

Early–Middle Pleistocene transitions: an overview and recommendation for the defining boundary

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Abstract: The Early–Middle Pleistocene transition (c. 1.2–0.5 Ma), sometimes known as the ‘mid-Pleistocene revolution’, represents a major episode in Earth history. Low-amplitude 41-ka obliquity-forced climate cycles of the earlier Pleistocene were replaced progressively in the later Pleistocene by high-amplitude 100-ka cycles. These later cycles are indicative of slow ice build-up and subsequent rapid melting, and imply a transition to a strongly non-linear forced climate system. Changes were accompanied by substantially increased global ice volume at 940 ka. These climate transformations, particularly the increasing severity and duration of cold stages, have had a profound effect on the biota and the physical landscape, especially in the northern hemisphere. This review assesses and integrates the marine and terrestrial evidence for change across this transition, based on the literature and especially the following 17 chapters in the present volume. Orbital and non-orbital climate forcing, palaeoceanography, stable isotopes, organic geochemistry, marine micropalaeontology, glacial history, loess–palaeosol sequences, pollen analysis, large and small mammal palaeoecology and stratigraphy, and human evolution and dispersal are all considered, and a series of discrete events is identified from Marine Isotope Stage (MIS) 36 (c. 1.2 Ma) to MIS 13 (c. 540–460 Ma). Of these, the cold MIS 22 (c. 880–870 ka) is perhaps the most profound. However, we here endorse earlier views that on practical grounds the Matuyama–Brunhes palaeomagnetic Chron boundary (mid-point at 773 ka, with an estimated duration of 7 ka) would serve as the best overall guide for establishing the Early–Middle Pleistocene Subseries boundary.

The transition from Early to Middle Pleistocene is marked by fundamental changes in Earth’s climatic cyclicality. Orbital obliquity at 41-ka cycles, which had dominated the earlier part of the Pleistocene, was superseded progressively by a 100-ka rhythm of climate change accompanied crucially by increased-amplitude climatic oscillations. Even prior to this time, global ice volume was increasing. This transition has become known as the ‘mid-Pleistocene revolution’ (Berger & Jansen 1994) or more prosaically the ‘mid-Pleistocene transition’. In its broadest sense, this interval occurred between about 1.2 Ma and 500 ka, and the 100-ka cyclicality is currently thought to be associated with every fourth or fifth precessional cycle. Precession has an important control on seasonality, and there is interest in whether increased seasonality featured in the evolving climate regime. The glacial–interglacial world in which we now live is the result of these changes, and stands against a backdrop of progressively cooling global temperatures since the Middle Eocene (Zachos *et al.* 2001). The impact of this transition on terrestrial and marine biota has been profound and manifold, but the dating and correlation of events has been problematic, leading to uncertainties over cause

and effect. Even the longest ice core, from Antarctica, has currently yielded data extending back only to 740 000 years BP (EPICA community members 2004), and direct evidence of earlier atmospheric change is presently unavailable. Despite these limitations, climate fluctuations are clearly driven by the Earth’s orbital variations and modulated by feedback mechanisms. An integrated global overview of the terrestrial and marine realms, based on this volume, is provided below to understand more fully the Earth’s response as a system (Figs 1 & 2).

A secondary objective of this compilation has been to consider the stratigraphic position of the Early–Middle Pleistocene Subseries boundary. This boundary, although not yet formally defined, is usually placed at the Matuyama–Brunhes palaeomagnetic Chron boundary (773 ka; Channell *et al.* 2004), following the recommendations of the International Union for Quaternary Research (INQUA) Working Group on Major Subdivisions of the Pleistocene (Richmond 1996). This boundary falls within Marine Isotope Stage (MIS) 19 (Channell *et al.* 2004, and references therein). Most of the contributors to this volume adopt this recommendation, although there are some exceptions: Reale &

Monechi do not commit to a specific boundary position but accept that it should be within the interval between MIS 25 and the Matuyama–Brunhes Chron boundary; Dodonov places the boundary at the base of MIS 21; and both Palombo *et al.* and Palombo & Valli place the boundary at MIS 25 following recommendations by Cita & Castradori (1994, 1995).

The International Union of Geological Societies has charged its International Commission on Stratigraphy (ICS) with the task of recommending global stratotype sections and points (GSSPs) for all major boundaries in the geological column so that they may be ratified by the year 2008. As part of this process, the ICS Subcommittee on Quaternary Stratigraphy, through the efforts of a designated Working Group on the Early–Middle Pleistocene Boundary, is required to recommend a GSSP for the Early–Middle Pleistocene Subseries boundary. It is clearly essential to understand the geological background to the mid-Pleistocene transition in advance of any final decisions on this matter.

Overview of mid-Pleistocene transitions

Maslin & Ridgwell set the scene by reviewing the Early–Middle Pleistocene transition in terms of the Earth as a system. They provide a helpful summary of the orbital forcing components, review current hypotheses explaining the ‘mid-Pleistocene revolution’, and refute the still frequently held view that the switch from 41-ka to 100-ka dominant cyclicity at the transition is driven by orbital eccentricity. This ‘eccentricity myth’ has arisen because eccentricity is the only orbital forcing factor with a 100-ka period. The authors explain, however, that the 100-ka glacial–interglacial cycles that characterize the Middle Pleistocene are more closely linked to precessional forcing, which has a period of 23 ka, with each cycle being defined by the fourth or fifth precessional cycle (see also Maslin *et al.* 2001). This is to be expected because precession, while having its greatest influence in the tropics, is still the dominant influence on solar influx received at the critical latitude of 65°N (e.g. Berger & Loutre 1991; Berger *et al.* 1999). Eccentricity may modulate glacial–interglacial cyclicity but does not drive it.

The ‘mid-Pleistocene revolution’, as **Maslin & Ridgwell** explain, has two phases. The first begins at about 940–890 ka and is characterized by a significant increase in global ice volume but the persistent dominance of 41-ka cyclicity. However, a speculated

‘attempt’ to move towards a Middle Pleistocene mode of glaciation occurred as early as 1.2 Ma (MIS 36). The second phase, beginning about 725–650 ka, is marked by strong 100-ka cyclicity and high amplitude (Mudelsee & Statterger 1997). However, none of the orbital-forcing parameters vary significantly across the ‘mid-Pleistocene revolution’ (Berger & Loutre 1991; Berger *et al.* 1999), implying that internal feedbacks in the Earth climate system are responsible for the glacial–interglacial cycles. In particular, the asymmetrical ‘sawtooth’ waveform of the 100-year cycle, characterized by slow ice build-up and subsequent rapid melting, implies a transition to a strongly non-linear forced climate system at about 725–650 ka. Several mechanisms, in particular CO₂ levels in the atmosphere (Ruddiman 2003) and a long-term decline in global temperatures (Zachos *et al.* 2001), have been proposed to account for this non-linearity, and are evaluated by the authors.

Marine realm

Schefuß *et al.* have used geochemistry to explore the eastern tropical Atlantic as a means of assessing the low-latitude response to the onset of glaciation in higher latitudes. They have looked specifically at lipid biomarkers and bulk organic geochemistry of Ocean Drilling Program (ODP) Site 1077, drilled on the lower Congo Fan and within the Angola Basin. This site was chosen because of its close proximity to the Congo (Zaire) River, which is the world’s second largest river. These sediments contain a detailed record of variations in transported terrestrial vegetation through time, both as river-transported organic material and wind-blown terrigenous plant waxes. River discharge, and its contained nutrients, have also left their mark on the local marine phytoplankton, although upwelling and other changes in hydrography have also affected the phytoplankton. By using lipid analysis to examine changes in phytoplankton composition through time, as well as by analysing the terrestrial organic input, the authors have been able to record detailed marine and terrestrial environmental changes through time. The influences of orbital forcing on these environmental changes have been evaluated by cross-spectral analyses of the data.

