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PALAEOLITHIC LANDSCAPES OF EUROPE AND ENVIRONS, 150,000–25,000 YEARS AGO: AN OVERVIEW

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Abstract — When considering the evolution and migrations of Neandertalers and early modern human beings, the harsh conditions of the last glacial maximum are often implicitly or explicitly assumed as their environmental background. This perception is false: the conditions of the high glacial apply to a small fraction of late Pleistocene time. Here we review the palaeoenvironmental history of Europe from 150,000 to 25,000 years ago with the aid of data from long cores of ice and marine and continental sediments. The results are displayed in four sketch maps that illustrate the landscapes of an interglacial–glacial cycle. The maps, connected by palaeoenvironmental histories, show that especially between 60,000 and 25,000 years ago, a critical part of the Palaeolithic, the glacial landscapes were for much of the time less barren than is generally assumed, but numerous climate changes on a scale of several millennia are evident, placing a premium on accurate dating of the co-evolution of humans and landscape. Moreover, during the glacial interval abrupt climatic changes lasting from a century to a few millennia were common. Their importance for landscape changes and their impact on human activity remain to be ascertained. Copyright © 1996 Elsevier Science Ltd

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INTRODUCTION

In the 20 years elapsed since CLIMAP Project Members (1976, 1981) produced the first quantitative global climate maps of the glacial world at its peak, much new information has become available, but the emphasis continues to be on the last glacial maximum and the subsequent deglaciation. This is a pity, because the full last glacial–interglacial–glacial cycle includes the arrival, evolution and global dispersal of modern human beings, our ancestors, and the demise of the Neandertalers. The ever-changing environmental setting against which this critical part of human history has played itself out is an essential factor in any attempt to understand that history.

This study is an experiment intended to assess to what degree our present knowledge suffices to create archaeologically useful images of the landscapes of Middle Palaeolithic Europe before the last glacial maximum. Our aim is archaeological, but without extensive treatment of the methods and the data base the Palaeolithic cannot be placed in the context of the changing European landscape. This treatment is presented here; an application to part of the Middle Palaeolithic is in progress.

The environmental database on which the reconstruction of Palaeolithic landscapes rests is still small and recent evidence for numerous brief but strong climatic oscillations is enough to give pause to even the most intrepid synthesizer. It is thus not surprising that, except for the final glacial maximum and deglaciation, only two broad syntheses, both covering the whole northern hemisphere but using very different methods, precede ours: a set of climate and vegetation maps ranging from the last interglacial (Eemian) to the Holocene by Frenzel *et al.* (1992) and two model-based reconstructions of the vegetation of the last interglacial by Harrison *et al.* (1995).

For this first stage in our experiment we have not used modelling and instead of compiling vast quantities of local information with their associated serious correlation problems, we have chosen to base our reconstructions on a small set of long records that each span several stratigraphic units of the late Pleistocene and so simplify (but not eliminate) many problems of chronology and correlation. We have further limited ourselves to a few broad variables: ice cover, sea-level, and the climatic and vegetation information contained in ice, ocean sediment and continental pollen cores.

We are very much aware of the speculative nature of our reconstructions, but believe that even the limited existing potential for spatial and temporal syntheses should no longer be ignored. What follows is an initial set of working hypotheses designed to lead to more advanced approaches. Ours is an historical science where, as Frodeman (1995) argued in similar terms, incomplete information produces tentative syntheses, which generate the inspiration for new observations that modify the existing syntheses as comprehension deepens in a circular fashion. It is in this spirit that we present our palaeoenvironmental hypotheses, hoping that these early, sparsely supported but coherent images of past worlds will accelerate our understanding of the glacial–interglacial cycle that looms so large in human history and will assist us by showing the best way to future, more solidly supported reconstructions.

FOUNDATIONS

The stratigraphy of the Pleistocene is burdened with a surfeit of local and regional divisions whose correlations with other regions are often weak or controversial. Here we have adopted as our basic time-scale the SPECMAP curve of Imbrie *et al.* (1984), a global oxygen isotope stratigraphy calibrated with the periodic changes of the earth's orbital geometry. Oxygen Isotope Stages (OIS) 6 to 3 inclusive cover the northern European sequence Saalian glacial–Eemian interglacial–Würmian glacial. As Fig. 1 shows, the global climate has fluctuated a great deal within each of the glacial and interglacial OI stages, mostly on a time-scale of several millennia, with obvious consequences for the activities of human beings. To consider the Palaeolithic against this backdrop of major environmental changes requires a matched high-resolution chronology good to at least within ±2–3 ka. This is difficult to achieve beyond the practical limit (40 kyr) of ^{14}C dating and currently nearly impossible beyond 100 or 120 kyr (Mellars *et al.*, 1993; Aitken *et al.*, 1993; Bar-Yosef and Kra, 1994).

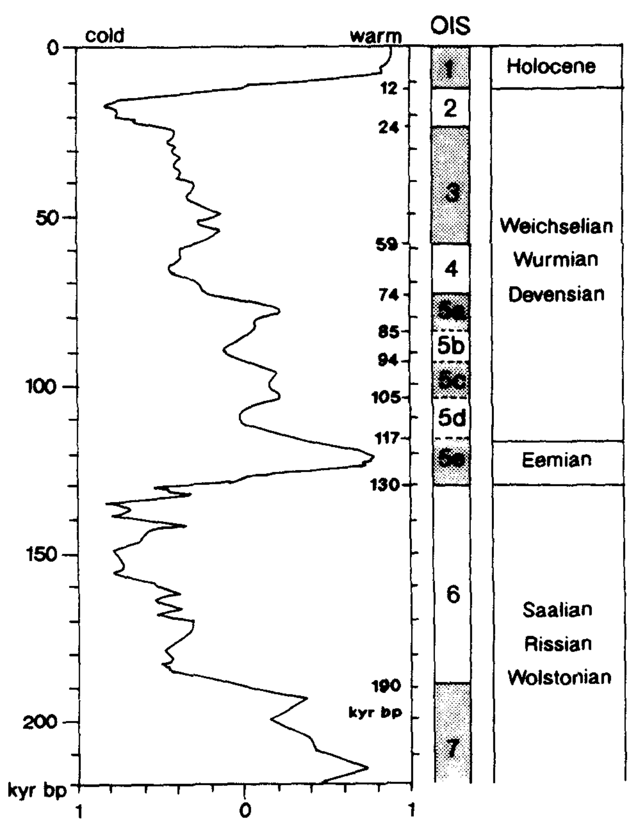

FIG. 1. Orbital-based global chronostratigraphy (SPECMAP) of the last 200,000 years; after Imbrie *et al.* (1984). Ages in millennia (kyr); Bottom: normalised benthonic $\delta^{18}O$. Oxygen isotope stage boundary ages (small numbers) after Martinson *et al.* (1987). The $\delta^{18}O$ curve can be regarded as a reasonable approximation of global climate oscillations between warm (interglacial or interstadial) and cold (glacial or stadial) states. Warmer intervals are shaded. A few stratigraphic names in common use are shown (but see Fig. 11).

[1]Ages are given in millennia (kyr) before present.

High-frequency Palaeoenvironmental Changes

Moreover, we can not be confident that this generalised curve reasonably reflects the climatic variations human beings have been exposed to in the past 150 ka. Recently, high-frequency climate oscillations on a centennial to millennial time-scale have been detected in Greenland ice cores that represent the last glacial period (Johnsen *et al.*, 1992; Dansgaard *et al.*, 1993; GRIP, 1993).

The ice cores display some 20 warm events between 20 and 105 kyr BP (Fig. 2), each ca. 7°C warmer than the intervening cold phases and only 2–10°C below the local Holocene average (Johnsen *et al.*, 1992; Dansgaard *et al.*, 1993; GRIP, 1993; Grootes *et al.*, 1993). Similar rapid oscillations have been detected in the sediments of the North Atlantic (Bond *et al.*, 1993; Keigwin *et al.*, 1994; McManus *et al.*, 1994) and North Pacific (Kotilainen and Shackleton, 1995; Thunell and Mortyn, 1995; Behl and Kennett, 1996). In the Antarctic, nine warm phases between 20 and 115 kyr BP seem to correlate with those Greenland events that lasted longer than 2000 years (Bender *et al.*, 1994). The observations suggest that at least some high-frequency climatic events of the northern hemisphere extended worldwide.

There is some evidence (GRIP, 1993; Johnsen *et al.*, 1995), as yet inconclusive (Peel, 1995), that similar brief but major climate instabilities may have marked the OIS 5e interglacial (Eemian) too. If true, this would contrast strikingly with the stable climate of the Holocene that, as

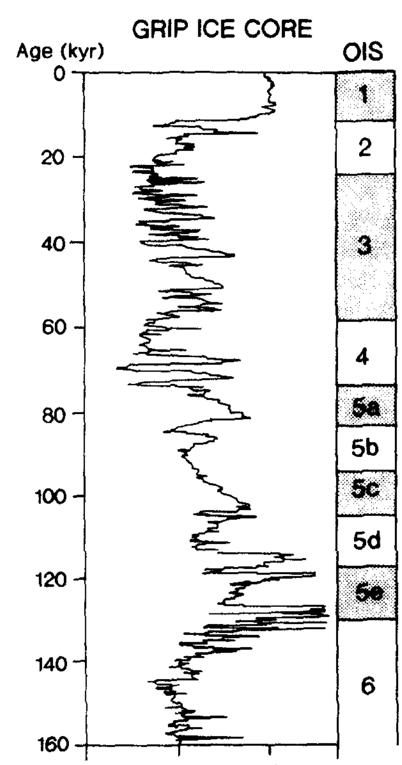

FIG. 2. Clusters of high-frequency climatic events recorded in a Greenland ice core (after GRIP, 1993: Fig. 1). The high-frequency events occurring in OIS 5e may in part be the result of ice flow deformation (Peel, 1995).

has been suggested, may have permitted the development of human civilisation.

An example, long known, of such a sharp oscillation is the Younger Dryas (11,000–10,500 BP[1]; e.g. Troelstra *et al.*, 1996) which, after the final deglaciation was well under way and the boreal forest had taken hold in northern Europe, brought the tundra back within a century. If these rapid climate changes were important in human terms—and why should they not be?—they may have been a major driving force in human biological and early cultural development. The study of human settlement and migration against a backdrop of climates and landscapes that varied on a scale of three to ten human generations would, however, become very difficult.

Waxing and Waning Ice-sheets, Falling and Rising Oceans

Because the temperature in the abyss changes little between glacial and interglacial conditions (Mix and Ruddiman, 1984; Shackleton, 1987), the $\delta^{18}O$ curve of bottom-dwelling Foraminifera is the one that best records the changing ice volume stored on land. The OIS 6 ice began to melt about 130 kyr ago (Fig. 1), rapidly reducing the large ice-sheets of the penultimate glacial to an interglacial volume that may have been a little smaller than that of today (CLIMAP Project Members, 1984). The OIS 5e interglacial *s.s.* was brief, however, and the following gradual climate deterioration raised ice volumes to an intermediate peak in OIS 4, followed by a long, partial withdrawal of the ice before the final maximum of OIS 2.

