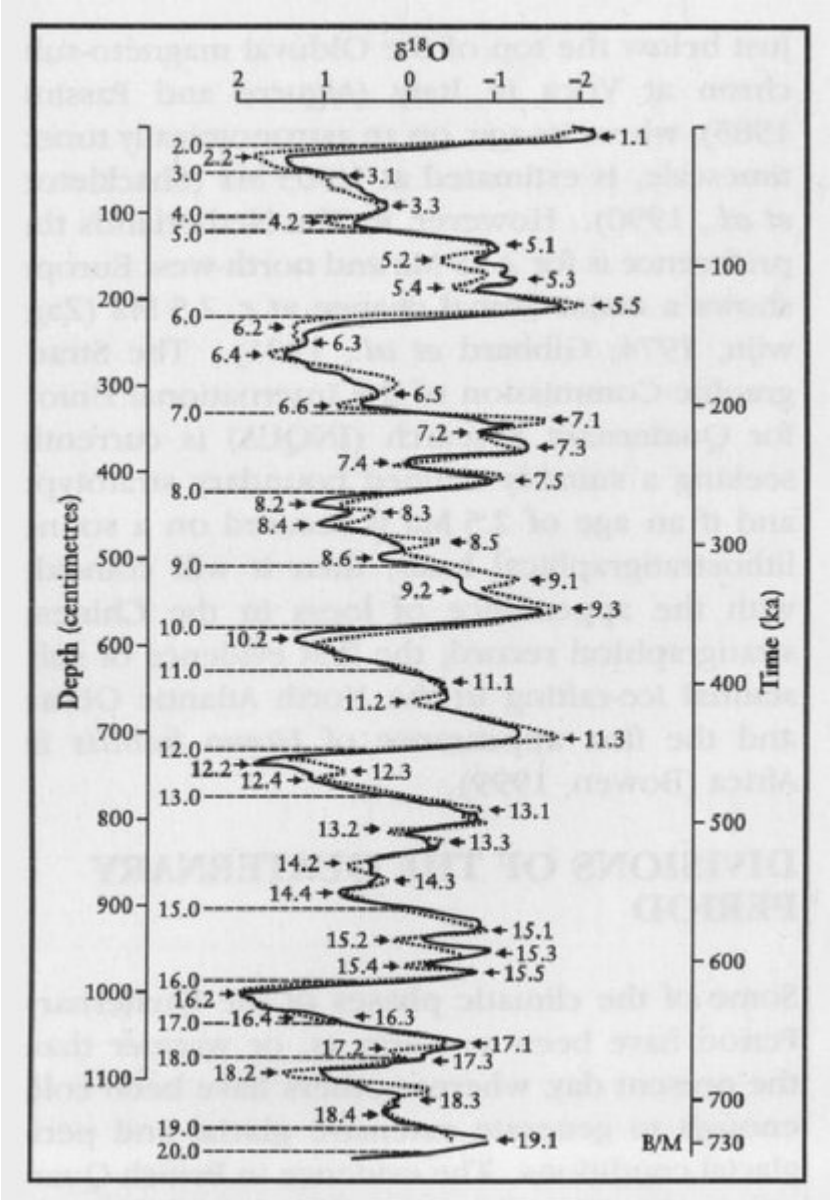
The Quaternary record in Britain

In the British Isles some 14 climato-stratigraphical stages of alternating cold (glacial) and warm (interglacial) conditions (Table 2.3) have been recognized (Bowen, 1978; Jones and Keen, 1993) and the correlation with oxygen isotope stratigraphy and polarity changes is given in (Table 2.2). These stages are often difficult to recognize in the terrestrial record and it is thought that most of the cold stages either did not generate extensive glacial expansion in Britain or are not preserved. In northern England the terrestrial record is dominated by deposits of the last, Devensian cold stage and evidence of earlier cold stages is limited and uncertain. Similarly, evidence of interglacial stages earlier than the current is also limited. Within the Devensian, evidence from Norway shows at least four periods of ice-sheet build up (Larsen and Sejrup, 1990; Baumann *et al.,* 1995), whereas in Britain there may have been only two (McCabe, 1987; Bowen, 1989; Rose, 1989a). In northern England the evidence is almost wholly from the last of these.

