22 • The Pleistocene and Holocene Epochs

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This chapter focuses on the major subdivisions and events in the terrestrial sequences of the Pleistocene and Holocene, with correlations to the marine record. Current proposals for formal subdivision are outlined.

22.1 PLEISTOCENE SERIES

22.1.1 Evolution of terminology

The classification and interpretation of the youngest stratigraphic sequences, variously known as Pleistocene, Holocene, and Quaternary, have been, and still are, a matter of debate. During the first two decades of the nineteenth century, many of the sequences were attributed to the biblical flood (the "Diluvial" theory). This theory could account for unconsolidated sediments that rested unconformably on "Tertiary" rocks and capped hills, and that commonly contained exotic boulders and the remains of animals, many still extant. This origin for the "Diluvium" was accepted by most eminent geologists of the time, including Buckland and Sedgwick.

Floating ice had frequently been seen transporting exotic materials, providing an explanation for the transport of the boulders, and reinforcing the Diluvial theory. This explanation lead to adoption of the term "drift" to characterize the sediments. However, geologists working in the Alps and northern Europe had been struck by the extraordinary similarity of the "drift" deposits and their associated landforms to those being formed by modern mountain glaciers. Several observers such as Perraudin, Venetz-Sitten, and de Charpentier proposed that the glaciers had formerly been more extensive, but it was the paleontologist Agassiz who first advocated that this extension represented a time that came to be termed the Ice Age by Goethe.

After having convinced Buckland and Lyell of the validity of his Glacial Theory in 1840, Agassiz's ideas became progressively accepted. The term *Drift* became established for the widespread sands, gravels, and boulder clays thought to have been deposited by glacial ice. Meanwhile, Lyell had already proposed the term *Pleistocene* in 1839 for the post-Pliocene period closest to the present. He defined this period on the basis of its molluscan faunal content, the majority of which are still extant. However, the term Quaternary (*Quaternaire or Tertiaire récent*) had already been proposed in 1829 by Desnoyers for marine sediments in the Seine Basin (Bourdier, 1957, p. 99) – although the term had been in use from the late eighteenth century.

Both terms – Pleistocene and Quaternary – have become synonymous with the Ice Age. However, unlike the Pleistocene concept, the span of the Quaternary included Lyell's original "Recent," later named *Holocene* by the Third International Geological Congress (IGC) in London in 1885. The term *Holocene* (meaning "wholly recent") refers to the percentage of living organisms and was defined by Gervais (1867–1869) "for the post-diluvial deposits approximately corresponding to the post-glacial period" (Bourdier, 1957, p. 101). The Holocene period was originally considered to represent a fifth era or *Quinquennaire* (Parandier, 1891), but this division was deemed to be "excessive;" details are given in Bourdier (1957) and de Lumley (1976).

Because the terms Primary, Secondary, and Tertiary have been abandoned, the continued use of Quaternary is regarded by some stratigraphers as somewhat archaic. Alternative terms of *Anthropogene* (extensively used in the former USSR) or *Pleistogene* (suggested by Harland *et al.*, 1990) have been proposed, but neither found favor. Other proposals place the Holocene in the Pleistocene epoch as a stage (cf. the Flandrian: see below).

The "Quaternary" is traditionally considered to be the interval of oscillating climatic extremes (glacial and interglacial episodes) that was initiated at about 2.6 Ma, therefore encompassing the Holocene and Pleistocene epochs and Gelasian stage of late Pliocene. A formal decision on its chronostratigraphic status is pending, as advocated by ICS and INQUA (Pillans, 2004).

22.1.2 The Pliocene–Pleistocene boundary and the status of the Quaternary

In 1948 at the IGC in London, an attempt was made to identify a basal-Pleistocene boundary. The requirement that it be located in exposed marine sediment led its placement near or at the base of the Calabrian strata in southern Italy (a stage introduced by Gignoux in 1910). This horizon was close to the first indication of climatic deterioration in Italy that took place after the deposition of the Italian Neogene (Oakly, 1950). The initial Calabrian boundary was thought to be marked by the first appearance of the cold-water mollus indicators *Arctica islandica* and *Hyalinea baltica* (Sibrava, 1978), but Ruggieri and Sprovieri (1979) showed that *Hyalinea baltica* appears slightly later. Moreover, he argued for the suppression of Calabrian and its replacement by Santernian, together with a revision of the rest of the sequence. Subsequently, various sections in southern Italy competed for the position of stratotype. Haq *et al.* (1977) correlated the boundary with the top of, or slightly above, the short-lived Olduvai magnetostratigraphic event at 1.8 Ma.