The benthic $\delta^{18}O$ curve informs us fairly well about the volume of water removed from the ocean and deposited as ice, but it is silent on the spatial limits of ice-sheets. For those we depend on the positions of terminal moraines and other glacial landforms, fragile evidence that is vulnerable to destruction by later ice advances. As a result most ice limits pre-dating the maximum of OIS 2 are to some degree uncertain unless they extended beyond the limits of OIS 2 ice-sheets.

All but the narrowest continental shelves form the flat surfaces of marine sediment wedges that, when exposed at low sea-level, form wide coastal plains, a landform rare at the high level of today. Although at no time in the last 150 kyr the sea fell enough to build major land bridges or reduce significantly the distance between, for example, Indonesia and Australia, wherever the often well-watered coastal plains emerged (Fig. 3) they offered important resources and migration paths to early humans (Shackleton *et al.*, 1984; van Andel, 1989a, b).

The global growth and decline of ice-sheets involved a volume of water equivalent to an ocean-wide layer about 165 m thick and so induced major sea-level changes. However, because the isotopic composition of ice-sheets varies with their size and latitudinal position (Mix and Ruddiman, 1984), the $\delta^{18}O$ ice volume curve displays only the general trend of changing levels of the sea (Fig. 4).

To obtain a true record of sea-level against time the $\delta^{18}O$ curve must be calibrated with past sea-level positions deduced from raised reefs or coastal terraces (e.g.

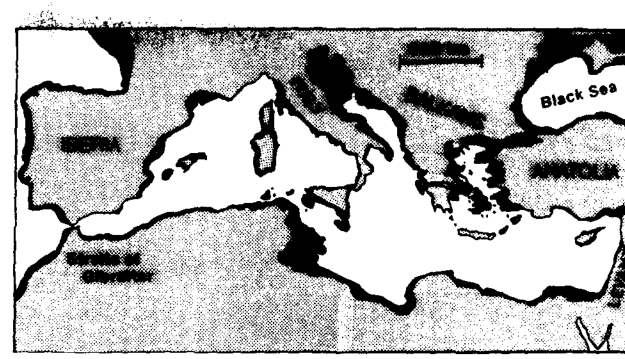

FIG. 3. The wide coastal plains of the Mediterranean (black) exposed during the glacial maximum of OIS 2 (after van Andel, 1989a: Fig. 3).

Bloom and Yonekura, 1985; Richards *et al.*, 1994) or observed as now submerged shore features on continental shelves (e.g. Pirazzoli and Pluet, 1991). The sea-level positions themselves require isostatic corrections to compensate for the weight of ice subtracted or added on land and of water added or subtracted on continental shelves and deep-sea floors (Lambeck, 1995). The glacio- and hydro-isostatic corrections may be positive or negative depending on the distance to the changed load and on the time elapsed since that load changed, and they may be large. Where applicable, adjustments must also be made for tectonic uplift or subsidence. It is not surprising that the best currently available time/depth sea-level curve (Fig. 4) only approximates the glacial–interglacial sea-level history, but it serves our purpose adequately.

Palaeoenvironmental Records

Aspects of the climatic changes of the last 150,000 years other than ice-sheets and sea-levels are documented by four major kinds of data. Oceanic sediments quantitatively record the properties of surface waters and the shallow and deep ocean circulation. Ice cores from Antarctica and Greenland reflect conditions at the ice surface as well as the history of atmospheric gases such as CO_2 and wind-borne dust. The ever-changing continental vegetation is displayed in long pollen sequences spanning several glacial–interglacial cycles and in the thick loess deposits of eastern Europe and Asia that provide manifold clues to the continental palaeoenvironmental history.

Our interpretation of the European landscape history rests mainly on long sediment cores (Fig. 5) which have yielded well-dated and correlated pollen records of the shifting vegetation patterns that accompanied glacial–interglacial cycles. The many tectonic and karst basins and volcanic lakes (maars) of central and mediterranean Europe have provided continuous pollen cores, but near the ice margins deposits may have been partly or wholly removed by ice advances or distorted by periglacial frost activity. From these fragmentary sections we have adopted only those that have a secure chronostratigraphy. Where available we have also used marine pollen sequences with good SPECMAP age control, sometimes supported by magnetic polarity reversals or volcanic ash beds.

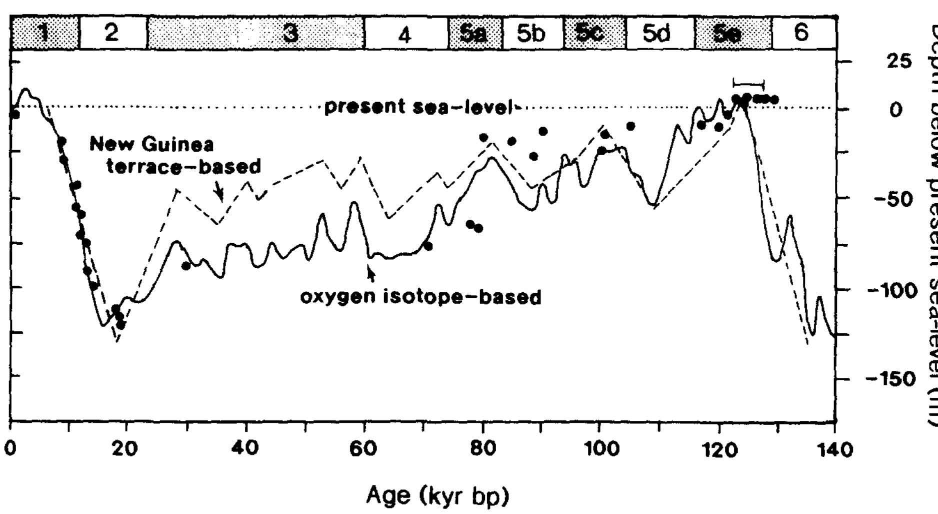

FIG. 4. Sea-level history for OIS 1–6. Solid line: sea-level derived from $^{18}O/^{16}O$ ratios of both planktonic and benthic Foraminifera (see Shackleton, 1987: Fig. 6). Dashed line: sea-level curve adjusted with dates and levels of raised New Guinea terraces (Chappell and Shackleton, 1986). Black dots: coral terraces dated with high-precision U/Th methods from Barbados (Bard *et al.*, 1990; Gallup *et al.*, 1994) and Bahamas (Chen *et al.*, 1991); bar above OIS 5e high sea-level covers the range of precision dates of Stirling *et al.* (1995).

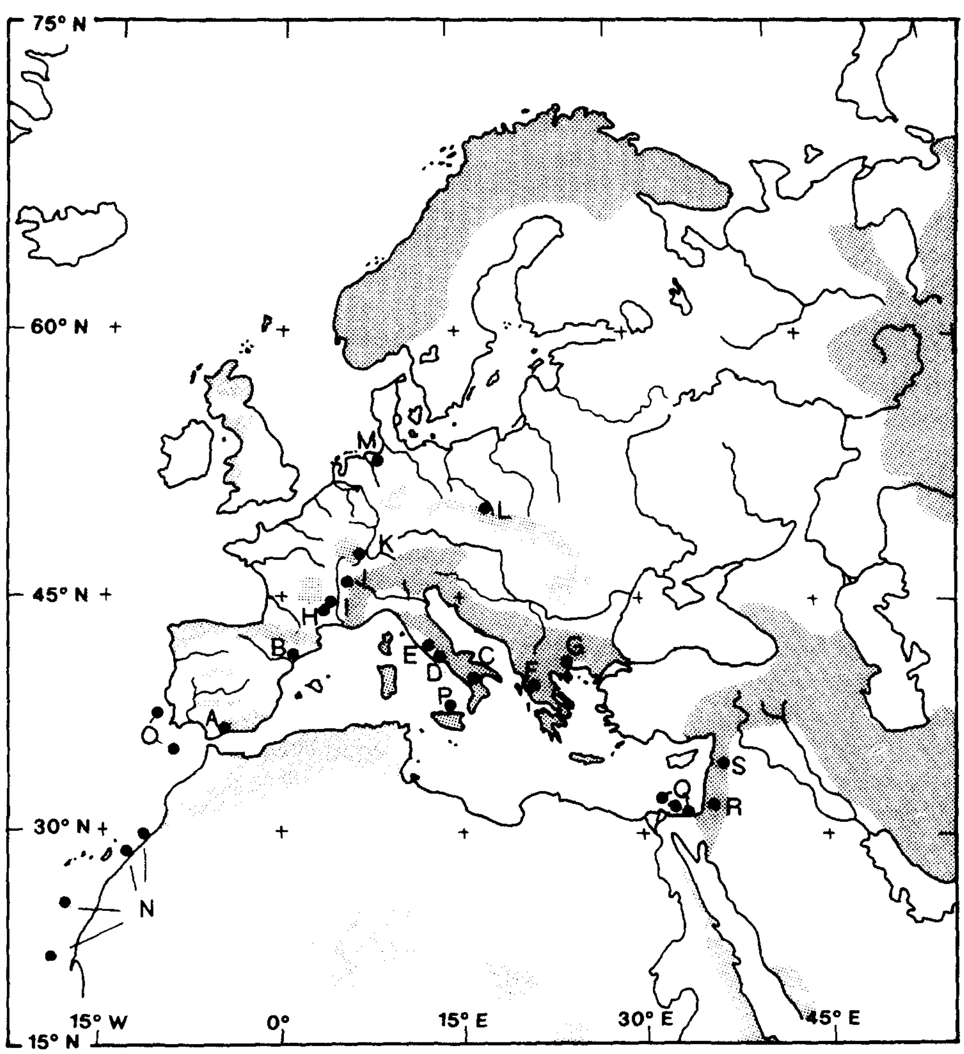

FIG. 5. Geographic base for Figs 6, 9, 13 and 14 with mountains and high plateaux (shaded) and major rivers. Main long records: **A.** *Padul* (Pons and Reille, 1988); **B.** *Banyoles* (Pérez-Obiol and Juliá, 1994); **C.** *Monticchio* (Watts, 1985); **D.** *Valle di Castiglione* (Follieri *et al.*, 1988); **E.** *Vico* (Leroy, 1994; Follieri *et al.*, in press); **F.** *Ioannina* (Tzedakis, 1994); **G.** *Tenaghi Philippon* (Wijmstra, 1969; Wijmstra and Smit, 1976); **H.** *Bouchet* (Reille and de Beaulieu, 1990); **I.** *Ribains* (de Beaulieu and Reille, 1992a); **J.** *Les Echets* (de Beaulieu and Reille, 1984); **K.** *La Grande Pile* (Woillard, 1978; de Beaulieu and Reille, 1992b); **L.** *Samerberg* (Grüger, 1979); **M.** *Oerel* (Behre and van der Plicht, 1992); **N** and **O.** *NW Africa* (Dupont, 1993 and Hooghiemstra *et al.*, 1992); **P.** *Tyrrhenian Sea* (Rossignol-Strick and Planchais, 1989); **Q.** *SE Mediterranean* (Cheddadi and Rossignol-Strick, 1995); **R.** *Ghab valley*, Syria; **S.** *Lake Huleh*, Israel (van Zeist and Bottema, 1991).