Schefuß *et al.* show from their wind-blown plant wax records that trade-wind strength and zonality increased significantly with the onset of 100-ka cyclicity, and that sea-surface temperature controlled the atmospheric water balance, leading to highest

Fig. 1. (See p. 2) Global chronostratigraphic correlation chart for the past 1.8 Ma. Modified from Gibbard *et al.* (2005a, b). Note that the base of the Quaternary, and of the Pleistocene, extends to c. 2.6 Ma based on the proposal of Gibbard *et al.* (2005a). The Matuyama–Brunhes boundary is here placed at 780 ka following Gibbard *et al.* (2005b), but we note that this boundary has recently been recalibrated as 773 ka (Channell *et al.* 2004); GSSP = global stratotype section and point.

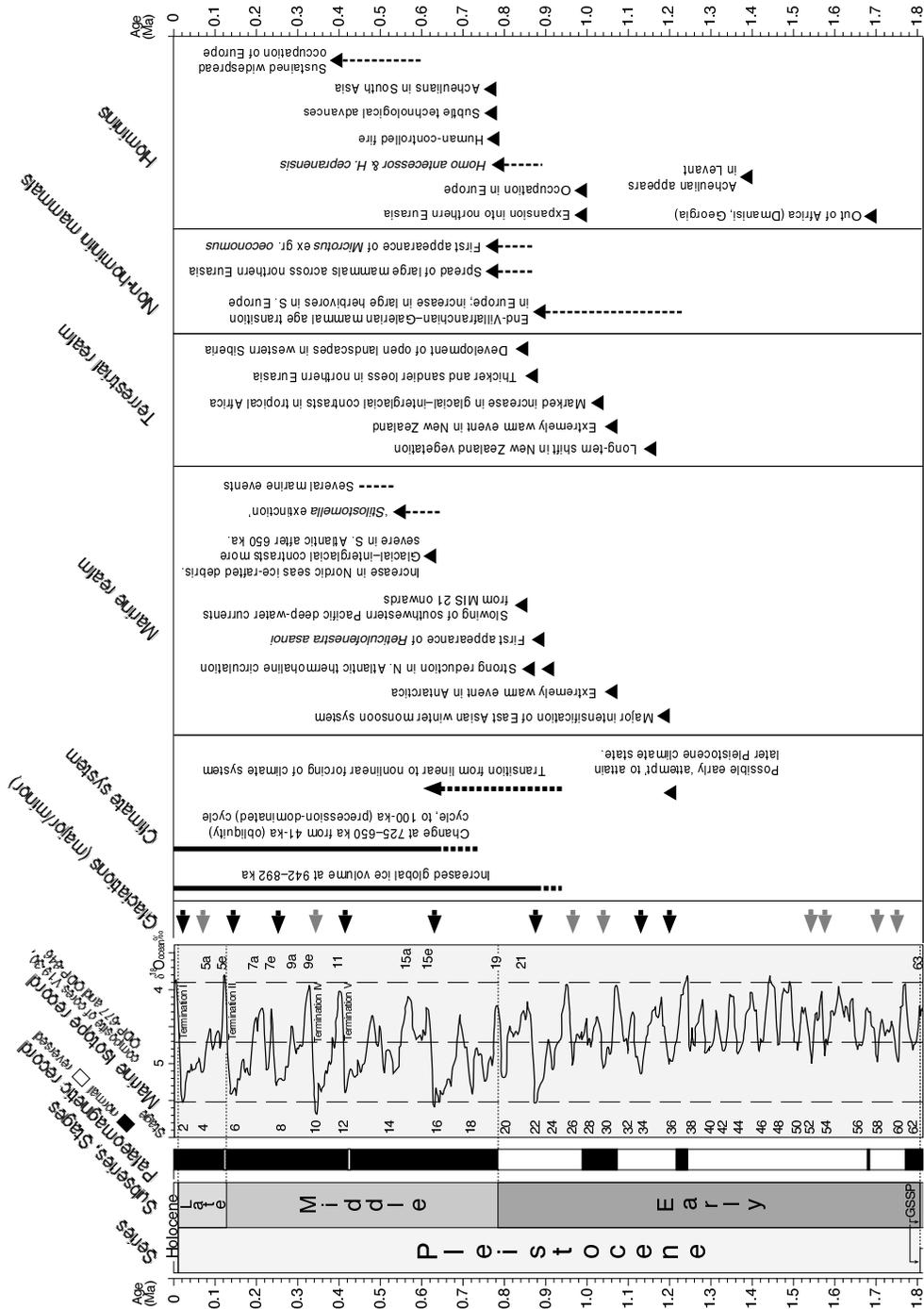


Fig. 2. Major events during the Early-Middle Pleistocene transition. Glaciations (major = solid arrows, minor = grey arrows) are compiled from Ehlers & Gibbard (2004a, b, c). See caption to Figure 1 and text for details.

continental aridity when sea-surface temperatures were at their lowest. High trade-wind strength in fact precedes continental aridity, and indicates that the vegetation became destabilized by increasing wind strength (see also Schefuß *et al.* 2003, and submitted). This interpretation agrees with an increase in dust accumulation in deep-sea cores off NW Africa and the equatorial Atlantic (e.g. Ruddiman & Janecek 1989; Tiedemann *et al.* 1994; deMenocal *et al.* 1993; Wagner 2000). Although marine productivity remained relatively constant throughout the mid-Pleistocene transition, there was a significant change in the marine ecosystem. Prior to the onset of 100-ka cyclicity, marine productivity was mainly controlled by monsoon runoff and hence enhanced siliceous production. After the onset, marine productivity was influenced mainly by wind-driven coastal and oceanic upwelling. This change illustrates the profound effect that aridification of the continent and strengthening of the trade winds exerted on tropical African environments and processes.

The Congo Fan is also the focus of the chapter by **Jahn *et al.*** These authors report on organic matter accumulation, organic carbon isotopic composition, and iron content during the past 1.7 Ma at ODP Site 1075, which is located on the lower Congo Fan about 70 km NW of ODP Site 1077 as analysed by Schefuß *et al.* (this volume). Indeed, **Jahn *et al.*** correlate the orbital signatures of these two sites precisely using magnetic susceptibility. These results are then compared with those of ODP Site 663, situated about 2400 km to the west, in the Gulf of Guinea. Milankovitch-scale analysis of ODP Site 1075 has yielded variations in marine productivity driven by changes in the trade winds, and in Congo River discharge which provides a measure of African precipitation. The African climate was mainly controlled by low-latitude insolation changes (indicated by a strong precessional signal) during the interval considered: a strong 100-ka signal in the total organic carbon signal is superimposed on this only for the past 0.6 Ma. However, results of terrestrial iron input to the Congo Fan show a pronounced 100 ka signal well before global glacial–interglacial oscillations increased in amplitude between 0.9 and 0.6 Ma. Eccentricity modulation of the low-latitude insolation seems to have directly influenced the equatorial African monsoon system. **Jahn *et al.*** moreover suggest that low-latitude precessional forcing and response of the monsoon in the tropics may have been important for 100-ka cyclicity in global climate well before the onset of major glaciation at high latitudes.

A shift in the vegetation of west equatorial Africa to pronounced glacial–interglacial cyclicity is known to have occurred at 1.05 Ma (MIS 30) (Dupont *et al.* 2001). Between 0.9 and 0.6 Ma, interglacials were characterized by warm dry conditions and glacials by cool humid conditions (Dupont *et al.*

2001). In the southeastern South Atlantic, glacial–interglacial contrasts in regional ocean circulation strengthened after 1.2 Ma and became more severe after 650 ka (Diekmann & Kuhn 2002).