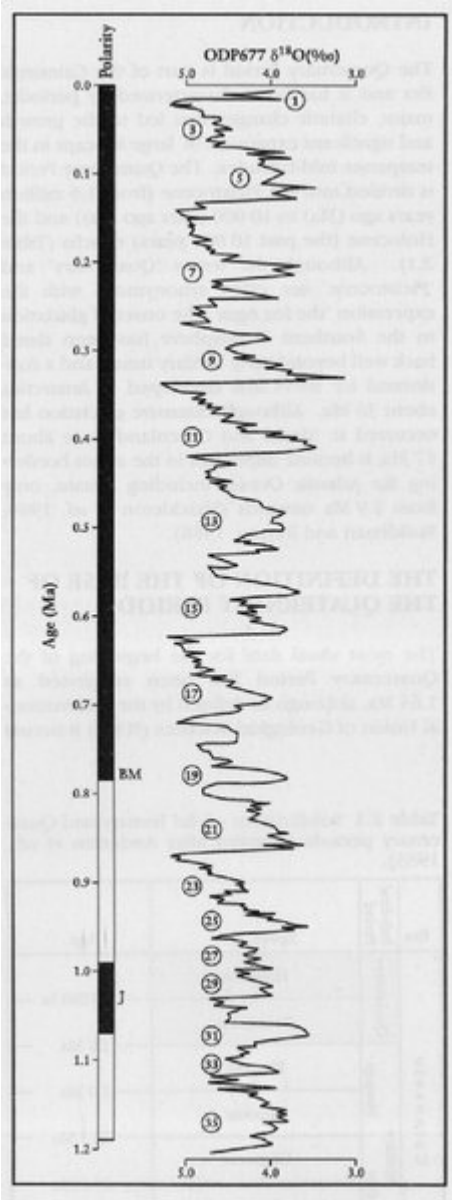
References

Era	Sub-Era/ Period		Epoch	Age
Cainozoic	Quaternary		Holocene	
				10 000 ka
			Pleistocene	
				1.6 Ma
	Tertiary	Neogene	Pliocene	
				5.0 Ma
			Miocene	
				22.5 Ma
		Palaeogene	Oligocene	
				37.5 Ma
			Eocene	
				53.5 Ma
			Palaeocene	
				65.0 Ma
Upper Cretaceous				

(Table 2.1) Subdivisions of the Tertiary and Quaternary periods (adapted after Anderton et al., 1983).



(Figure 2.1) Oxygen isotope stratigraphy (after Prell et al., 1986). A deep-sea core shows the variation of oxygen isotopes with depth (arbitrary units) and time (ka). Fluctuations to the left correspond to larger ice volumes on land (glacials) and conversely fluctuations to the right to reduced on-land ice, corresponding to interglacials. Curves represent different ben-thic foraminifera.



(Figure 2.2) Oxygen isotope record and polarity from Ocean Drilling Program (ODP) site 677 (after Shackleton et al., 1990). Normal polarity is indicated by black, reverse by white. No.s 1–35 are Oxygen Isotope Stages (O1S). BM: Brunhes–Matuyama boundary; J: Jaramillo event.