The quantitative interpretations from pollen records we present are based on the premise that climatic change is the primary determinant of the observed shifts in the temporal and geographic ranges and abundances of taxa (e.g. Webb, 1986, 1988; Huntley and Webb, 1989). The method has been criticised (e.g. Davis, 1981; Birks, 1981, 1986), because the vegetation development is also influenced by historical and biotic factors and so may lag behind climatic changes or be subject to internal forces that make pollen-based climatic reconstructions unreliable. While bearing this in mind, it is still useful to examine pollen-based climate reconstructions, such as that based on palaeobioclimatic analogues (Guiot, 1990; Guiot *et al.*, 1993), which has been applied to several European pollen sequences. This technique rests on identifying analogues of fossil pollen spectra among modern reference spectra. The taxa are then weighted by their response to climatic variables by means of eigenvector in order to emphasise those taxa with the most coherent behaviour in terms of vegetation dynamics. The estimates can be constrained by plant macrofossils and insect remains (e.g. Guiot *et al.*, 1993). Reconstructions of interglacial periods tend to be more robust than for glacial and late glacial times whose floristic and faunal assemblages often have no modern counterparts.

THE CHANGING LANDSCAPES OF EUROPE AND ITS ENVIRONS

Our synthesis has been summarised on four palaeoenvironmental sketch maps of Europe and environs connected by general palaeoenvironmental histories: (1) the penultimate glaciation (OIS 6) at 150 kyr BP, (2) the optimum of the last interglacial (OIS 5e) at 125 kyr BP, (3) the first major ice advance (OIS 4) at 65 kyr BP, and (4) a warm phase at 39–36 kyr BP in the long interstadial of OIS 3 that preceded the last glacial maximum. The maximum itself (OIS 2) and its deglaciation have been described fully elsewhere (see CLIMAP Project Members, 1976, 1984; Peterson *et al.*, 1979; COHMAP Members, 1988; Wright *et al.*, 1993) and are not discussed here.

The Late Penultimate Glacial: ca. 150 kyr

The northern ice-sheets of OIS 6 (the Saalian/Warthe complex) were among the most extensive of the whole Pleistocene (Fig. 6) and persisted much longer than those of the recent glacial maximum (OIS 2). No coherent data set exists for the North Atlantic Ocean at this time, but because the OIS 6 ice margin was on average no more than two or three degrees south of the OIS 2 ice limits, the Atlantic and Mediterranean conditions have been adapted from those prevailing during OIS 2.

For the Red Sea also we have assumed that OIS 6 conditions were similar to those of OIS 2. During the last glacial maximum sea surface temperatures dropped some 5°C below the present value (Ivanova, 1985). In addition, the global sea-level drop of more than 100 m sharply restricted the exchange of Red Sea water with the Gulf of Aden through the Straits of Bab el Mandeb (Rohling, 1994a). As a result evaporation in the Red Sea much exceeded the input of ocean water and the surface salinity in the central area reached 47‰ against 38–41‰ at present (Locke and Thunell, 1988).

A level of the sea much below that of OIS 2 would be the consequence of the greater ice extent of OIS 6, but the oxygen isotope curve (Fig. 4) does not strongly support this. Anyway, even if the sea had fallen to −140 or −150 m, the impact would have been small, because below 100–120 m the continental slope is steep. Therefore, we have assumed the same area of emerged continental margins as during OIS 2.

Long pollen sequences show that in Europe a moderately severe climate with fluctuating tree abundances prevailed in the early part of OIS 6, followed by more extreme conditions marked by a mainly treeless landscape (Fig. 6). At that time a polar desert existed south of the ice margin, while the rest of Europe was under a discontinuous, herbaceous plant cover.

North of the Alps this was a tundra-steppe dominated by grasses, sedges, chenopods and *Artemisia* (mugwort). No modern analogue exists for this mixture of tundra and steppe, but it probably represents a mosaic of separate communities that became integrated in the pollen record. During glacial periods the region north of about 45° N was subject to permafrost (van Vliet-Lanoë, 1989) and the lowlands, wet from the superficial summer thaw, may have supported tundra, while grass steppe covered the better drained uplands. Fossil beetle assemblages from the late penultimate glacial at Grande Pile (Vosges) correspond to modern tundra faunas with a few temperate species (Ponel, 1995).

South of the Alps a discontinuous steppe vegetation with *Artemisia*, chenopods (goosefoot family, indicative of aridity) and grasses predominated. Detailed studies of cold-stage chenopod pollen from Tenaghi Philippon in Macedonia, Greece (Smit and Wijmstra, 1970), showed the presence of *Eurotia ceratoides* and *Kochia laniflora* found today in steppe and semi-desert environments in central Asia, leading the authors to infer a cold, arid climate. In sheltered areas, mainly in the western Balkans and the mountains of Italy, scattered temperate tree populations survived in refugia where temperature variations were not extreme and precipitation was sufficient (Bennett *et al.*, 1991; Tzedakis, 1993). In the Iberian peninsula woodland populations occurred mainly in the south, probably because of aridity elsewhere. A semi-desert vegetation covered the Levant, but the Taurus mountains may have provided refugia for warmth-loving trees (Cheddadi and Rossignol-Strick, 1995).

Although the temperate tree refugia in southern Europe contained a mixture of various species, possibly one without modern analogues, a latitudinal floristic differentiation can be seen. Coniferous populations occurred mainly in the northern Balkans, northern Italy and some in northeast Spain, while deciduous trees were found further south with evergreens in the extreme south and coastal lowlands.

Long pollen sequences from France have yielded esti-

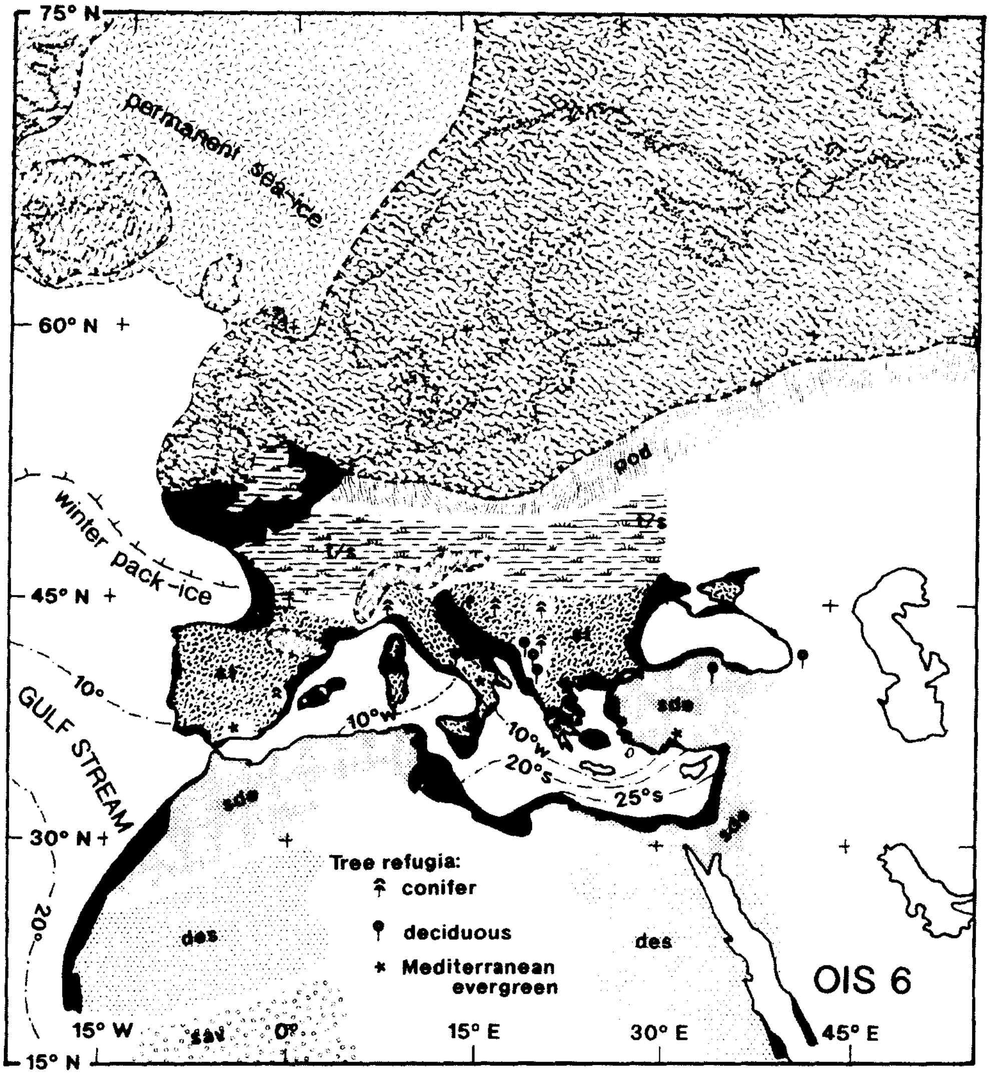

FIG. 6. Europe in the penultimate glacial (OIS 6) at 150 kyr BP. Ice margin from Donner (1995, Fig. 3.2), Nilsson (1983, Fig. 12.3) and P. Gibbard (*pers. commun.*, 1995). Pack-ice limits and ocean surface isotherms from CLIMAP Project Members (1984). Mediterranean surface isotherms adapted from OIS 2 data (Thiede, 1978, 1980; Thunell, 1979; Thunell and Williams, 1983) and Gulfstream position after Keffer *et al.* (1988). Coasts based on −100 m isobath. Cloudy pattern: ice-sheets; black: emerged coastal plains. Palaeoenvironments: pod: polar desert; t/s: tundra and cold steppe mosaic; st: arid cold steppe; med: Mediterranean evergreen woodland; sde: semi-desert; des: desert.

mates of annual temperature and precipitation (Guiot *et al.*, 1989, 1993). At La Grande Pile (Vosges) the annual temperature was 1–2°C (now 9.5°C) with a July average of 10–12°C and 300 mm of precipitation compared to 1080 mm today. In south-central France the Les Echets area had an annual temperature of 1.5°C compared to 9.5°C today and a precipitation of ca. 200 mm (830 mm today). These values, even the warm summer temperature, are reasonable, because the mid-latitude sites enjoyed a high midday sun and a long summer.

In north-western Africa land data are scarce, but insight in the vegetation zones of the past comes from oceanic sediment cores (Lézine and Casanova, 1991; Hooghiemstra *et al.*, 1992; Dupont, 1992, 1993). The cores are located off western Africa and between Portugal and Morocco (Fig. 5) and provide a long, almost uninterrupted record of pollen wind-borne from North Africa (Fig. 7). Early in OIS 6 encroachment of the sagebrush steppe caused the Mediterranean oak woodland to disappear, leaving only scattered populations surviving in favourable areas in the mountains. Northward movement of the semi-desert that fringed the Sahara desert in the north came a little later (Hooghiemstra *et al.*, 1992) and the desert itself widened, pushing the savanna and tropical rainforest southward.

The temporal shifts of the zones of vegetation in northwestern Africa were due mainly to southward migration of the dry subtropical high pressure zone during glacials (Dupont and Beug, 1991; Hooghiemstra *et al.*, 1992; Dupont, 1993; Frédoux, 1994). In north-eastern Africa and south-western Asia, on the other hand, precipitation and hence the vegetation were influenced mainly by the response of the monsoon to the Milankovitch precession cycle (Kutzbach and Street-Perrott, 1985; Kutzbach, 1987; Kutzbach and Gallimore, 1988).