Ferretti *et al.* emphasize that thermohaline circulation in the North Atlantic is a critical component of the climate system because it transports heat to the high northern latitudes. This circulation system is driven largely by the sinking of dense, cold, saline water masses in the Norwegian–Greenland seas and Labrador Sea, which form south-flowing North Atlantic Deep Water (NADW). Circulation in the North Atlantic Ocean is not a closed system, however, because NADW continues to flow southwards into the South Atlantic and beyond. Indeed, today NADW ventilates more than half of the volume of the deep oceans, and has a global influence on deep-water characteristics. Moreover, Antarctic Bottom Water (AABW), which forms around the margins of Antarctica, flows sufficiently far northwards that it may penetrate the deep basins of the North Atlantic. Because AABW has lower $\delta^{13}\text{C}$ values than NADW, its influence over time is reflected in the isotopic ratio of tests of benthic foraminifera in the fossil record. Isotopic studies of these fossils in the North Atlantic have inferred that during Quaternary glacial intervals, the contribution of north-flowing AABW increased (declining $\delta^{13}\text{C}$ values). This is attributed to a decrease in the flux of NADW and therefore of thermohaline circulation in the North Atlantic (Raymo *et al.* 1990). It follows that $\delta^{13}\text{C}$ values in benthic foraminifera can be used to reconstruct the intensity of thermohaline circulation in the North Atlantic over deeper time, and this has been one of the key aims of **Ferretti *et al.***

These authors have examined the oxygen and carbon isotopes of benthic and planktonic foraminifera at ODP Site 1063 in the western North Atlantic. The southerly position of this site within the North Atlantic and its considerable water depth makes it ideal for assessing the influence of AABW relative to NADW. The 1020 to 738 ka time interval has been examined at high resolution because this critically spans the mid-Pleistocene transition. The $\delta^{18}\text{O}$ planktonic signal from this site represents one of the highest resolution (less than 0.6 ka on average) records available for the mid-Pleistocene transition. It shows detailed variations of global ice volume through time and allows precise orbital tuning of the record. Of particular interest is the presence of millennial-scale (sub-Milankovitch) fluctuations. These occur throughout the entire interval, but show highest amplitudes during interglacials, particularly after about 900 ka. The MIS 25–24, 23–22 and 19–18 transitions all feature brief fluctuations, and climatic instability appears to be a pervasive feature of the record (see also Kleiven *et al.* 2003). The benthic $\delta^{13}\text{C}$ record shows predictable shifts to low

values during glacials, implying a reduced NADW flux and concomitant strong influence of southern-sourced bottom water. This is particularly notable during MIS 24 and 22 where nearly pure AABW is detected, indicating a strong production of AABW relative to NADW. The same kind of millennial-scale fluctuations present in the $\delta^{18}\text{O}$ are also seen in the $\delta^{13}\text{C}$ record, suggesting that suborbital fluctuations in bottom water circulation occurred throughout the 1020 to 738 ka time interval.

In the South Atlantic, there is evidence of a marked stagnation of deep-water circulation, and hence reduction in the thermohaline circulation, occurring as early as 920 ka which corresponds to MIS 24 (Schmieder *et al.* 2000). As far back as MIS 31 (base of the Jaramillo Subchron; 1.07 Ma), an extremely warm interglacial event in the Ross Sea, Antarctica, has been suggested as representing a significant step in the transition to Middle Pleistocene glacial–interglacial cycles (Scherer *et al.* 2004). Nonetheless, in the central Nordic seas, ice-rafted debris does not increase substantially until MIS 16 (about 650 ka; Helmke *et al.* 2005).

The chapter by **Reale & Monechi** emphasizes the important role of calcareous nannofossil biostratigraphy in defining the Early–Middle Pleistocene boundary in the Mediterranean region. Indeed, the lowest occurrence of *Gephyrocapsa* sp. 3 of Rio (1982) at the base of the *Pseudoemiliania lacunosa* Zone was used by Cita & Castradori (1994, 1995) to define the boundary between the Calabrian and Ionian marine stages (Fig. 1), which they also proposed to serve as the Early–Middle Pleistocene Subseries Boundary (a proposal later considered premature; Castradori 2002). This event is close to MIS 25. *Gephyrocapsa* sp. 3 belongs to a group of medium-sized *Gephyrocapsa* spp. that are warm-water taxa. Unfortunately for boundary purposes, their re-entry datum is strongly diachronous (Raffi 2002, p. 45). Another important and slightly younger datum in the Mediterranean is the highest occurrence of *Reticulofenestra asanoi*. This datum corresponds to the MIS 23–22 transition and is considered synchronous across mid- and low latitudes globally. It is clearly related to the severe cooling of MIS 22 (Raffi 2002).

The stratigraphic importance of *Reticulofenestra asanoi* has prompted **Reale & Monechi** to undertake a detailed investigation of the distribution of this and other species in the North Atlantic Ocean (Deep Sea Drilling Project (DSDP) Hole 610A) and western and eastern Mediterranean Sea (ODP Sites 976 and 963 respectively), allowing them to refine the nannofossil biostratigraphy across the mid-Pleistocene transition. The authors have found that the lowest common occurrence of *R. asanoi* is apparently diachronous, correlating to MIS 35 in DSDP Hole 610A, the MIS 34–33 transition at ODP

Site 976, and MIS 30 at ODP Site 963. This pattern is consistent with other published records for the Atlantic and Mediterranean. The lowest occurrence of *Gephyrocapsa* sp. 3 of Rio (1982) has an inferred correlation to the MIS 25–24 transition in DSDP Hole 610A, the MIS 26–25 transition at ODP Site 976, and MIS 27 at ODP Site 963. This confirms earlier studies that this datum is latitudinally diachronous, and argues against its use for defining the Early–Middle Pleistocene Subseries boundary as proposed by Cita & Castradori (1994, 1995). **Reale & Monechi** have found that the highest occurrence of *R. asanoi* correlates to about MIS 22 in DSDP Hole 610A, to the MIS 23–22 transition at ODP Site 976, and to about MIS 23 at ODP Site 963. These findings strengthen the claim that this datum is more or less synchronous with the MIS 23–22 transition, and emphasize the importance of this datum when considering the placement of the Early–Middle Pleistocene boundary.

The focus of changes in the marine realm moves to the SW Pacific with the chapter by **Hayward *et al.*** These authors note that the connection between northern hemisphere glaciations and reduced production of NADW has long been drawn for the Pleistocene in general (e.g. Mix & Fairbanks 1985), and the mid-Pleistocene in particular (Raymo *et al.* 1990). Indeed, investigations of numerous cores from the South Atlantic imply that global ice increase at 920 ka was accompanied by a sudden stagnation in deep-water circulation, and that an interim state consistent with reduced deep-water circulation was terminated at about 540 to 530 ka (early MIS 13) by another event of probably global significance (Schmieder *et al.* 2000). Against this background, **Hayward *et al.*** have investigated the record of deep-sea benthic foraminifera in the SW Pacific to assess whether there has been any major oceanographic reorganization in this region during the mid-Pleistocene transition. An improved knowledge of bottom-water changes might help explain why a group of at least 50 elongate, cylindrical, deep-sea foraminiferal species (including the families Pleurostomellidae and Stilostomellidae) became extinct during the mid-Pleistocene transition (the so called ‘*Stilostomella* extinction’). The authors have examined three bathyal sequences in the SW Pacific, DSDP Site 594 and ODP Sites 1119 and 1125, situated above and below the present-day Subtropical Front. Canonical correspondence analysis shows that short-term and local influences were more important than major oceanographic changes in controlling assemblages. Nonetheless, north of the Subtropical Front, faunas between MIS 21 and 15 show increased fluctuations in composition that suggest increased food supply and a decrease in dissolved bottom oxygen. This appears to be caused by slowing of deep-water currents during the mid-Pleistocene tran-

sition in the region; a similar conclusion was reached based on sortable silt investigations of the nearby ODP Site 1123 (Hall *et al.* 2001). **Hayward *et al.*** therefore conclude that the ‘*Stilostomella* extinction’, which occurred during MIS 15, was linked to slower deep-water currents and a decrease in bottom oxygenation. The most conspicuous event at the end of the mid-Pleistocene transition is an abrupt rise and decline in the abundance of *Abditodentrix pseudothalmanni* south of the Subtropical Front during MIS 15–12 (*c.* 620–420 Ma). This event suggests an increase in food supply and lower dissolved oxygen, and may be related to several pronounced climatic changes during the close of the mid-Pleistocene transition including a rearrangement of ocean circulation inferred from an unusual sediment facies in the South Atlantic at *c.* 540–530 ka (early MIS 13; Schmieder *et al.* 2000).