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* Age established at another locality of the aminozone
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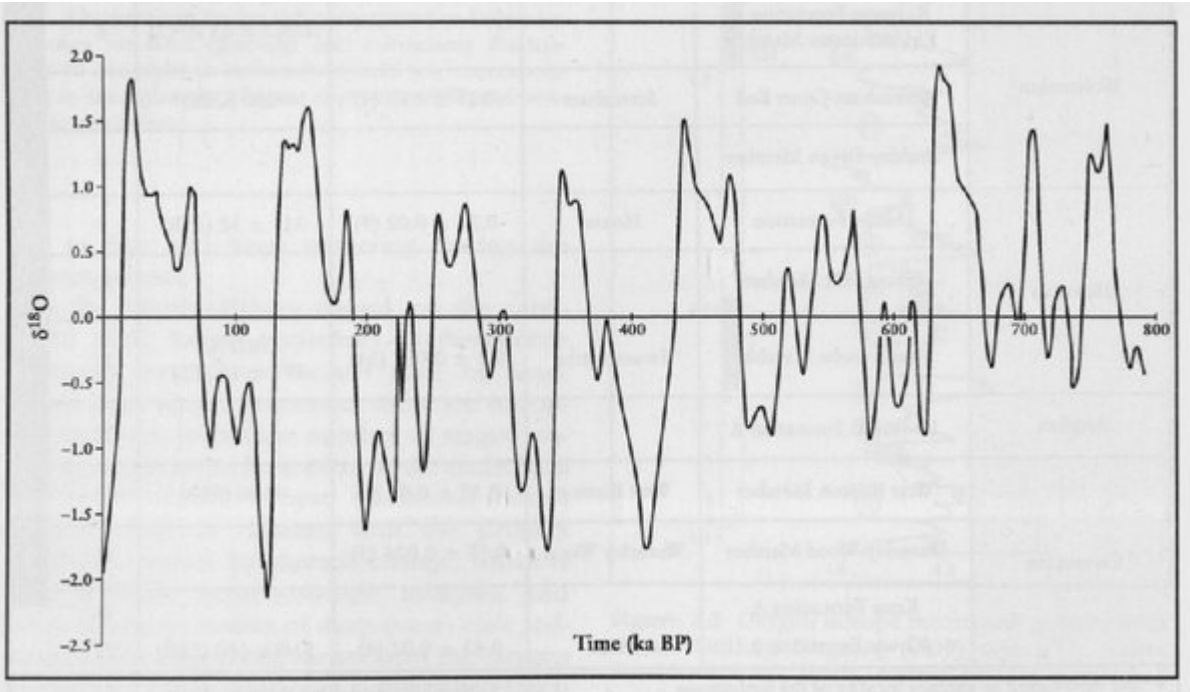
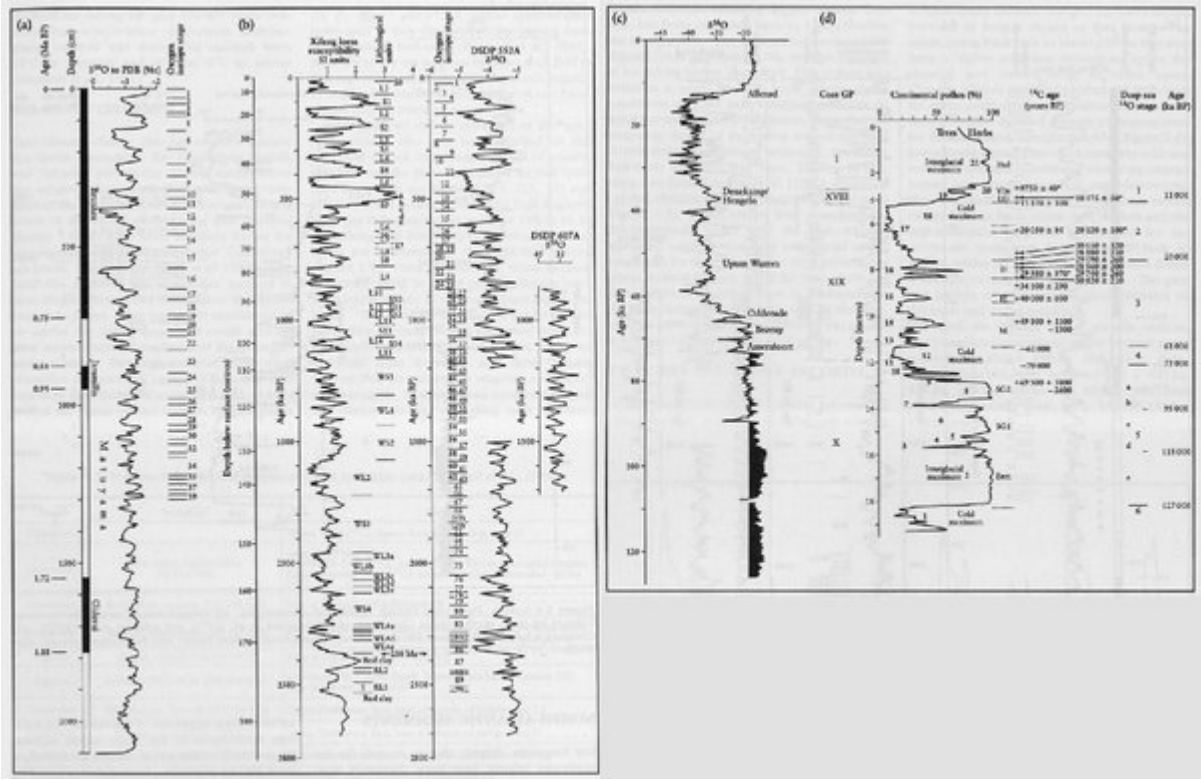


(Figure 2.3) A global climatic index (after Thorne, 1996) based on the SPECMAP $\delta^{18}O$ record (Imbrie et al., 1984).



(Figure 2.4) Pleistocene climatic signals and chronologies. (a) Oxygen isotope and palaeomagnetic record in core V28–239, after Shackleton and Opdyke (1976) and Berggren et al. (1980) PDB: Pee Dee Belemnite. (b) Magnetic susceptibility record in the Xifeng Loess and proposed correlation of the Chinese loess and soil units with the Oxygen Isotype Stages in the core 552A (Deep Sea Drilling Project), after Shackleton et al. (1984), Ruddiman et al. (1986) and Kukla (1987). See overleaf for (c) and (d).Pleistocene climatic signals and chronologies. (c) Oxygen isotope record for the Camp Century ice core, north-western Greenland, after Dansgaard et al. (1971) and Johnsen et al. (1972). (d) Correlation between the Grand Pile continental deposit, north-eastern France and the deep-sea record, after Woillard (1978) and Woillard and Mook (1982).

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