During interglacials, when the northern summer coincides with the earth's closest proximity to the sun, the temperature contrast between the Indian Ocean and continental interiors is greatest. The monsoon is then strongest and most extensive and affects northeast African and

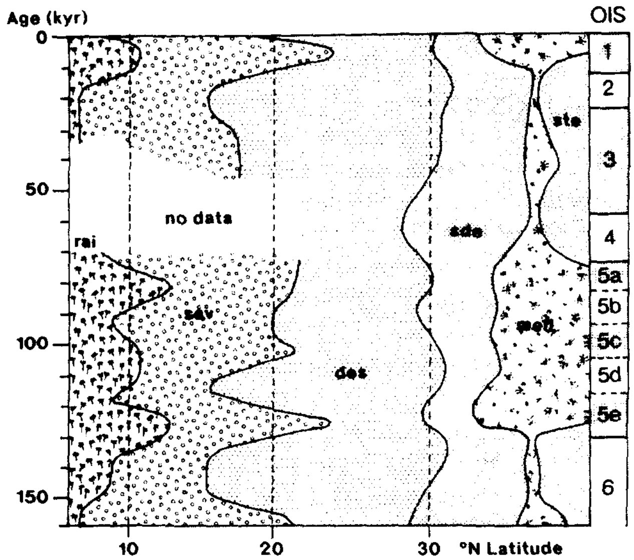

FIG. 7. Vegetation history of northwestern Africa during the last 150 kyr based on pollen in offshore cores. Modified after Dupont (1993, Fig. 4a); locations on Fig. 5. Palaeoenvironments: med: Mediterranean woodland; sde: semi-desert; des: desert; sav: Sahelian savanna or parkland; rai: African rainforest.

southern Asian climates by its north-westward expansion (van Campo *et al.*, 1982; Rossignol-Strick, 1983; Prell, 1984; Pokras and Mix, 1985; Prell and van Campo, 1986; Prell and Kutzbach, 1987). During glacial periods such as OIS 6 these regions were but little affected by the monsoon and like northwestern Africa had a dry climate caused by the southward shift of the subtropical dry zone.

The Last Interglacial: 130–117 kyr

The last interglacial (OIS 5e) is usually thought of as warmer than its Holocene counterpart, with a reduced ice cover and a sea-level higher than today by 2–12 m. How well is this view supported by our current knowledge?

After the rapid deglaciation that terminated OIS 6 the interglacial quickly turned warm for about 10,000 years, before another slow gradual decline (Fig. 8). OIS 5e ocean surface temperatures differed little from those of today (CLIMAP Project Members, 1984) and the few warm anomalies are generally within error limits, but warmer surface waters may have briefly existed in the Norwegian Sea (Duplessy *et al.*, 1988).

High interglacial sea-level stands ranging in age from 140 to 122 kyr (Fig. 8) have been reported for many stable continental coasts and oceanic islands (Kaufman, 1986; Edwards *et al.*, 1987; Bard *et al.*, 1990; Ku *et al.*, 1990; Chen *et al.*, 1991; Gallup *et al.*, 1994). The earliest dates imply that the last interglacial might have begun as much as 10 kyr before the insolation maximum at 125–126 kyr BP and so contradict the SPECMAP dates (Smart and Richards, 1992; Crowley, 1994; Crowley and Kim, 1994; Gallup *et al.*, 1994), but new high-precision U/Th dates obtained by Stirling *et al.* (1995) place the start of the high stand of the sea no earlier than 130 kyr BP.

These same high sea-levels have been cited as evidence for an ice volume reduced well below its present value through the melting of parts of the Canadian Arctic, Greenland and West Antarctic ice-sheets (CLIMAP Project Members, 1984, pp. 205–212; Funder,

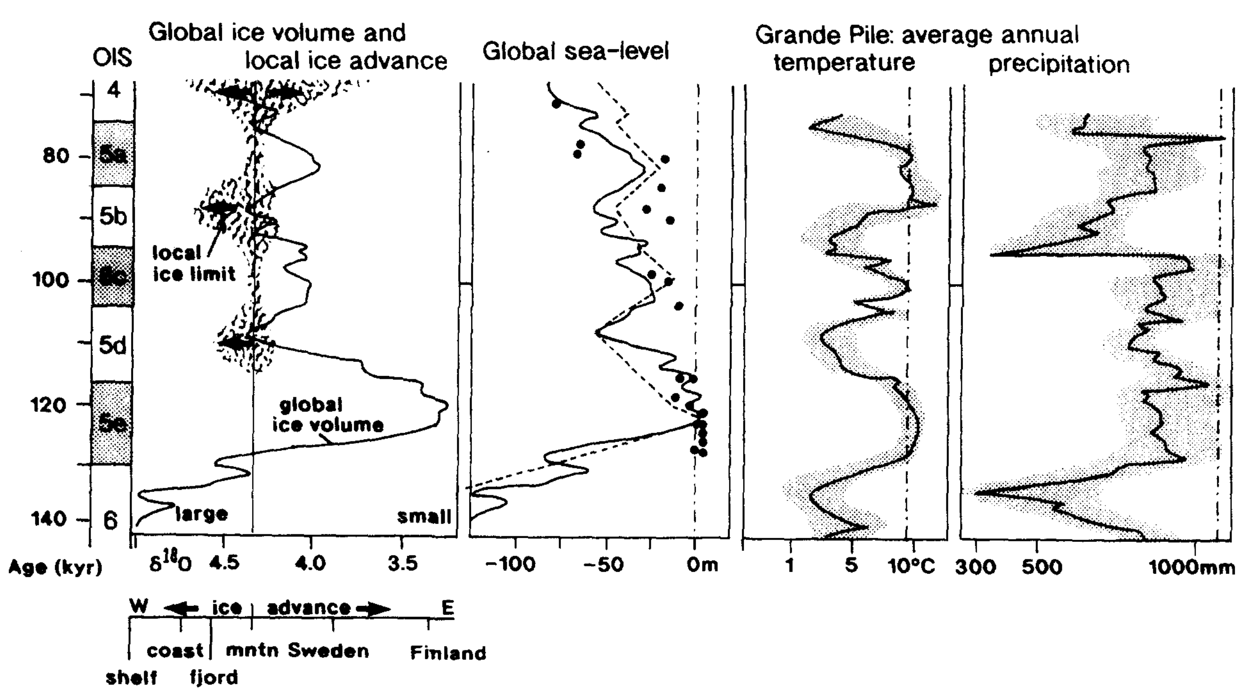

FIG. 8. Palaeoenvironmental changes from 130 kyr (OIS 6 deglaciation) to the approach of the last pleniglacial (OIS 6 to OIS 4). Global ice volume from Shackleton (1987, Fig. 1); Scandinavian ice advances (shaded) after Mangerud (1991a, Fig. 5); diagram at lower left relates ice edge to major geographic features. Sea-level changes from Fig. 4; black dots are coral dates of past sea levels from Barbados (Bard *et al.*, 1990; Gallup *et al.*, 1994) and the Bahamas (Chen *et al.*, 1991). Central European mean annual temperature and precipitation after Guiot *et al.* (1989, Fig. 3): confidence intervals shaded. Dashed vertical lines mark present values.

1989; Koerner, 1989). However, Lambeck and Nakada (1992) have argued that isostatic compensation could produce the same elevations for an interglacial ice volume and sea-level equivalent to those of today. Their conclusion was confirmed by Stirling *et al.* (1995, Fig. 4) who also showed that the brevity of the period of high sea-levels does not invalidate the SPECMAP interglacial dates (130–117 kyr), because a reasonable, slightly different mantle relaxation could have extended its duration (Stirling *et al.*, 1995, Fig. 5).

With sea-levels about the same as now, we may equate most OIS 5e shorelines with present coasts except where deposition was especially rapid. A striking exception are the coasts of the Eemian Sea in northern Europe (Fig. 9), which differ greatly from those of the present North Sea and Baltic (Mangerud *et al.*, 1979, 1981; Nilsson, 1983, pp. 205–212); Miller and Mangerud, 1986; Mangerud, 1989). The rate of sea-level rise that created the Eemian Sea was ca. 20 m/kyr (Zagwijn, 1983; Streif, 1991), twice that of the recent postglacial rise, indicating exceptionally rapid deglaciation at the OIS 6/5e boundary.

The Eemian Sea was at its highest level from about 126 to 116 kyr BP (Fig. 8; Streif, 1991) and had regressed before 110 BP. Along the North Sea coast it flooded only main river valleys, but it occupied the entire Baltic basin, connecting the North Sea with the White Sea and the Arctic Ocean and turning Fennoscandia into an island (Zans, 1936; Gross, 1967; Forsström *et al.*, 1988). Its western waters contained warm faunas related to more southerly waters warmer than the present North Sea, while colder faunal elements in the north-east suggest connections with the Arctic Ocean. In the North Sea region itself summer temperatures rose from 10°C at the start of the interglacial to 22°C at its peak, well above present values (Zagwijn, 1961, 1983; Müller, 1974; Larsen *et al.*, 1995). The shores of this remarkable transcontinental sea are likely to have had a variety of environments and resources that would have invited human exploitation even at this early date.

High temperatures have also been claimed for the Mediterranean. Planktonic Foraminifera from a marine core east of Crete indicate that warming began 127 kyr