Terrestrial realm

The oceans play a critical role in driving and moderating global climate, but it is on land that evidence for climate change is most visible. The chapter by **Dodonov** begins a series of contributions assessing the terrestrial evidence for mid-Pleistocene change. The loess–palaeosol sequences document the alternation of arid and humid conditions on land, and have provided high-resolution records of climatic development across northern Eurasia from the Loess Plateau of China, through central Asia and into Europe. Throughout the Quaternary, these alternations show good correspondence with the marine isotope record, implying that they are controlled by orbital perturbations (e.g. Kukla 1978, 1987; Ding *et al.* 1991; Fig. 1). Indeed, loess–palaeosol sequences in the Loess Plateau of China appear to provide a detailed record of past variations in the monsoon (Ding *et al.* 1999, and references therein). **Dodonov** has reviewed and integrated the evidence for climatic change across the mid-Pleistocene transition in northern Eurasia, drawing on literature from across the region including poorly accessible Russian sources. It is apparent that during the mid-Pleistocene transition, individual loess horizons within loess–palaeosol sequences become thicker, a shift tentatively related to increasingly pronounced 100-ka climate cycles accompanied by relatively cooler and drier atmospheric conditions. **Dodonov** explains that this appears to result from high atmospheric pressure becoming established over the expanding north European ice sheet, hindering the transfer of western air masses to the east, and increasing aridity in inner parts of Eurasia. This would have led to an expansion of the Siberian–Mongolian high-pressure system and thence to an intensification of the winter monsoon in eastern Asia as well as increasing loess formation in central Asia and north China.

Heslop *et al.* (2002), based on loess–soil records of northern China, have already shown that a major intensification of the east Asian winter monsoon system occurred at about 1.20 Ma (MIS 36). **Dodonov**’s detailed comparisons of loess–palaeosol sequences across the mid-Pleistocene transition show clear potential for correlation throughout northern Eurasia despite difficulties owing to variable sedimentation rates and weathering depths, and incomplete sequences. In particular, the base of the climatostratigraphic unit representing MIS 21 is advocated as being a convenient and logical marker for recognizing the Early–Middle Pleistocene boundary in loess–palaeosol sequences. This is because MIS 22, which represents the first strongly negative (cold) deviation in the marine isotope record of the mid-Pleistocene transition, is reflected by thicker or sandier loess deposits in northern Eurasia that signal the onset of severe palaeoclimatic conditions (see also Heslop *et al.* 2002, and below). In northwestern Europe, the pronounced glacial–interglacial cycles that typify the Middle Pleistocene and onwards appear to begin with the Menapian (see p. 12). Perhaps surprisingly, the Dorst ‘Glacial’ of the Bavelian Stage in the Netherlands, which is probably equivalent to MIS 22 (De Jong 1988; Funnell 1995; Westerhoff *et al.* 2003; Fig. 1), is not an especially pronounced cold stage, although there are reservations about the veracity of the Early Pleistocene chronology in the Netherlands (Drees 2005).

Dodonov notes that the conventional use of the Matuyama–Brunhes palaeomagnetic Chron boundary as representing the Early–Middle Pleistocene boundary in northern Eurasia is compromised by problems of displaced magnetic remanence in loess–palaeosol sequences. This explains why the Matuyama–Brunhes boundary occurs in a (cold) loess horizon representing MIS 20, when it should in fact be in a palaeosol representing MIS 19 (Zhou & Shackleton 1999).

Vegetational response to the climate changes that culminated in the ‘mid-Pleistocene revolution’ are addressed in the paper by **Byrami *et al.*** who present an investigation of a 50-m-deep sequence at Auckland in northernmost North Island, New Zealand. Fission-track dating of tephras, palaeomagnetism, biostratigraphy and orbital tuning to the ocean-core marine isotope stratigraphy indicate that the upper 28 m of the sequence represents late Early Pleistocene MIS 45–28, or *c.* 1.4 to 1.0 Ma. During this period detailed vegetational changes are indicated by well-preserved pollen assemblages recovered from the sediments. **Byrami *et al.*** record repeated cyclic changes from a coniferous- to *Nothofagus*-dominated regional woodland vegetation that are in step with the marine isotope events during this interval. These changes are interpreted to result from warmer to cooler climatic oscillations that reflect

moist-temperate stable climate for the coniferous vegetational community and cool but not cold climates for the *Nothofagus*.

A more permanent increase in the *Nothofagus* forest from MIS 35 onwards implies a response to a longer-term climatic shift interpreted as reflecting greater temperature extremes between warm and cool episodes after that time. This was accompanied by decreased humidity and increased seasonality during the cool periods, a condition characteristic of the Late Pleistocene in the area. **Byrami et al.** consider that although their sequence predates the 'mid-Pleistocene revolution' *sensu stricto* by 100 ka, the changes detected imply that the climate of northern New Zealand was already becoming progressively more similar to that of Late Pleistocene in the region by 1.0 Ma.

The chapter by **Capraro et al.** investigates the c. 50-m-thick Valle di Manche section of the San Mauro succession in the Crotone Basin, Calabria (southern Italy) in order to calibrate and compare marine and terrestrial climate signals. The Crotone Basin, a classic site for Pleistocene stratigraphy, contains the Vrica section in which the Pliocene–Pleistocene Series boundary is defined by a GSSP. The Valle di Manche section represents an upper-shelf depositional environment, contains the Matuyama–Brunhes magnetic reversal, and represents a critical interval from MIS 21 to MIS 18 or 17. Pollen analyses have been undertaken to reconstruct vegetational and climatic changes, and a highly resolved marine isotope stratigraphy is established from planktonic and benthic foraminifera. Even deposits representing glacial maxima are represented owing to rapid tectonic subsidence of the sub-basin. **Capraro et al.** have shown that the vegetational development of the area is clearly linked to the marine signal. In particular, mesic forest succeeded wooded steppe just before the Matuyama–Brunhes boundary, continues across this boundary, and represents the most negative (warmest) values of $\delta^{18}\text{O}$ for MIS 19. This indicates that the vegetation was fully in phase with the marine isotope record during MIS 19. Results correspond with those of Bassinot *et al.* (1994) in placing the Matuyama–Brunhes boundary within the middle part of MIS 19, rather than at the MIS 20–19 transition or the upper part of MIS 19 as suggested by Berger *et al.* (1995) and Channell & Kleiven (2000), respectively. **Capraro et al.** also caution, however, that precise placement is difficult because the three substages of MIS 19 are not generally identified in the isotope records.

A surprising discovery is the existence of alpine forest conditions in the Crotone area during the glacial intervals of MIS 20 and 18, in contrast to earlier views that the central Mediterranean area was characterized by arid conditions during glacial intervals.

A broader view of vegetational and climatic evolution for the Mediterranean is given by **Suc & Popescu**. These authors have synthesized the 16 most important pollen records available for the north Mediterranean region during the past 2.7 Ma. The pollen records are replotted as ecological groups, allowing their climatic significance to be evaluated and interpreted in terms of orbital cyclicity. The Mediterranean region today is particularly characterized by dry summers, and its bioclimatic subdivision has existed in its modern form since mid-Pliocene times, with later episodes merely amplifying these subdivisions (Suc 1984; Suc *et al.* 1995a). Moreover, the same overall pattern of warming and cooling observed in NW Europe for the Late Pliocene and Pleistocene (Zagwijn 1975) is also recognized in the northern Mediterranean pollen record, although the latter is complicated by the effects of gradients caused by longitude, latitude and altitude. The longitudinal gradient is imposed by the influence of the Asiatic monsoon, as shown for example by the persistence of thermophilous trees in the eastern Mediterranean. Regarding the critical interval from 1.2 to 0.7 Ma, **Suc & Popescu** perhaps surprisingly report no strikingly important change in the pattern of vegetational development from glacials to interglacials during this time, nor any especially pronounced expression of MIS 22. There was a floral change in the NW Mediterranean; for example, the taxonomic composition of *Artemisia* steppes changed considerably between 2.6 and 1.0 Ma, probably in response to a lowering of temperatures. However, this change is not reflected in southern Italy, indicating the strong latitudinal (and its altitudinal equivalent) control on Mediterranean floral development over this time interval.