The GSSP for the Pliocene–Pleistocene boundary and the beginning of the Pleistocene was placed by a joint INQUA and ICS working group (IGCP Project 41) near the top of the Olduvai subchron (1.8 Ma) and approved by ICS in 1983. The GSSP is at Vrica (39° 32′ 18.61″ N, 17° 08′ 05.79″ E), approximately 4 km south of Crotone in the Marchesato Peninsula, Calabria, southern Italy (Aquirre and Pasini, 1985; Bassett, 1985). Stratigraphic details are given in Chapter 21 (Section 21.1.4)

The decision to assign the base-Pleistocene GSSP was "*isolated from other more or less related problems, such as ... the status of the Quaternary within the chronostratigraphic scale*" (Aguirre and Pasini, 1985). Many "Quaternary" workers, especially those working with terrestrial and climatic records, now favor defining "Quaternary" as beginning significantly before the base-Pleistocene GSSP. As a result, the status (as of 2004) and chronostratigraphic rank of Quaternary has not been established, and different options for formally defining "Quaternary" are being considered (e.g. Ogg, 2004; Pillans, 2004; Pillans and Naish, 2004).

The London 1948 IUGS recommendations included the notion that the base-Pleistocene boundary should be placed at the first evidence of climatic cooling. However, the Vrica GSSP boundary level is not the first severe cold climate oscillation of the late Cenozoic. It can be argued that the first severe cold climate takes place at a stratigraphic position equivalent to the base of the Dutch terrestrial Praetiglian Stage, and some earth scientists studying Quaternary strata in northern Europe tend to begin their Pleistocene at this level. Since the record on land is highly fragmentary and difficult to correlate, eastern Europeans have their own terminology in which the Eopleistocene follows the Pliocene (cf. Section 22.2 below). These alternative sequence terminologies have been included in Fig. 22.1.

This older level corresponds to the Gauss/Matuyama magnetic epoch boundary (2.6 Ma) and the base of the Pliocene Gelasian Stage (Rio *et al.*, 1998). An equivalent level in marine sediments occurs at Monte San Nicola in Sicily and can be easily correlated with Marine Isotope Stage (MIS) 104 in the ocean sediments (see discussion in Suc *et al.*, 1997). The event is clearly defined in the marine oxygen isotope stratigraphy and coincides with the first major influx of ice-rafted debris into the middle latitude of the North Atlantic (Shackleton *et al.*, 1984; Shackleton, 1997; Partridge, 1997a). The fossil mammalian record also shows changes that are obvious near the Gauss/Matuyama reversal. Opposing views on the position for the Pliocene–Pleistocene boundary have been discussed by Van Couvering (1997) and by Partridge (1997b).

22.1.3 Division of the Pleistocene

Two major types of subdivisions have been proposed for the Pleistocene Series. A standard subdivision at stage level has been advocated by workers based on sections in elevated shallow-marine sediments in Italy (see Chapter 21 and Fig. 22.1). Earth scientists concerned with terrestrial and to a lesser extent shallow-marine sequences have adopted regional subdivision schemes. The regional schemes have found favor despite the difficulties of world-wide correlation. In these schemes, larger, subseries- (sub epoch) scale units have been adopted and are advocated here.

A quasi-formal tripartite subdivision of the Pleistocene into Lower, Middle, and Upper has been in use since the 1930s. The first usage of the terms Lower, Middle, and Upper Pleistocene was at the second International Quaternary Association (INQUA) Congress in Leningrad 1932 (Woldstedt, 1962), although they may have been used in a loose way before this time. Their first use in a formal sense in English was by Zeuner (1935, 1959) and Hopwood (1935) and was based on characteristic assemblages of vertebrate fossils in the European sequence.

The desire to make these units identifiable world-wide led the INQUA Commission on Stratigraphy/ICS Working Group on Major Subdivision of the Pleistocene (Richmond, 1996) to place the Lower–Middle boundary at the Brunhes–Matuyama magnetic reversal epoch boundary; the "Toronto Proposal" of Richmond (1996). Unfortunately, it is less easy to define the Middle–Upper boundary in the same fashion and therefore it seems expedient to consider it equivalent to the base of marine isotope stage 5 (MIS 5) following the long-established convention that the basal boundary of the Upper Pleistocene corresponds with that of the last interglacial stage (proposed at INQUA Commission on Stratigraphy working group meeting, Berlin 1995, unpublished). This proposal naturally follows from the acceptance that MIS 5, substage e, is the ocean equivalent of the terrestrial northwest European Eemian Stage interglacial (Shackleton, 1977).