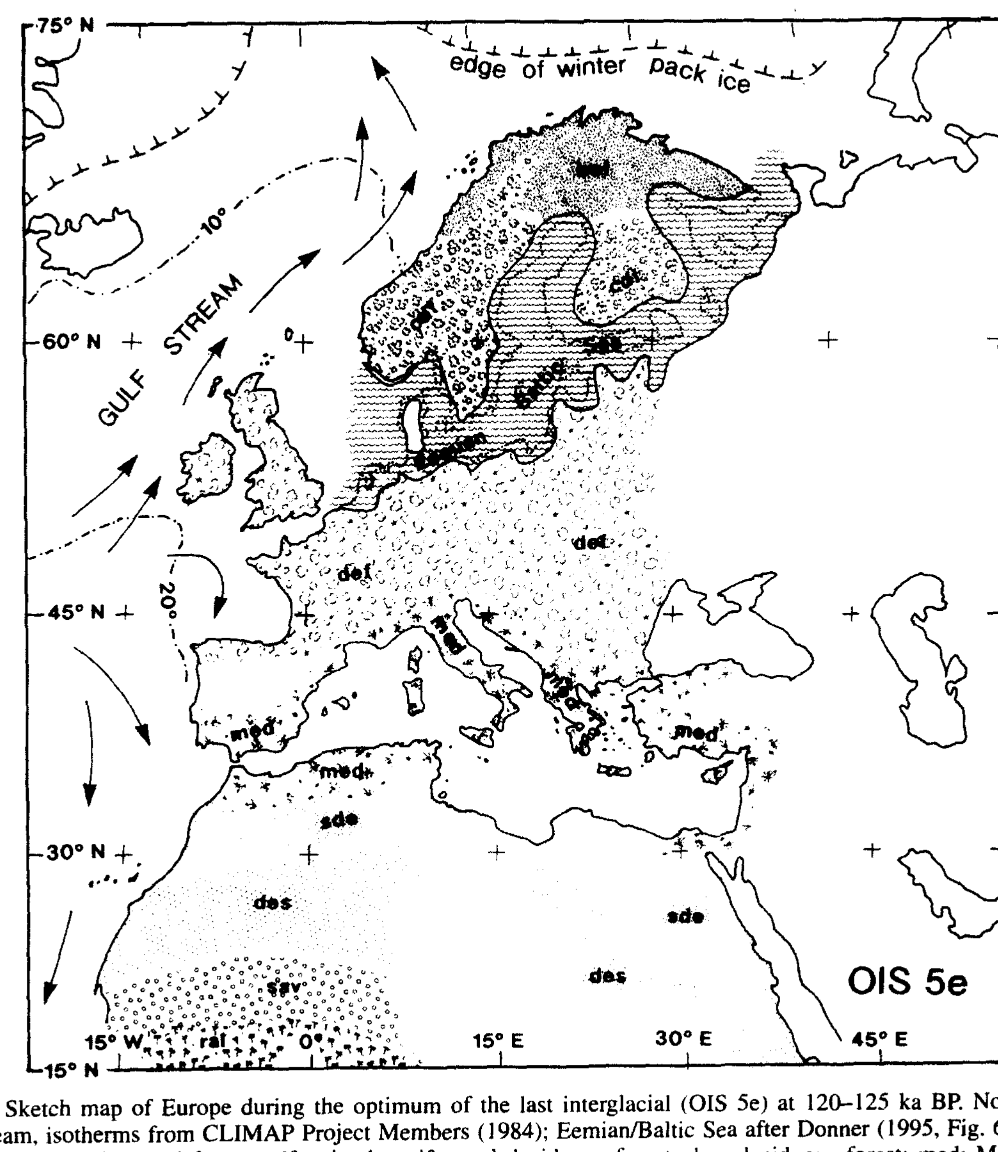

FIG. 9. Sketch map of Europe during the optimum of the last interglacial (OIS 5e) at 120–125 ka BP. North Atlantic Gulfstream, isotherms from CLIMAP Project Members (1984); Eemian/Baltic Sea after Donner (1995, Fig. 6.1). Palaeoenvironments: bof: boreal forest; cdf: mixed conifer and deciduous forest; dec: deciduous forest; med: Mediterranean evergreen woodland; sde: semi-desert; des: desert; sav: Sahelian savanna or parkland; rai: African rainforest.

ago and that the temperature rapidly rose to ca. 3°C above the present value (Thunell and Williams, 1983). In the western Mediterranean, 3°C warmer water is implied by $\delta^{18}O$ values of *Strombus bubonius* in Tyrrhenian raised beaches (Cornu *et al.*, 1993), a species today not found north of the Gulf of Guinea. Unfortunately, the temperature estimates are not very robust, because they depend on surface salinity values (Thunell and Williams, 1983, 1989). In the Mediterranean the OIS 6/5e deglaciation was accompanied by a large ^{18}O depletion, probably because of the combined effect of meltwater input from the Black Sea (Thunell and Williams, 1983), increased Nile discharge caused by an enhanced monsoon (Rossignol-Strick, 1983) and perhaps increased summer rains in Anatolia and the Near East (Rohling, 1994b; Rohling and Hilgen, 1991). In contrast, the surface salinity and temperature of the Red Sea appear to have been similar to those of today (Ivanova, 1985).

Turning to north-west Africa, the interglacial vegetation zone boundaries (Fig. 7) differed markedly from those of OIS 6 (Dupont, 1993; Hooghiemstra *et al.*, 1992). Mediterranean scrubs and oak woodland covered the region from the coast to the southern slopes of the Atlas where a narrow transitional zone (semi-desert) graded into true Saharan desert. The desert itself was much reduced due to the northward shift of the Sahelian savanna which in turn was encroached upon by a northward expansion of the tropical rainforest. Farther east, in southern Tunisia, high lake stands with U/Th ages from 136–191 kyr were assigned by Causse *et al.* (1989) to a wet phase at 150 kyr "corresponding closely to the last interglacial" (sic!), but the ages belong to OIS 6. Besides, Mollusca are notorious for yielding poor U/Th dates (Kaufman, 1986).

In northeast Africa and southwest Asia many Pleistocene lake deposits indicate intermittently increased precipitation. In the western desert of Egypt. Some of the lakes have been assigned to OIS 5e on the basis of U/Th dates of lake marls (Wendorf *et al.*, 1994, p. 165; Szabo *et al.*, 1995), but others range to 155 kyr BP, thus including the penultimate glacial (Szabo *et al.*, 1995).

When the interglacial began in Europe, trees spread outward from refugial centres, diachronically colonising the hitherto open landscape as different populations successively became dominant under the influence of the changing climate, biotic interactions and soil maturation.

Sites across the whole of Europe display an interglacial vegetation succession (Fig. 10) which began with deciduous *Quercus* (oak) and *Ulmus* (elm). Afterwards, *Corylus* (hazel) and then *Taxus* (yew) became prominent north of the Alps (Fig. 10). They were followed by a *Carpinus* (hornbeam) forest that spread across the whole of Europe, soon accompanied by expansion of *Abies* (fir), then *Picea* (spruce) and *Pinus* (pine). The *Carpinus* expansion into areas far beyond its present geographical range and the almost total absence of *Fagus* (beech) occur across Europe (Tzedakis, 1994), but it is not clear whether this was due to climate or to competitive interac-

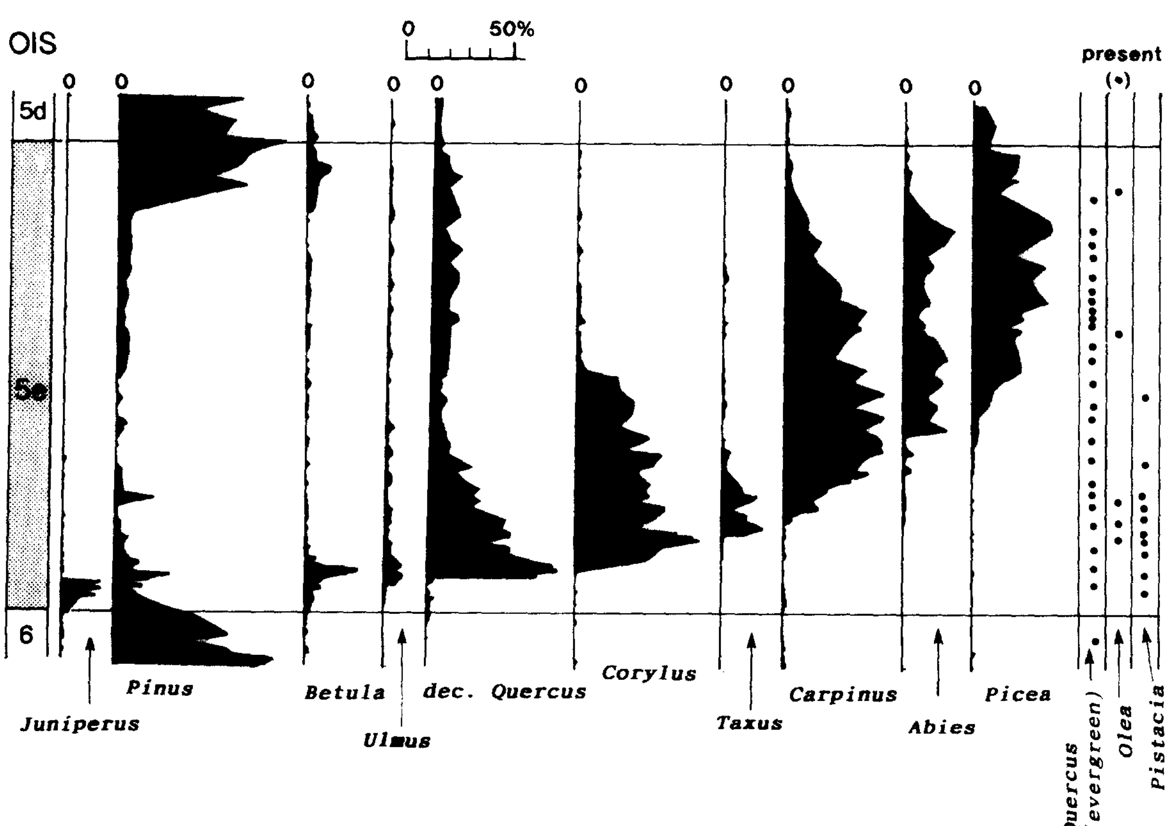

FIG. 10. Forest succession in western Europe during the last interglacial (OIS 5e) as shown by selected taxa from the Ribains pollen core (de Beaulieu and Reille, 1992a: Fig. 2). Dots or blanks in three right hand columns indicate presence or absence.

tions. Might, for example, the absence of *Fagus* have permitted the expansion of *Carpinus* (Huntley and Birks, 1983)?

Towards the end of OIS 5e, a brief double peak of fir or spruce interrupted by *Pinus* peaks has been recorded at some sites that may point to some climatic instability at the end of the Eemian (Tzedakis *et al.*, 1994).

In southern Europe the *Quercus/Ulmus* interval was followed by a major expansion of Mediterranean forest characterised by *Olea* (wild olive) and evergreen oak. In the circum-Mediterranean region olive reached values higher than even in the Holocene (Tzedakis, 1994), and marine cores in the eastern Mediterranean (Cheddadi and Rossignol-Strick, 1995) show that the olive event was contemporaneous with the deposition of a marine deposit dated at 126–125 kyr (sapropel S5: Muerdter *et al.*, 1984). This was the time of maximum summer insolation (12–13% above the present value) in the last interglacial and indeed the last 150 kyr (Berger, 1978).

The Mediterranean olive event appears to have been synchronous with the *Taxus* phase in the rest of Europe, because olive (and also *Pistacia* (pistacio)) and evergreen oak pollen grains appear in that phase at Ribains in the Massif Central (de Beaulieu and Reille, 1992a), conveniently linking the stratigraphies north and south of the Alps.

Palynological and $\delta^{18}O$ data from a marine core west of Portugal (Turon, 1984) show that temperatures in western Europe remained high to the end of the *Carpinus* expansion and throughout the *Abies/Picea/Pinus* phase. Still, ice-caps must have begun to form somewhere, because late in OIS 5e (Fig. 8) the level of the Eemian Sea in The Netherlands began to drop (Zagwijn, 1983). The culprit can only have been the North American ice-sheet, because the forest that was present at this time at Fjøsanger in western Norway (Mangerud *et al.*, 1981; Mangerud, 1989) makes the existence of a Scandinavian ice-sheet unlikely (Mangerud, 1991a; see also Cortijo *et al.*, 1994). This raises the question whether the climate changes reflected in European records resulted only from external radiative forcing or partly from the internal dynamics of the climate system. A more accurate terrestrial chronology is needed before this matter can be pursued further.

Palaeobotanical records with high counts of the pollen of warmth-loving taxa and of macroscopic plant remains well beyond their Holocene ranges provide ample evidence that the **continental** climate at the peak of Eemian was warmer than at any time during the Holocene. Mediterranean shrubs like *Acer monspessulanum* (Montpelier maple) and *Pyracantha coccinea* (firethorn), together with the common occurrence of *Hedera helix* (ivy), *Ilex* (holly), *Mariscus serratus* and plants now absent in Britain and the Low Countries such as *Trapa natans* (water chestnut), *Najas minor* (Lesser Najad) and *Salvinia natans*, point to warm, dry summers and mild winters (Phillips, 1974).