Non-hominin mammals

The large-mammal record shows a significant response to climatic and environmental change across the mid-Pleistocene transition in Europe, with the famous 'end-Villafranchian event' representing a major faunal turnover (Azzaroli 1983; Azzaroli *et al.* 1988). The event is marked for example by the first European occurrences at around 0.9 Ma of the extant leopard *Panthera pardus*, spotted hyaena *Crocuta crocuta* and extinct hyaenid *Pliocrocuta perrieri* (O'Regan *et al.* 2002).

Palombo et al. have assessed environmental changes during the mid-Pleistocene transition in Italy by statistically analysing the fossil record of large mammals. Italy served as a refuge for warm-adapted mammals because of its north–south orientation, and the Alpine mountain chain at its northern limit has acted as an ecological filter. The authors

use a new method involving ‘bootstrapped’ cluster analysis to assess body size and diversity trends, as well as changes to the community structure across the transition. The shift from 40- to 100-ka climatic periodicity coincides in Italy with the successive arrival of large ungulates from eastern and central Europe, culminating in a sharp increase in diversity in the early Middle Pleistocene. This led to a major reorganization of large-mammal communities within the interval from c. 920 to 750 ka (from the early to middle Galerian mammal age). Statistical analyses of the mammalian record reveal open and arid conditions in the late Early Pleistocene followed by a wetter and substantially cooler climate in the early Middle Pleistocene. These findings support the view that climatic changes affected the vegetation, which in turn drove the reorganization of the mammal community. These wetter conditions in the early Middle Pleistocene seem to have supported a larger proportion of large herbivore species relative to the number of carnivores present, resulting in a decrease in the predator–prey ratio. By about 605 ka, the faunal renewal marking the early Middle Pleistocene had come to a close, with taxa characteristic of the late Early Pleistocene (Villafranchian) having been fully replaced by modern taxa.

Palombo & Valli continue the theme of large-mammal faunal changes during the mid-Pleistocene transition by statistically analysing and comparing the French and Italian records. Southern Europe, and especially Italy because of its north–south elongation, was a refuge for northern species during Pleistocene glacial periods. In assessing waves of immigration into Italy, France is important because it acted as a route for migration from central Europe. Accordingly, the authors have used multivariate statistical analysis to compare large-mammal assemblages both within and between France and Italy. The Italian record shows a progressive faunal renewal from the late Early Pliocene and into the early Middle Pleistocene. An extinction phase began at about 1.4 Ma (the Farneta faunal unit; late Villafranchian mammal age) and the appearance of new arrivals culminated at about 1.0–0.9 Ma. Indeed, from about 1.0 to 0.9 Ma assemblages began to acquire typical ‘Quaternary’ characters. New arrivals dominated during the later Middle Pleistocene.

As with Italy, the French record shows an extinction event followed by a dispersal event, although both events seem to occur later in France, and the Middle Pleistocene dispersal event seems to have been more gradual. Clearly the transition event was a complex and possibly diachronous one, in which an extinction phase was followed by a partially overlapping dispersal event. In both areas, faunal reorganization seems to relate to a general transition in climate cyclicality, as reflected in the vegetation that changed from ‘warm’ steppe–deciduous forest alter-

nations during the Early Pleistocene, to ‘cold’ steppe–warm temperate deciduous forest after about 1 Ma (Suc *et al.* 1995b; Bertini 2003; Palombo & Valli, this volume).

The ecological structure of the French assemblages appears to parallel these vegetational changes because of the prevalence of taxa that inhabited open landscape in the early Middle Pleistocene. The analyses of French and Italian faunas also show that the ‘end-Villafranchian event’ of earlier authors was a mammal reorganization event of more profound importance than occurred at the Middle–Late Pliocene or Pliocene–Pleistocene boundaries.

By about 1 Ma, the northwest Mediterranean (Spain, France and Italy) fauna had assumed a modern aspect that marked the transition from Villafranchian to Galerian mammal ages. This transition represents an increase in large herbivores and megaherbivores, coincident with the onset of high-intensity glacial cycles (Rodríguez *et al.* 2004). A somewhat comparable shift, at c. 0.86 Ma (= early part of MIS 21), is seen in western Siberia where the fauna became dominated by the large mammals *Mammuthus trogontherii*, *Bison ex gr. priscus* and *Equus mosbachensis*, reflecting the development of open landscapes (Foronova 2005). Indeed, the steppe mammoth *Mammuthus trogontherii* first appears just before the Matuyama–Brunhes boundary (MIS 21–early MIS 19) across Eurasia (Foronova 2005; Van Kolfschoten & Markova, this volume).

O’Regan *et al.* have statistically examined the large-mammal record in Africa and the Levant during the interval 1.0–0.5 Ma. Their aim has been to determine whether there existed a large-scale turnover perhaps similar to the ‘end-Villafranchian event’ that occurred in Europe at about 0.9 Ma, and to an event in Indonesia occurring between 0.8 Ma and 0.7 Ma (van den Bergh *et al.* 2001). Major turnovers in mammalian faunas of the African continent are known for the Pliocene, but data have previously been too sparse to assess broad patterns of change across the mid-Pleistocene transition. An assessment is now feasible owing to a marked increase in published sites during the past 10 years. **O’Regan *et al.*** have included the Levant in their study because it may have served as an important route for faunal interchange. Perhaps surprisingly, the African large mammal faunas seem to have been relatively stable throughout this interval. Several new species, including the modern hippopotamus, reedbeek and black wildebeest, seem to appear at about 0.7 Ma, but these are not seen as part of a significant turnover. The continent of Africa is presently separated into Palaearctic (Levant and north Africa, but also including Europe and much of Asia) and African (sub-Saharan Africa) zoogeographical regions. **O’Regan *et al.*** show that this separation appears to originate before the mid-Pleistocene transition.

The 'end-Villafranchian event', although best known for the migration of larger animals, was also a time of important radiation in the middle-sized voles. **Van Kolfschoten & Markova** review and explain these evolutionary changes in the voles, particularly by examining the record of eastern Europe where well-dated stratigraphic sections have been studied. The authors then compare eastern Europe with the less complete record for central and western Europe. Despite the incompleteness, general similarities are apparent throughout these regions, with a clear pattern of evolution evident from the Jaramillo Subchron to the beginning of the Brunhes Chron. This is particularly illustrated by evolution in the *Allophaiomys*–*Microtus* lineage (Alexeeva & Erbaeva 2005; Markova 2005; Van Kolfschoten & Markova, this volume). Moreover, the mid-Pleistocene evolution of widely dispersed rodent species appears to have been more or less simultaneous all over Eurasia, and the same seems to be true of the steppe mammoth *Mammuthus trogontherii* which first appears just before the Matuyama–Brunhes boundary. In detail there are discrepancies, and geographical gradients cannot always be dismissed, but in general the stages in faunal evolution are clearly related to climate changes. **Van Kolfschoten & Markova** emphasize the utility of the vole record for dating and correlating mid-Pleistocene continental deposits throughout Eurasia, and note that the lowest occurrence of *Microtus* ex gr. *oeconomus* just below the Matuyama–Brunhes reversal might serve as a useful practical guide in recognizing the Early–Middle Pleistocene boundary.

The Atapuerca cave complex in northern Spain is famed for having yielded one of western Europe's oldest fossil hominins, the c. 800–850 ka *Homo antecessor* (Carbonell *et al.* 1995, 2005; Bermúdez de Castro *et al.* 1997). This site represents a very early incursion, although apparently not permanent establishment, of hominins in Europe, and there is an intense interest in understanding its climatic and palaeoecological context. The level at which *Homo antecessor* occurs has been dated at about 800–850 ka using combined ESR–U/Th analyses (Falguères *et al.* 1999) and is just below the Matuyama–Brunhes Chron boundary (Parés & Pérez-González 1999). The site has also produced the best collection of remains of the Middle Pleistocene *Homo heidelbergensis*, dated at c. 400 ka, by which time hominins were dispersed throughout Europe. The site of Atapuerca therefore represents an important archaeological transition from Early to Middle Pleistocene.

Cuenca-Bescós *et al.* have analysed the distribution of small-mammal remains across this transition at Atapuerca, allowing them to reconstruct shifts in local habitat and thereby detect landscape and other environmental changes. Small mammals represent the most abundant and diverse macrofossils at

Atapuerca. The evidence from these small mammals is combined with earlier evidence from large mammals, pollen and palaeosol analyses. The authors conclude that *Homo antecessor* lived during a warm, wet and wooded interval probably corresponding to MIS 21 to 19, and more likely to MIS 21. The small-mammal record indicates an important shift at the beginning of the Middle Pleistocene, reflecting cooler and drier conditions. The small mammals are almost exclusively from open, dry land habitats. This more open landscape may have favoured the dispersal of hominins (*Homo heidelbergensis*) across western Europe.

Hominins

The evolution and dispersal of hominins holds particular interest for the mid-Pleistocene transition. Hominins had migrated out of Africa to Dmanisi in Georgia by about 1.7 Ma (Gabunia & Vekua 1995; Gabunia *et al.* 2000), soon after the evolution of *Homo erectus*, a hominin of modern-like body pattern and increased brain size. By about 1 Ma, they had penetrated into higher northern latitudes including those of China and Europe (Dennell 2003). European evidence for earliest occupation is primarily based on lithic assemblages from Orce in south-eastern Spain at c. 1 Ma (Oms *et al.* 2000) and Monte Poggiolo in Italy at 900 ka (Milliken 1999), and from Atapuerca in northern Spain where the 850–800 ka *Homo antecessor* has been found (Falguères 2003; Cuenca-Bescós *et al.*, this volume), and Ceprano in Italy where the 900–800 ka *Homo cepranensis* has been recovered (Ascenzi *et al.* 2000; Manzi *et al.* 2001; Mallegni *et al.* 2003). Manzi (2004) considered *Homo antecessor* (from Atapuerca) and *H. cepranensis* (from Ceprano) to be closely related to each other, and puzzlingly more similar to later African lineages than European ones. Alternatively, whereas *H. cepranensis* (from Ceprano) has uncontested African affinities (Mallegni *et al.* 2003; Manzi 2004), the suggestion has been advanced that *Homo antecessor* has Asiatic affinities (Carbonell *et al.* 2005). Either scenario invokes a complex history of evolution and early migration into Europe, and elsewhere, during the mid-Pleistocene transition. Such early incursions into higher northern latitudes were rare and sporadic, and appear to be associated with the appearance of the Galerian large-mammal fauna and expansion of the northern Eurasian steppes (Dennell 2003). Consequently, the colonizing ability in general of *Homo erectus* before the early Middle Pleistocene appears to have been very limited (Dennell 2003). Of course, migration out of Africa should not be viewed as a single movement but as a series of events (Smith 2002). A sustained and widespread occupation of Europe does not appear to have begun until 600–500

ka (Roebroeks & van Kolfschoten 1994; Dennell & Roebroeks 1996; Klein 1999), and widespread evidence of hominins in western Europe occurs only after MIS 12 (c. 400 ka; McNabb, this volume).

The two final chapters, by **Petraglia** and **McNabb**, deal with hominin evolution and dispersal during the mid-Pleistocene transition, a time of increasingly variable climate and strengthening monsoon conditions. The subject, however, is fraught with sparse evidence and uncertain dates.

The dominant stone-tool industry of the mid-Pleistocene transition was the Acheulean, defined by large bifacial tools that include hand axes. These tools were not principally used as weapons, but were hand-held implements for such tasks as processing carcasses for food.

Petraglia has considered two poorly studied regions, the Arabian peninsula and the Indian subcontinent, that were evidently pathways for migrations out of Africa. The Acheulean industry appeared in Africa by about 1.6 Ma and is attributed to *Homo erectus*. The Early Acheulean is found in the Levant by 1.4 Ma (Bar-Yosef & Goren-Inbar 1993; Bar-Yosef 1998), but the earliest clear indication of the dispersal of Acheulean populations into south Asia is not until 780 ka (Bar-Yosef 1998), although mounting evidence suggests an earlier migration. **Petraglia** has reviewed stone-tool technology for the Arabian peninsula and the Indian subcontinent throughout the mid-Pleistocene transition. Perhaps surprisingly, there is maintenance of standardized bifacial technology, with little evidence for innovation and change. Gamble (1998) proposed that the selection of generic skills, such as those required by subsistence on meat, may have been responsible for such technological uniformity. However, **Petraglia's** detailed analysis of stone-tool reduction techniques and spatial distribution of tools across the landscape suggests an increasing degree of forethought and behavioural flexibility, these adaptations allowing Acheulean hominins to survive the increasingly unpredictable climatic conditions of the mid-Pleistocene transition.

McNabb has reviewed the archaeological evidence for hominin evolution and cultural change across the mid-Pleistocene transition by reference to migrations out of Africa and into Europe. Referring to geochemical data (Schefuß *et al.* 2003, this volume) he notes that contemporaneous African climates were every bit as dynamic as those in higher latitudes, and that oscillations between dry and humid climate were more extreme and of longer duration from the time of the mid-Pleistocene transition. However, as with the Arabian peninsula and Indian subcontinent, the Acheulean of Africa and Europe can be seen as a generalized hand-held processing technology with little technological change discernible over a million years of time. This does

not imply a lack of behavioural or biological change, which must have occurred in response to climatic and environmental change. Rather it demonstrates that adaptive success was based on biological and behavioural traits rather than technological dependency (Cachel & Harris 1995).

McNabb's review of the evidence for earlier *versus* later migrations out of Africa, the so-called long *versus* short chronologies, favours early migration. But it is only after about 650 ka that hominins appear with any frequency in the geological record above 45°N. Indeed, there is a dramatic shift in the hominin signal in Europe after about 650 ka. **McNabb** concludes that hominins appear to have remained generalists throughout the mid-Pleistocene transition, and it was not until after the full transition to 100-ka cyclicality that natural selection had adapted hominins to survive conditions they would meet in northern and northwestern Europe.

Notwithstanding this broad technological stasis, subtle evidence of technological innovation has been reported not only from the Arabian peninsula and the Indian subcontinent (Petraglia, this volume), but also from the Israeli site of Geshar Benot Ya'aqov, dated at 780 ka (Goren-Inbar *et al.* 2000). This same Israeli site suggests that hominins were able to control fire by at least 790 ka (Goren-Inbar *et al.* 2004; Balter 2004).

The penetration of hominins into higher northern latitudes by about 1 Ma, presumably because of the increasing availability of large herbivores as a food and commodity source, would surely have required adaptation. This could have been a critical early step in hominin evolution that led much later to increased colonizing capacity and encephalization.

Glacial history during the mid-Pleistocene

Although direct evidence of glaciation is not assessed comprehensively in the chapters of this volume, glaciation is a key aspect of the climate. An examination of evidence accumulated in the INQUA project 'Extent and Chronology of Quaternary Glaciations' (Ehlers & Gibbard 2003, 2004a, b, c, in press) demonstrates a strong consistency with other lines of evidence already noted. This compilation shows that glaciation had already occurred in many regions during the Neogene. The Early Pleistocene (2.6 to c. 0.7 Ma, *sensu* Gibbard *et al.* 2005a) included relatively few intervals that were sufficiently cold and long to allow the development of substantial ice sheets beyond mountain or high-latitude settings, such as Antarctica or northernmost North America. Only 14 Early Pleistocene cold Marine Isotope Stages are currently known to have evidence of major glaciation. They include the Pliocene–Pleistocene boundary events MIS 104,

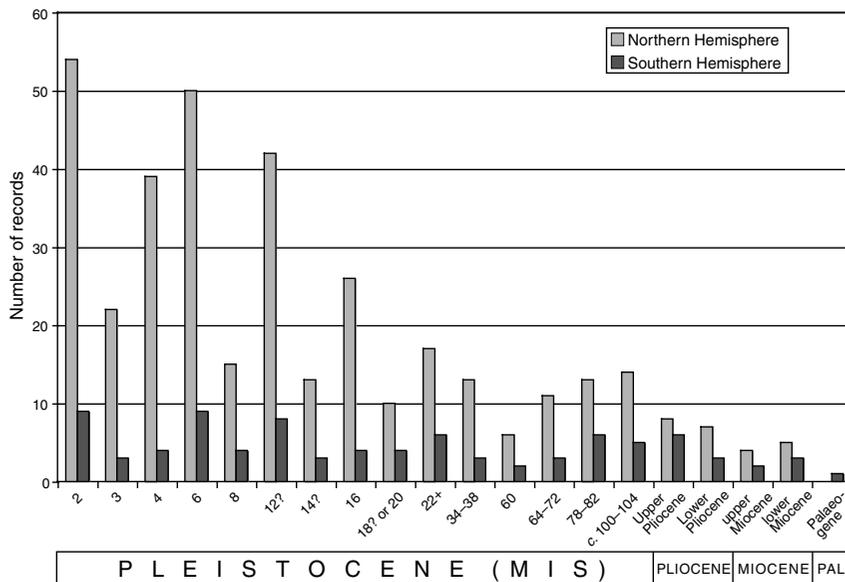


Fig. 3. Cenozoic worldwide glaciation through time, showing the relative importance of glaciations as broadly indicated by the number of literature records cited in the compilation by Ehlers & Gibbard (2004a, b, c). MIS = Marine Isotope Stages.