In Ostrobothnia the OIS 5e optimum was marked by warmth-loving species (Forsström *et al.*, 1988) beyond their northern Holocene limits, and a flora and fauna more temperate than in the Holocene characterised Eemian deposits in Swedish Lapland at latitude 67°38'N (Lundqvist, 1971). Pollen analysis of these same deposits suggest the proximity of *Corylus* and *Alnus* (alder) populations. Plant macrofossils and beetle remains indicate a mean annual temperature of 4°C above that of the Holocene.

In contrast, farther south a climatic reconstruction (Fig. 8) based on beetle remains from the *Taxus* phase at La Grande Pile (Vosges; Ponel, 1995), the time of maximum summer heat, yielded a mean annual temperature of 10–12°C, not far above the present 9.5°C. The mean July temperature (16–18°C: Guiot *et al.*, 1993) was also unexceptional and so was the temperature of the Atlantic ocean (CLIMAP Project Members, 1984). Do we see here climatic differences between northern and southern latitudes and/or between maritime and continental regions, or are we merely looking at wide error limits of the estimates?

A climate and vegetation model where the solar radiative forcing of the OIS 5e interglacial maximum was prescribed by means of an atmospheric general circulation model helps here (Harrison *et al.*, 1995). When, due to the combined effects of tilt, precession and eccentricity (Berger, 1978), insolation reached a maximum at 125 kyr BP, summer temperatures higher than today by 4°C prevailed especially in middle and high latitudes, while somewhat greater winter cooling in lower latitudes and increased seasonal contrasts marked the mid-continental regions of Eurasia. In the Arctic, simulated winter temperatures were 2–8°C higher than now, especially in northern Scandinavia, with large annual and seasonal contractions of sea-ice.

Although the model did not take into account the Eemian Sea and its connection with the Arctic nor any ocean/atmosphere interaction, the results agree with the data summarised here encouraging the use of carefully designed simulations of this kind.

Approaching The Last Glacial: 74–116 kyr

This interval (OIS 5a–d; Fig. 8) comprises the slow climate deterioration that led to the first major European ice advance in OIS 4. Sea-level, which had started to fall late in OIS 5e, reached –50 m towards the end of OIS 5d (Herning or Melisey I stadial). Although the depth is not confirmed by independent data, it suggests a sizeable global ice volume (Fig. 8), but the Scandinavian ice-sheet was not yet extensive (Mangerud *et al.*, 1981; Mangerud, 1991a, b) and the lack of ice-rafted deposits in the Norwegian Sea (Baumann *et al.*, 1995) confirms that the ice front had not reached the coast of Norway. East of the Scandinavian mountains Lapland and northern Finland may have been ice-covered, but elsewhere the ice limit was probably located well up on the mountain slopes (Mangerud, 1991a).

The next interstadial (OIS 5c: Brørup, St Germain I) was quite warm and a large part of the Scandinavian ice-sheet may have vanished (Mangerud, 1991a; data in Donner, 1995, Chapter 8) as is also suggested by a U/Th dated sea-level at –20 m (Pickett *et al.*, 1985; Bard *et al.*, 1990; Smart and Richards, 1992). The vegetation reflects conditions like those of the late Eemian.

Sea-level dropped again in OIS 5b (Rederstall or Melisey II stadial), but the –25 m fall, now backed by coral dates, implies only a modest global ice volume (Fig. 8). In Norway the ice edge locally advanced beyond the west coast (Mangerud, 1991a) and cores from the Norwegian Sea provide evidence for meltwater and icebergs, but ice-rafted material remained sparse (Baumann *et al.*, 1995). Little is known about the regions east of the Scandinavian mountains and south of the Baltic, but conditions there probably resembled those of OIS 5d.

During OIS 5a (Odderade or St Germain II interstadial), the sea rose again but not quite to the level of OIS 5c and the ice volume curve confirms a gradual global increase beyond that of the previous interstadial, although there is no direct evidence for an advance of the Fennoscandian ice-sheet (Mangerud, 1991a).

In pollen sequences the climate oscillations during OIS 5d–a appear as alternations between expanding open vegetation and returns of forest conditions. During interstadials, the north–south forest composition across Europe had a steeper gradient than at the present time. The tundra of northwest Scandinavia was replaced by birch forest in southern Scandinavia. Conifer forests with *Pinus*, *Picea*, *Abies*, *Larix* (larch) and *Betula* (birch) covered northern Germany, Denmark, the Netherlands and Britain, while a mixed forest with warmth-loving deciduous trees expanded in south central Europe and France (Behre, 1989; Emontspohl, 1995). South of the Alps, a deciduous forest with Mediterranean elements existed (Wijmstra, 1969; Pons and Reille, 1988; Tzedakis, 1994; Follieri *et al.*, *in press*). According to estimates of temperature and precipitation (Fig. 8) the stadials were cold and dry and the interstadials more continental than those of OIS 5e.

In the eastern Mediterranean the cold, dry chenopod and *Artemisia* steppe alternated in a similar fashion with returns of the Mediterranean mixed evergreen and deciduous woodland. The interstadial landscapes were more open than in OIS 5e, however, and semi-desert and desert plant communities were present even in warmer phases (Cheddadi and Rossignol-Strick, 1995).

The pollen record of La Grande Pile (Vosges) has been correlated with the marine oxygen isotope stratigraphy by Woillard (1978; Woillard and Mook, 1982), a correlation now also achieved in southern and northern Europe. In each interstadial the tree expansion was rapid in all regions, implying that their refugia were relatively close, although stadial conditions were more severe in Melisey II than Melisey I. In southern Europe this decline was marked by an increased aridity indicated by major expansion of chenopods and *Artemisia* (Follieri *et al.*, *in press*).

Early in the St Germain I interstadial the development of a warmer forest was interrupted briefly by the cold Montaigu event (Reille *et al.*, 1992), when a cold steppe expanded for 500–1000 years until the forest re-established itself (de Beaulieu and Reille, 1992a). This event, similar to the Younger Dryas, has now been recognised in all sequences in France and southern Europe.

OIS 5a–d conditions in northern Africa are documented only for the northwest (Fig. 7); elsewhere the chronology is too insecure. Vegetation differences between interstadials and stadials were small (Dupont, 1993) except for some shifts of the boundary between the desert and the Sahelian savanna. The Mediterranean forest was somewhat impoverished (Hooghiemstra *et al.*, 1992) even in the interstadials (OIS 5a, 5c), and the semi-desert extended slightly farther north.

The First Major Ice Advance: 74–59 kyr

The long, slow descent into the full glacial culminated in a large increase of the global ice volume during OIS 4 (Fig. 12). The Scandinavian ice-sheet advanced considerably, but its marginal deposits, over-ridden by OIS 2 ice-sheets, have been much obscured by later erosion and/or burial. Glacial deposits thought to belong to OIS 4 exist not only in Fennoscandia, but also in Denmark (Strand Strand-Petersen and Kronborg, 1991), Poland (Mojski, 1991), Estonia (Liivrand (1991) and northwest Germany (Benda, 1995, Chapter 8, pp. 106–107; Menke, 1991), but they are highly controversial. Unfortunately, OIS 4 is beyond the safe limit of conventional ^{14}C dates and the distinction between OIS 4 and OIS 2 glacial deposits based on age is thus insecure (Donner, 1995, pp. 70–73, Table 9.1). Thermoluminescence dating has been attempted (e.g. Strand-Petersen and Kronborg, 1991), but the results, such as dates placing OIS 4 15–20 kyr before its SPECMAP age (Mojski, 1991) justify great caution. As a result, Andersen and Mangerud (1989, Fig. 9) gave two ice limits for OIS 4 (Fig. 13) and Donner (1995, Fig. 8.4) omitted trans-Baltic lobes altogether. It is clear, however, that the OIS 4 ice sheet was much smaller than that of the last glacial maximum (OIS 2).

If there ever was an early advance of the Alpine ice-caps corresponding to that of Fennoscandia (Benda, 1995, p. 268; de Beaulieu *et al.*, 1991; Frenzel, 1991; Schlüchter, 1991), it was too localised to affect Fig. 13. In Scotland, a small OIS 4 ice-cap or glacier deposit exists underneath OIS 3 interstadial beds (Sutherland and Gordon, 1993), but England was probably ice-free until 25 kyr ago (Ehlers *et al.*, 1991).

Instead of an early establishment of glacial conditions and their persistence throughout OIS 4, the vegetation record responded to the expansion of the Scandinavian ice-sheet by a complex series of progressive changes. A brief time of open vegetation at the OIS 5a/4 transition was followed in France and southern Germany by the Ognon warm phase (Fig. 11) characterised by renewed expansion of conifers (Woillard, 1978) while south of the Alps *Juniperus* (juniper) also increased. Only towards the end of OIS 4 extensive open vegetation appeared in response to the lower temperatures resulting from the expansion of the Fennoscandian ice-sheet (Figs 12, 13 and 14). Farther south, tundra and cold-arid steppe expanded and warmth-loving tree populations contracted, even in refugia such as those of north-west Greece (Tzedakis, 1993). Pollen-based climatic reconstructions

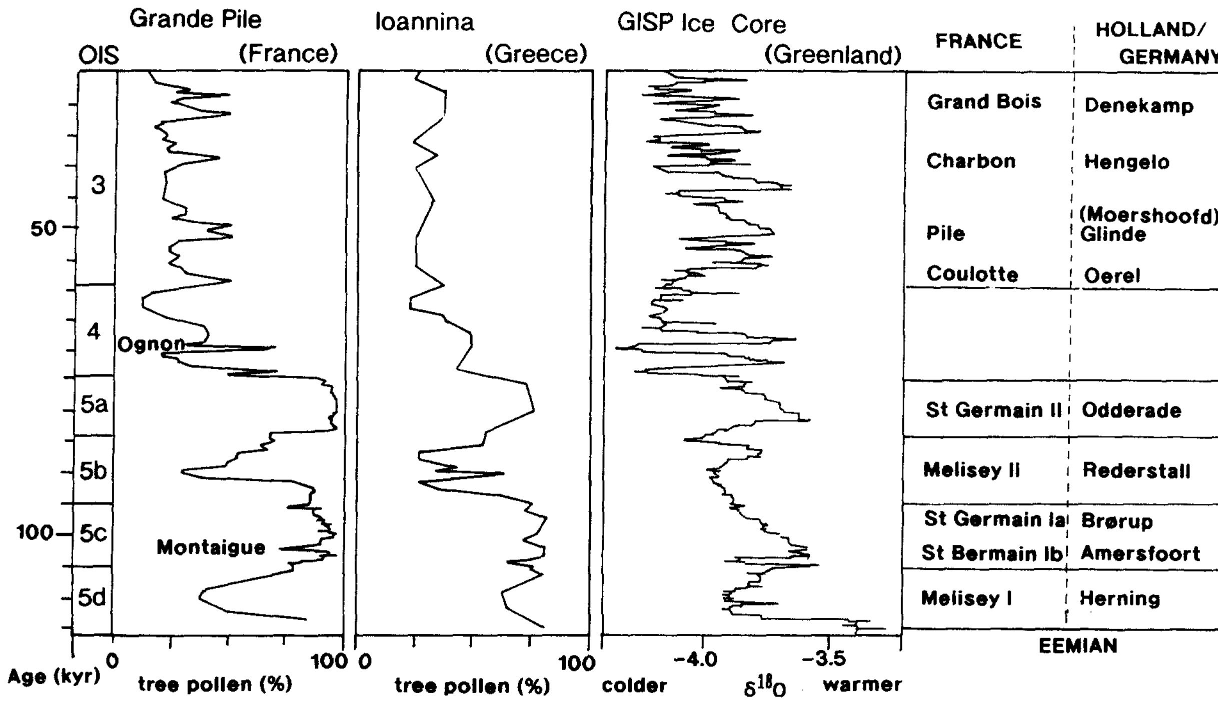

FIG. 11. Late Quaternary European vegetation history based on total tree pollen percentages in France and NW Greece, compared to high-frequency climate oscillations of the GRIP ice core (GRIP, 1993, Fig. 1) during OIS 3–5d. Current status of fine-scale European stratigraphy is shown. NOTE: due to scale and dating problems correlations between the three records are approximate.

from France (Guiot *et al.*, 1989) indicate annual temperatures 12–13°C lower and precipitation 650–800 mm less than at present (Fig. 12).

Marine pollen cores show, although with reduced resolution nearly the same sequence of events for the Near East (Cheddadi and Rossignol-Strick, 1995). A slight expansion of evergreen oak and other Mediterranean elements early in OIS 4 may have been the equivalent of the Ognon warm phase, but afterwards the treeless landscape was covered by a semi-desert vegetation. In north-west Africa (Fig. 7) the Mediterranean forest almost vanished, the semi-desert moved north (Dupont, 1993) and increasing chenopods suggest desert expansion (Hooghiemstra *et al.*, 1992).

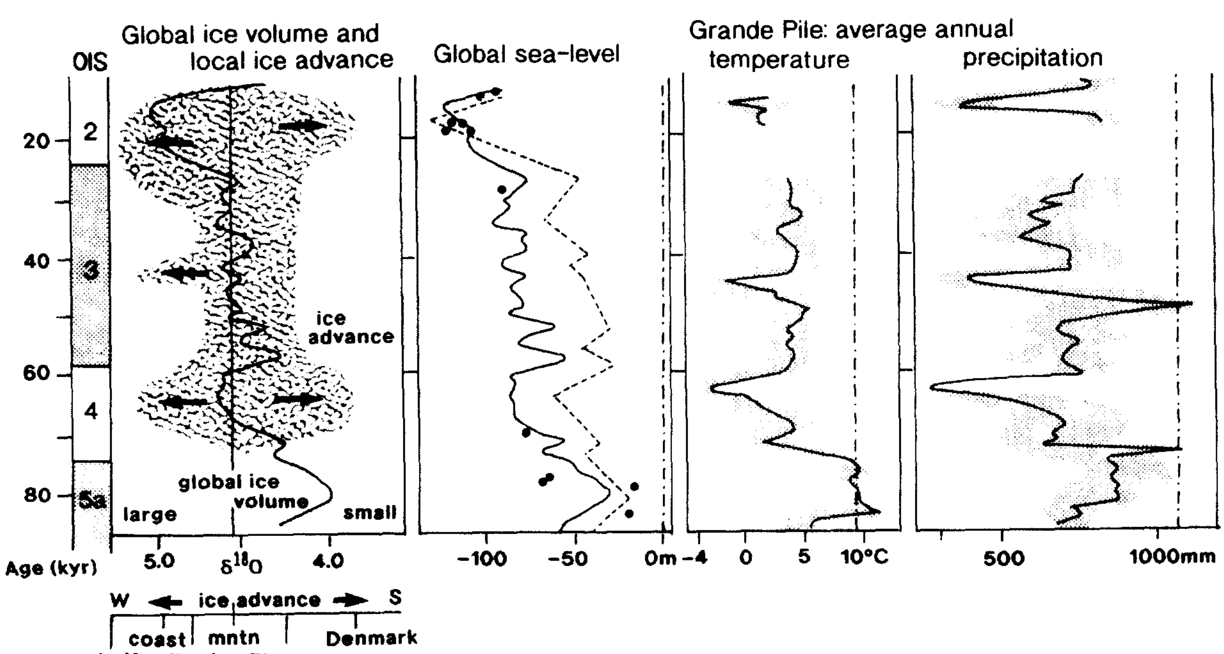

FIG. 12. Palaeoenvironmental changes between 75 and 25 kyr BP (OIS 4–3). Ice volume from N.J Shackleton (1987, Fig. 1); ice advances (shaded) in western Scandinavia (left) and Sweden/Finland to Denmark (right) after Mangerud (1991a, Fig. 5); diagram at lower left relates ice edge to major geographic features. Sea-level changes from Fig. 4; black dots are dated Barbados coral terraces after Bard *et al.* (1990). Central European mean annual temperature and precipitation history after Guiot *et al.* (1989, Fig. 3); confidence intervals shaded. Dashed vertical lines mark present value.

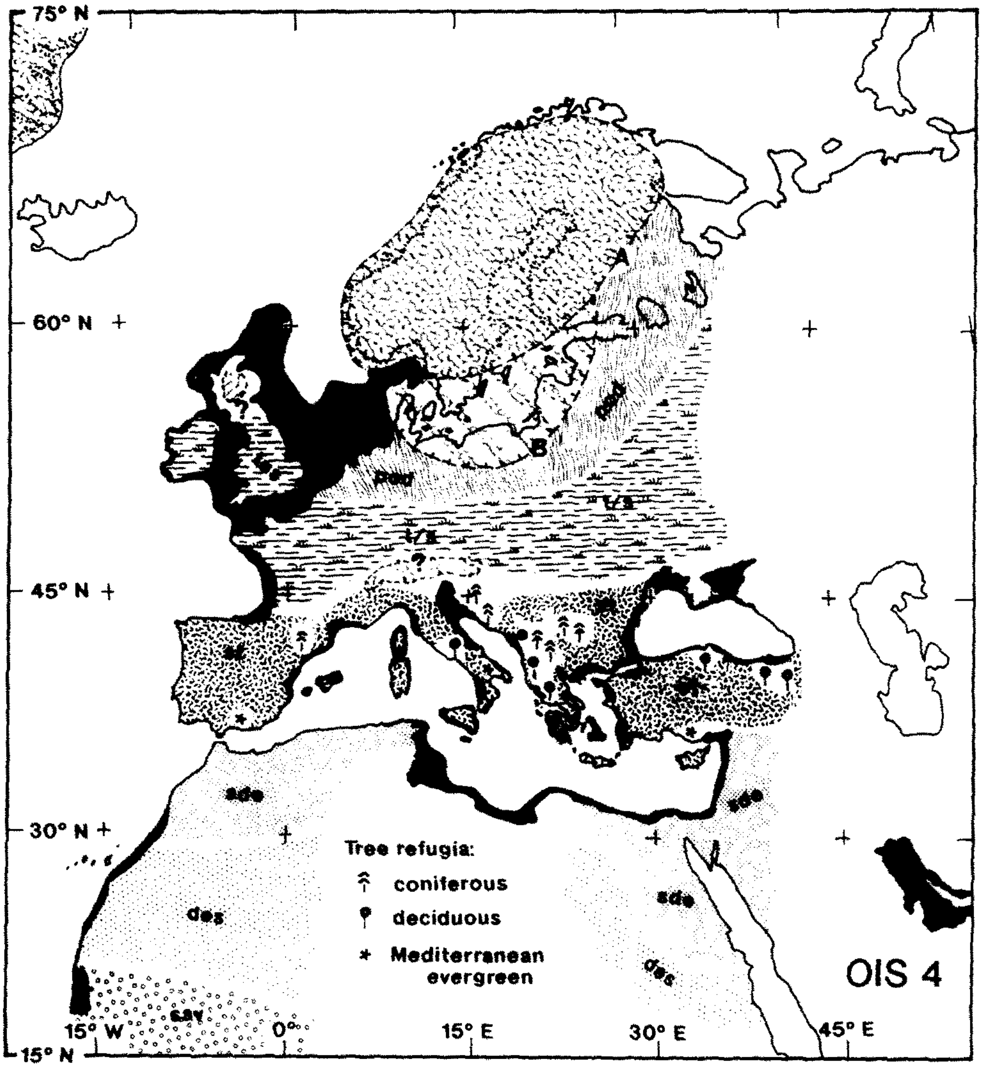

FIG. 13. Sketch map of Europe at 65 kyr BP (late OIS-4). Minimum (A) and maximum (B) Fennoscandian ice margins after Andersen and Mangerud (1989). Pack ice extent and North Atlantic isotherms unknown. Coasts based on −75 m isobath. Palaeoenvironments: cloudy pattern: ice-sheets; black: emerged coastal plains. Palaeoenvironments: pod: polar desert; t/s: tundra and cold steppe mosaic; st: arid cold steppe; sde: semi-desert; des: desert; sav: Sahelian savanna or park land.

The Long Middle Glacial: 59–24 kyr

OIS 3 began with a temporary increase of ice-rafted sediments in the Norwegian Sea (Baumann *et al.*, 1995) that marked a retreat of the Fennoscandian ice-sheet. The ice-sheet remained limited to only a part of Fennoscandia until OIS 2, but its geographic limits (Fig. 14) are as controversial as those of OIS 4 (Andersen and Mangerud, 1989; Mangerud, 1991a; Donner, 1995). Several sites imply nearly complete deglaciation of the Fennoscandian lowland (Donner, 1995, pp. 69–74), but the dates are not robust.

At the start of the OIS 3 interstadial global ice volume decreased somewhat, then slowly increased over the next 30 kyr (Fig. 12), while the level of the sea fluctuated for long around the −50 m isobath, then slowly fell to reach −80 m towards the end of OIS 3 (Bard *et al.*, 1990). In the Mediterranean the low sea-level (Fig. 12) exposed many coastal lowlands (Fig. 14) but they were less extensive than those of full glacials.

A brief mid-OIS 3 cold event (Fig. 12) in western Norway (Mangerud, 1991a; Baumann *et al.*, 1995) may have been due to local conditions as Donner (1995, pp. 79–80) believes who regarded its existence in eastern and southern Fennoscandia as uncertain. Whether it was a true ice advance or not, pollen records show a spread of arctic steppe and tundra at this time.