100 and 98, together with Early Pleistocene MIS 82, 78?, 68, 60, 58, 54, 52, 36, 34, 30? and 26. For example, in the northern Andes the record begins at 2.5 Ma, followed by an extensive glaciation at 2.2 Ma (c. MIS 82). In Patagonia, widespread lowland glaciation occurred between 2.05 and 1.86 Ma (c. MIS 78–68), followed by a major event at 1.15–1.00 Ma (c. MIS 36–34). On the mountains of east Africa, glaciation also apparently began at 2.0 Ma (MIS 68?). Figure 3 shows the relative importance of these glacial events through the Pleistocene (see also Fig. 2).

Across Europe and Siberia the first major glacial event in lowland areas is indicated by the ‘Hattem Beds’ of the Netherlands which are putatively Menapian in age (1.2–1.1 Ma; c. MIS 36–34; Fig. 1). These deposits indicate substantial glaciation of the Baltic region late in the Early Pleistocene (Doppert *et al.* 1975; Bijlsma 1981; Gibbard 1988). Glaciation of the same age is recorded off Norway by Sejrup *et al.* (2000).

However, collective evidence from all the northern continents indicates that MIS 22 (c. 870–880 ka) is the first of the ‘major’ cold events that typify glaciations of the later Pleistocene (Fig. 3). Widespread lowland glaciation is first seen throughout North America during this stage (pre-Illinoian F). Equally, glaciation of the Alpine mountains becomes established north of the Alps in Switzerland, southern Germany and in the Dolomites by MIS 22 (Muttoni *et al.* 2003). Significant glaciation is also recorded

before the Matuyama–Brunhes magnetic reversal (c. MIS 22) on east African mountains.

There is relatively scant evidence of glaciation related to MIS 20 or 18, but MIS 16 (c. 650 ka) corresponds to the Don glaciation in Europe, the most substantial glaciation yet experienced in the northern hemisphere during the mid-Pleistocene transition. This is followed by another extensive glaciation, known as the Elsterian glaciation in Europe, that is tentatively correlated to MIS 12 (c. 0.42 ka; Fig. 1).

It therefore appears from direct evidence that it is not until c. 1.2 Ma that the cold periods (glacials) are regularly cold and long enough to allow ice-sheet development on a continental scale, outside mountain or polar regions, and that the most extensive glaciations in the Quaternary (five or six intervals) were restricted to the last 900 ka (c. MIS 22 to present) in the northern hemisphere. In contrast, southern hemispheric glaciation is a much longer-established phenomenon with substantial glaciation already a regular occurrence in the Tertiary (Fig. 3). The striking development of ice sheets through the Quaternary clearly emphasizes that world-wide glaciation is in effect a northern-hemispheric phenomenon (Ehlers & Gibbard, in press).

In this context, it is interesting to note that river systems also apparently responded significantly to the increased intensity of climatic deterioration and duration of climatic events during the mid-Pleistocene transition. The increased activity has

been attributed to intensification and prolongation of periglacial regimes, characterized by frost weathering and related to coarse detrital sediment supply, in the mid-latitudes, particularly in Europe (Gibbard 1988; Gibbard & Lewin 2003). Here they are marked by increased incision and depositional terrace sequence and incision cycles in river systems that begin during the interval 1.2–0.8 Ma (Gibbard & Lewin, unpublished work), e.g. in the Lower Rhine (Meyer & Stets 2002), the Thames (Gibbard 1988), Somme and Seine (Antoine *et al.* 2003) and the Danube (Gábris & Nádor, in press).

Events during the mid-Pleistocene transition

Notable events are listed in decreasing chronological order from MIS 36 (*c.* 1.20 Ma) to MIS 13 (*c.* 540–460 ka). Most mammalian (including human) evolutionary and dispersal events are not included because too few data are available to constrain them at the resolution of Marine Isotope Stages, but they are shown on Figure 2.

MIS 36 (c. 1.20 Ma). Mudelsee & Statterger (1997) speculated that this event was a precursor to the later Pleistocene climate state; and it coincides with a major intensification of the east Asian winter monsoon system (Heslop *et al.* 2002). Ice sheets developed on a continental scale in MIS 36–34 for the first time during the mid-Pleistocene (see above), and in the southeastern South Atlantic glacial–interglacial contrasts in ocean circulation strengthened after 1.2 Ma (Diekmann & Kuhn 2002).

MIS 35 and 34 (c. 1.17–1.11 Ma). A long-term shift in the New Zealand vegetation occurs at MIS 35 and is succeeded by the coldest climate phase which corresponds to MIS 34 (Byrami *et al.*, this volume), a probable time of major glaciation (see above).

MIS 31 (c. 1.07 Ma). An extremely warm event is recorded both in Antarctica (Scherer *et al.* 2004) and New Zealand (Byrami *et al.*, this volume).

MIS 30 (c. 1.05 Ma). A marked increase occurs in the contrast between glacial–interglacial stages in the tropical African vegetation (Dupont *et al.* 2001).

MIS 25 (c. 940 ka). The lowest occurrence of the warm-water calcareous nannofossil *Gephyrocapsa* sp. 3 characterizes the boundary of the marine Calabrian and Ionian stages in Italy, which Cita & Castradori (1994, 1995) proposed to serve also as the Early–Middle Pleistocene Subseries boundary.

MIS 24 (c. 910 ka). The detection of nearly pure AABW in the western North Atlantic implies a strong reduction in the North Atlantic thermohaline circulation (Ferretti *et al.*, this volume), and a similar stagnation of bottom-water circulation is also detected in the South Atlantic (Schmieder *et al.*

2000). MIS 24 represents an important step towards a state of long-term increased global ice volume, which Mudelsee & Statterger (1997) have determined as being between 942 and 892 ka.

MIS 23–22 (c. 900–880 ka). The MIS 23–22 transition is characterized by the highest occurrence of the calcareous nannofossil *Reticulofenestra asanoi*, which is synchronous across low to mid-latitudes globally and related to severe cooling in MIS 22 (Raffi 2002; Reale & Monechi, this volume).

MIS 22 (c. 880–870 ka). MIS 22 represents the first of the major cold events that characterizes the later Pleistocene (see above). Thicker and sandier loess deposits in northern Eurasia are also indicative of severe palaeoclimatic conditions (Heslop *et al.* 2002; Dodonov, this volume). As with MIS 24, a strong reduction in North Atlantic thermohaline circulation is implied also for this interval (Ferretti *et al.*, this volume). By deduction, MIS 22 must represent the first interval of substantial sea-level fall due to the glacioeustatic effects of ice build-up during this glaciation, although the magnitude of this sea-level fall has not yet been documented.

MIS 22–21 (c. 870–850 ka). The MIS 22–21 transition represents a distinctive loess–palaeosol boundary in northern Eurasia that Dodonov (this volume) proposes as representing the Early–Middle Pleistocene Subseries boundary. The MIS 22–21 transition coincides with the base of the following units: the ‘Cromerian Complex’ Stage of northwestern Europe, the Tiraspolian mammal Stage of Russia (Dodonov, this volume), the Petropavlovian (sub)stage of the Russian Plain, and the Neopleistocene Subseries of the Russian Plain (Zhamoïda 2004; Gibbard & van Kolfschoten 2004). In the southwestern Pacific from MIS 21 time there seems to have been a slowing of deep-water currents (Haywood *et al.*, this volume).