Brief (100–1000 years), sharp climate oscillations mark OIS 3 in the Greenland GISP2 and GRIP ice cores (Fig. 11), events which Bond *et al.* (1993) have correlated with sea-surface temperature excursions recorded in North Atlantic deep-sea cores. The events combine to cooling cycles which lasted 10–15 kyr and culminated in Heinrich events, major iceberg discharges from the North American (Bond *et al.*, 1992, 1993; Bond and Lotti, 1995) and Norwegian ice-sheets (Fronval *et al.*, 1995). The cycles may reflect alternate ice-sheet growth and collapse associated with periodic position changes of the jet stream as the North Atlantic switched back and forth between different circulation modes (MacAyeal, 1993; Keigwin *et al.*, 1994; Haflidason *et al.*, 1995). When finally the Polar Front retreated north and the

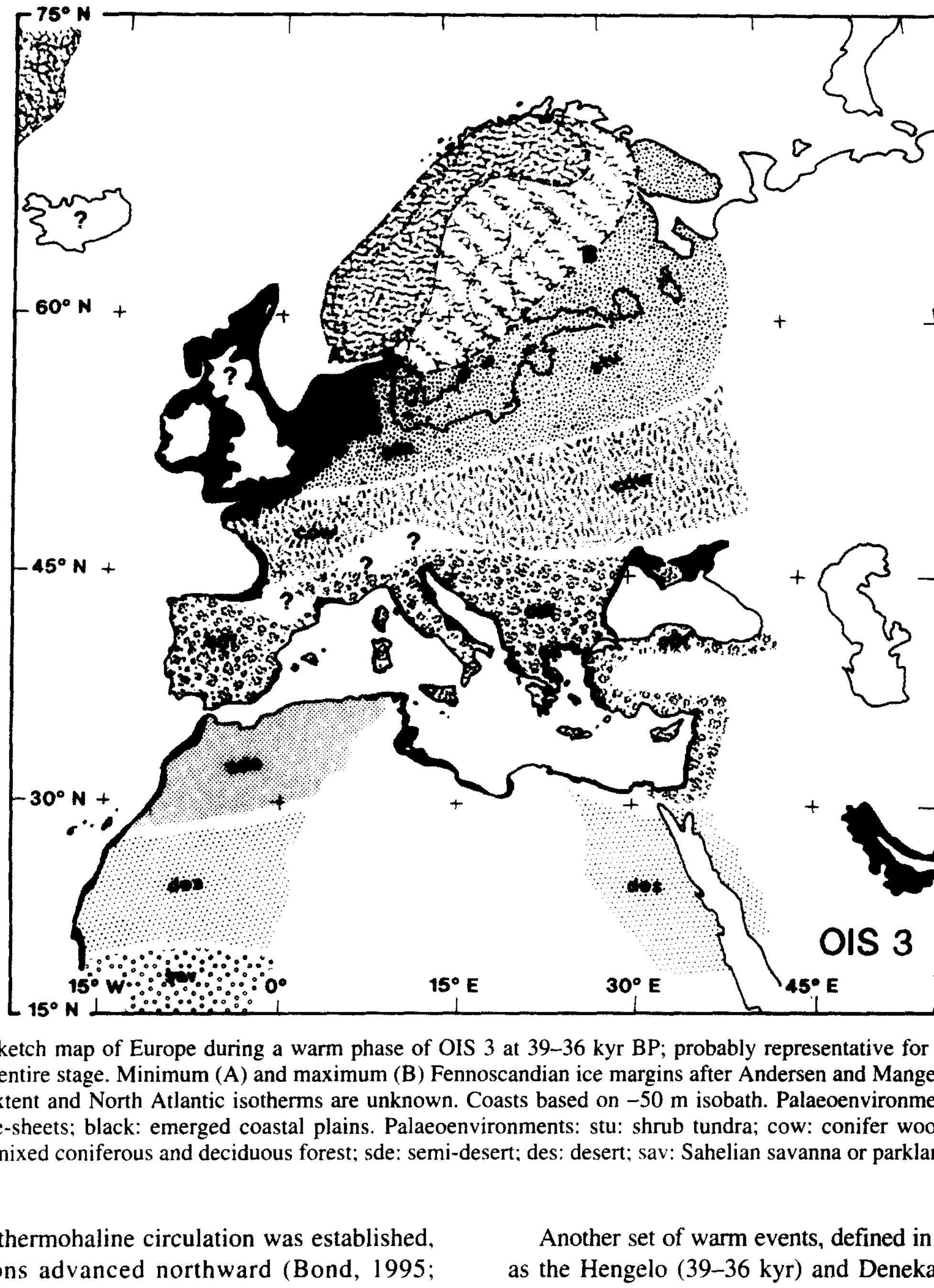

FIG. 14. Sketch map of Europe during a warm phase of OIS 3 at 39–36 kyr BP; probably representative for as much as half of the entire stage. Minimum (A) and maximum (B) Fennoscandian ice margins after Andersen and Mangerud (1989). Pack ice extent and North Atlantic isotherms are unknown. Coasts based on –50 m isobath. Palaeoenvironments: cloudy pattern: ice-sheets; black: emerged coastal plains. Palaeoenvironments: stu: shrub tundra; cow: conifer woodland; cdf: mixed coniferous and deciduous forest; sde: semi-desert; des: desert; sav: Sahelian savanna or parkland.

North Atlantic thermohaline circulation was established, warm conditions advanced northward (Bond, 1995; Broecker, 1994a, b). If this hypothesis is correct, the rapid climatic changes should be reflected in European land records and render feasible correlations between ice and pollen cores (Thouveny *et al.*, 1994).

The north-western European pollen record for OIS 3 displays a number of warm intervals (Fig. 11), such as the Oerel and Glinde events (Behre and Lade, 1986; Behre, 1989), dated at 58–54 and 51–48 kyr BP respectively (Behre and van der Plicht, 1992) and correlated with the Goulotte and Pile phases at La Grande Pile (Woillard and Mook, 1982). An early OIS 3 warm complex noted in the Netherlands (Moershoofd; Zagwijn, 1961) dated at 49–41 kyr may correspond with the Glinde interval (Ran, 1990), but this is considered doubtful by Behre and van der Plicht (1992). Because in terms of vegetation the events differ between sites, the correlations rely on the relative stratigraphic positions and on ^{14}C dates which are not uniformly reliable beyond 35–40 kyr and partly conflict with SPECMAP ages.

Another set of warm events, defined in The Netherlands as the Hengelo (39–36 kyr) and Denekamp (32–28 kyr) intervals (van der Hammen *et al.*, 1967; Ran, 1990; Ran and van Huissteden, 1990), has been placed after 40 kyr BP. At La Grande Pile, this period contains the Charbon and Grand Bois intervals, but their equivalence with the Dutch sequence is not clear. In southern Europe also oscillations have been identified that may correspond to events north of the Alps (e.g. Leroy, 1994), but needed confirmation.

The long pollen sequences from southern Europe provide an opportunity to develop regional correlations based on ^{14}C dating and tephrochronology (e.g. Follieri *et al.*, *in press*) which may lead to a coherent pollen-stratigraphic scheme that can be compared with other records. The correlations between the high-frequency oscillations of the GRIP and GISP2 ice cores and the northern European pollen record (Fig. 11) may then provide the climatic underpinning for the fine structure of OIS 3. However, chronological improvements are needed to develop the true history of the vegetational and climatic changes of the European pleniglacial.

Our reconstruction of OIS 3 conditions at 36–39 kyr BP (Fig. 14) refers to one of the many mild intervals of OIS 3, but is probably representative of the many other mild times which together make up as much as half of OIS 3. The vegetation picture that emerges is one of steep north–south gradients. Pollen records indicate a shrub tundra characterised by *Betula nana* (dwarf birch), *Salix* (willow) and *Juniperus* in northern Germany (Behre, 1989), The Netherlands and the eastern Baltic, with scattered spruce near St Petersburg (Liivrand, 1991). Open *Pinus*, *Picea* and *Betula* woodland covered eastern France and the alpine foreland, but other deciduous trees seem to have been absent north of the Alps (de Beaulieu and Reille, 1984, 1992a, b; Grüger, 1989; Reille and de Beaulieu, 1990). Pollen-based estimates for La Grande Pile (Fig. 12) and Les Echets (Guiot *et al.*, 1989) give mean annual temperatures about 4°C below today's value and a precipitation that was reduced by 200–400 mm (Fig. 12). La Grande Pile beetle assemblages give a mean temperature of 8°C for the warmest and –5°C for the coldest months just before the Grand Bois warm event at ca. 32 kyr BP (Ponel, 1995).

During mild intervals in the Mediterranean, pine woodland with some deciduous oak existed in Catalunya (Pérez-Obiol and Juliá, 1994) while in southern Spain evergreen and deciduous oak populations spread, together with pine and juniper (Pons and Reille, 1988). In central and southern Italy deciduous woodland with *Quercus*, *Corylus*, *Fagus*, *Tilia* (lime) and *Ulmus* expanded (Follieri *et al.*, 1988, *in press*; Leroy, 1994; Rossignol-Strick and Planchais, 1989; Watts, 1985). A similar development took place in northern Greece (Wijmstra, 1969; Tzedakis, 1994), while southern Greece was dominated by the expansion of deciduous and evergreen *Quercus* and *Pinus* and *Juniperus* populations. The OIS 3 Mediterranean woodland was open in character, however; highest tree densities are recorded in only a few places where moisture was sufficient and soil conditions optimal such as in northwestern Greece and central Italy.

In the Near East areas of evergreen woodland expanded within a mainly semi-desert landscape (van Zeist and Bottema, 1991; Cheddadi and Rossignol-Strick, 1995). In north-western Africa, gaps in the southernmost cores leave us with fragmentary evidence for late OIS 4 and early OIS 3 (Fig. 7), but the desert and semi-desert remained close to their widest extent and the tropical rainforest kept to the far south. An arid grass and *Artemisia* steppe marked the Mediterranean coastal zone where the woodland had withdrawn to a narrow strip in the Atlas mountains, except during a slightly warmer phase in late OIS 3 (Hooghiemstra *et al.*, 1992). In north-eastern Africa high desert conditions prevailed and pluvial conditions were absent (Szabo *et al.*, 1995).

EPILOGUE

Two fundamental points have been emphasised in the preceding discussion: the need to date human history accurately before it can be compared with a palaeoenvironmental record that changes on timescales of millennia or less, and the fact that much of the time the glacial climate was far less severe than during its maxima. The palaeoenvironmental graphs and maps that depict the changing European landscape vividly illustrate these points. The importance of the first point is widely recognised in Palaeolithic studies and although its implementation is very difficult, it deserves a high priority, but the truth of the second one appears to be rarely recognised.

Secondarily, the probability of sharp climatic excursions lasting a few centuries to two millennia deserves attention. Their impact on the pleniglacial landscape and on early human affairs is potentially great. The Little Ice Age (Grove, 1988) and current concerns about global warming underscore the point, but similar changes may well have gone unnoticed by early hunters and gatherers or were compensated by slow migrations. Until we learn much more about the nature of those brief but strong mid- and high-latitude events, their influence on human affairs can only be speculated upon.

We recognise that our main effort, the construction of a spatial and temporal palaeoenvironmental history of Europe and its periphery from 140 to 25 kyr BP, is but a start and that it omits large parts of the Old World at least as important as Europe in terms of human wanderings. Moreover, the local detail that we are unable to give may well be of greater interest to archaeologists than the broad perspective presented here. This is inevitable now, but better and more strongly supported syntheses should soon appear.

Maps are persuasive. By their nature they fill the space available, but the degree of confidence the filling deserves is not easily perceived by the reader. They are also static, showing the world as a series of discontinuous images without transitions. In reality the transitions not only occupy much of the total time but also might have provided by themselves some of the forces that drove the migrations of human tribes.

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