MIS 21 to early 19 (c. 850–770 ka). This interval, which includes the earliest members of the Tiraspolian Russian mammal stage, represents the spread of large mammals including the steppe mammoth *Mammuthus trogontherii* across northern Eurasia (Foronova 2005), and is marked also by the first appearance of the vole *Microtus ex gr. oeconomus* (van Kolfschoten & Markova, this volume).

MIS 16 (c. 650–620 ka). This stage corresponds to the Russian Plain Donian Stage (Fig. 1) which is characterized by the extensive Don glaciation (see above). Ice-rafted debris also increased substantially in the Nordic Sea during this time (Helmke *et al.* 2005). In the South Atlantic, glacial–interglacial contrasts became more severe after 650 ka (Diekmann & Kuhn 2002).

MIS 15 (c. 620–550 ka). A particular group of cylindrical benthic foraminifera had a global extinction (the ‘*Stilostomella* extinction’) during MIS 16–15, which is linked to the onset of slower deep-

water currents and decreasing bottom oxygenation (Hayward *et al.*, this volume).

MIS 13 (c. 540–460 ka). Various events mark what has been considered a terminal phase in the mid-Pleistocene transition, including the development of a distinctive sediment facies in the South Atlantic at c. 540–530 ka, unusual variations in $\delta^{18}\text{O}$ values in the Indian Ocean during MIS 13.2 and 13.3, an anomalous sapropel layer in the Mediterranean Sea dated at 528–525 ka, and exceptionally high magnetic susceptibilities in Chinese palaeosol–loess deposits at about 500 ka (Schmieder *et al.* 2000, and references therein). Moreover, MIS 13 precedes an extensive glaciation tentatively assigned to MIS 12, known in Europe as the Elsterian, Okian or Anglian and in North America as Pre-Illinoian B (Figs 2 & 3). An acme of the benthic foraminifera *Abitodentrix pseudothalmanni* between MIS 15 and 12 (c. 620–420 Ma) in the SW Pacific appears to signal an increase in food supply and lower dissolved oxygen, and occurs at a time of slower deep-water currents in the area (Hayward *et al.*, this volume).

The Matuyama–Brunhes boundary

For practical purposes, the Early–Middle Pleistocene Subseries boundary is typically drawn at the Matuyama–Brunhes palaeomagnetic Chron boundary (Richmond 1996), which coincides with the middle of MIS 19 (Bassinot *et al.* 1994; Channell *et al.* 2004; Capraro *et al.*, this volume). This boundary was astrochronologically dated at 780 ka by Shackleton *et al.* (1990) who used the Ice Volume Model of Imbrie & Imbrie (1980) as a tuning target for marine isotope data from ODP Site 677 in the eastern Pacific. Because polarity reversal stratigraphy was not resolved for ODP Site 677, Shackleton *et al.* (1990) used DSDP Sites 552 and 607 in the North Atlantic where both isotope data and palaeomagnetic records were available. More recently, Channell *et al.* (2004) obtained high-resolution magnetostratigraphic and isotope data from ODP Sites 984 and 983 situated in deep-sea drift deposits south of Iceland, where sedimentation rates are very high. Channell *et al.* (2004) tuned these records to the Ice Volume Model, and determined that for ODP Site 984 the Matuyama–Brunhes polarity transition has an estimated duration of 7 ka, with an onset at 777 ka and a mid-point at c. 773.5 ka. For Site 983, the duration is about 5 ka and the mid-point of the transition is 772.5 ka (see also Channell & Kleiven 2000). Older previous estimates based on marine sediments (e.g. Shackleton *et al.* 1990) might be a function of their lower rates of deposition, as the lock-in depth of remanence acquisition will then represent a greater amount of time (Channell *et al.* 2004).

The Matuyama–Brunhes palaeomagnetic bound-

dary is also affected by problems of displaced magnetic remanence in loess–palaeosol sequences of northern Eurasia, causing it to occur in sediments assigned to MIS 20 when it should in fact be in MIS 19 (Zhou & Shackleton 1999; Dodonov, this volume). Despite the fact that chemical weathering of continental sediments may result in secondary Brunhes-age overprinting, the Matuyama–Brunhes boundary was found to be the most easily recognized chronostratigraphic marker in Australian continental sediments (Pillans 2003).

A recommendation for the Early–Middle Pleistocene boundary

A chronostratigraphic boundary serves to create a hierarchical subdivision of a time scale, and therefore should be ideally defined by an event of appropriate magnitude. It must also be an effective tool for correlation in order to be of practical use. The challenge in selecting a stratigraphic position for the Early–Middle Pleistocene Subseries boundary lies in the fact that the mid-Pleistocene transition is characterized not by any single event, but by a series of palaeoenvironmental events of varying expression and importance. These events overlie what Mudelsee & Statterger (1997), from statistical analysis of marine oxygen isotope records, have called a multiple-transition phenomenon.

Previous recommendations for the stratigraphic position of the Early–Middle Pleistocene Subseries boundary have been: (1) the boundary of the Calabrian and Ionian marine stages in Italy (Cita & Castradori 1994, 1995; although later considered premature by Castradori 2002); (2) the MIS 22–21 transition corresponding to the base of the Tiraspolian mammal stage of Russia, the base of the Petropavlovian (sub)stage of the Russian Plain and of the Neopleistocene Subseries of the Russian Plain (Gibbard & van Kolfschoten 2004; Zhamoida 2004; Dodonov, this volume); and (3) the Matuyama–Brunhes palaeomagnetic Chron boundary (Richmond, 1996; Pillans 2003). In addition, Pillans (2003) considered the Jaramillo Subchron as a possible position for the boundary, but for practical reasons rejected it in favour of the Matuyama–Brunhes boundary.

The Calabrian–Ionian boundary, occurring within MIS 25, is problematic in that it is biostratigraphically expressed in Italy primarily by the lowest occurrence of the warm-water calcareous nannofossil *Gephyrocapsa* sp. 3, a datum now shown to be strongly diachronous (Raffi 2002; Reale & Monechi, this volume).

A boundary associated with the cold MIS 22 is appealing on the basis that this is probably the single most profound time of environmental change within

the Early–Middle Pleistocene transition. This is particularly true of the MIS 22–21 transition which sees a rapid change from very low to high $\delta^{18}\text{O}$ values. In many continental Eurasian deposits this pronounced MIS 22–21 transition is easily recognized (Dodonov, this volume). However, the MIS 22 event may be less useful for correlating in continental records in the southern hemisphere, given that the glaciation represented by MIS 22 is principally a northern hemisphere phenomenon. In the marine realm, the boundary would be better suited to the MIS 23–22 transition based on the extinction of the calcareous nannofossil *Reticulofenestra asanoi* (Raffi 2002; Reale & Monechi, this volume). A practical difficulty here arises in that calcareous microfossils become diachronous and poorly represented at high latitudes.

Our own recommendation, based on new information on the Early–Middle Pleistocene transition (including that contained in the chapters of this volume), is to endorse that of Richmond (1996) and Pillans (2003) in placing the Early–Middle Pleistocene Subseries boundary at, or close to, the Matuyama–Brunhes boundary. It is worth noting in this regard that the traditional use of biostratigraphic datums as primary marker events for establishing boundaries is no longer seen as a requirement. Other events, such as palaeomagnetic reversals and isotopic signals, are now also considered acceptable, but should be accompanied by a diagnostic and diverse fossil record (Remane *et al.* 1996).

Placing the boundary at or near the Matuyama–Brunhes reversal seems to have overwhelming practical benefits even though MIS 19, in which the Matuyama–Brunhes boundary occurs, is not an exceptionally significant palaeoenvironmental event. Such a placement would allow the boundary to be recognized in both marine and terrestrial deposits, even when isotope stratigraphy was not available (which is the case with most marine cores), and at high as well as lower latitudes. The Matuyama–Brunhes boundary also has the advantage of being reasonably close to MIS 22, which is indeed probably the most significant climatic shift within the Early–Middle Pleistocene transition.

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