The proposals of the INQUA/ICS Working Group on Major Subdivision of the Pleistocene (Richmond, 1996) can be summarized as follows:

> It was proposed that the initial Middle Pleistocene boundary be placed at the Matuyama–Brunhes magnetic polarity reversal. The reversal has not been dated directly by radiometric controls. It is significantly older than the Bishop Tuff (revised K–Ar age 738 ka; Izett, 1982), and the estimated K–Ar age of 730 ka assigned to the reversal by Mankinen & Dalrymple (1979) is too young. In Utah, the Bishop volcanic ash bed overlies a major paleosol developed in sediments that record the Matuyama–Brunhes reversal (Eardley *et al.*, 1973). The terrestrial geologic record is compatible with the astronomical age of 788 ka assigned to the reversal by Johnson (1982).

[Note: The age of this Matuyama–Brunhes reversal is estimated as 781 ka in the astronomical-tuned time scale, see Chapter 20.] The initial Late Pleistocene boundary, placed arbitrarily at the beginning of MIS 5 (at the midpoint of Termination II or the MIS 6–5 transition), is not dated directly. It was assigned provisional ages of 127 ka by CLIMAP Project members (1984) and 128 ka by SPECMAP Project members (Ruddiman and Mcintyre, 1984), based on uranium series ages of the MI substage 5e high eustatic sea-level stand. However, more recent re-evaluation of the boundary indicates that following historical precedent in northwest Europe, the Middle–Upper Pleistocene subseries boundary should correspond to the Saalian–Eemian Stage boundary rather than to the boundary in marine isotope records which is not coeval (see below). The former is positioned at the boundary stratotype of the latter at 63.5 m below surface in the Amsterdam Terminal borehole (52 E 0913: 52 22 45 N; 4 54 52 E). This parastratotype locality is also the Eemian Stage unit-stratotype (Cleveringa *et al.* 2000; van Kolfschoten and Gibbard, 2000; van Leeuwen *et al.*, 2000). Both the stage and the stage boundary are recognized on the basis of multidisciplinary biostratigraphy, the boundary being placed at the expansion of forest tree pollen above 50% of the total pollen assemblage, the standard practice in northwest Europe (Gibbard, 2003). The Saalian–Eemian Stage boundary is identified at 126 kyr in deep-sea sediment off Iberia by Sanchez-Goñi *et al.* (1999) and Shackleton *et al.* (2002).

Independently, groups of workers have advocated a subdivision based on "standard stages" comparable in scale to those defined for the Neogene, as already noted. Of particular importance is the scheme that has been developed for shallow-marine sequences in southern Italy (Fig. 22.1) and summarized in Chapter 21.

Other shallow-marine sequences, such as that from New Zealand, have also been developed. In the former USSR, and particularly in European Russia, the Pleistocene is divided into the Eopleistocene, equivalent to the Early Pleistocene subseries, and the Neopleistocene, equivalent to the Middle and Late Pleistocene subseries (Anonymous, 1982, 1984; Krasnenkov *et al.*, 1997). The most recent proposal for a revised stratigraphical scheme for the last 1 Ma in the Eastern European Plain is given by Shik *et al.* (2002).

22.2 TERRESTRIAL SEQUENCES

In contrast to the rest of the Phanerozoic, the uppermost Cenozoic has a long-established tradition of sediment sequences being divided on the basis of represented climatic changes, particularly sequences based on glacial deposits in central Europe and mid-latitude North America. This approach was adopted by early workers for terrestrial sequences because it seemed logical to divide till (glacial diamicton) sheets and non-glacial deposits or stratigraphical sequences into *glacial (Glaciation)* and *interglacial* periods, respectively (cf. West, 1968, 1977; Bowen, 1978). In other words, the divisions were fundamentally lithological. The overriding influence of climatic change on sedimentation and erosion has meant that, despite the enormous advances in knowledge during the last century and a half, climate-based classification has remained central to the subdivision of the succession. Indeed, the subdivision of the modern ocean sediment isotope stage sequence is itself based on the same basic concept (see below). It is this approach which has brought Quaternary geology so far, but at the same time causes considerable confusion to workers attempting to correlate sequences from enormously differing geographical and, thus, environmental settings. This is because of the great complexity of climatic change and the very variable effects of the changes on natural systems.

The recognition of climatic events from sediments is an inferential method and by no means straightforward. Sediments are not unambiguous indicators of contemporaneous climate, and other evidence such as fossil assemblages, characteristic sedimentary structures (including periglacial

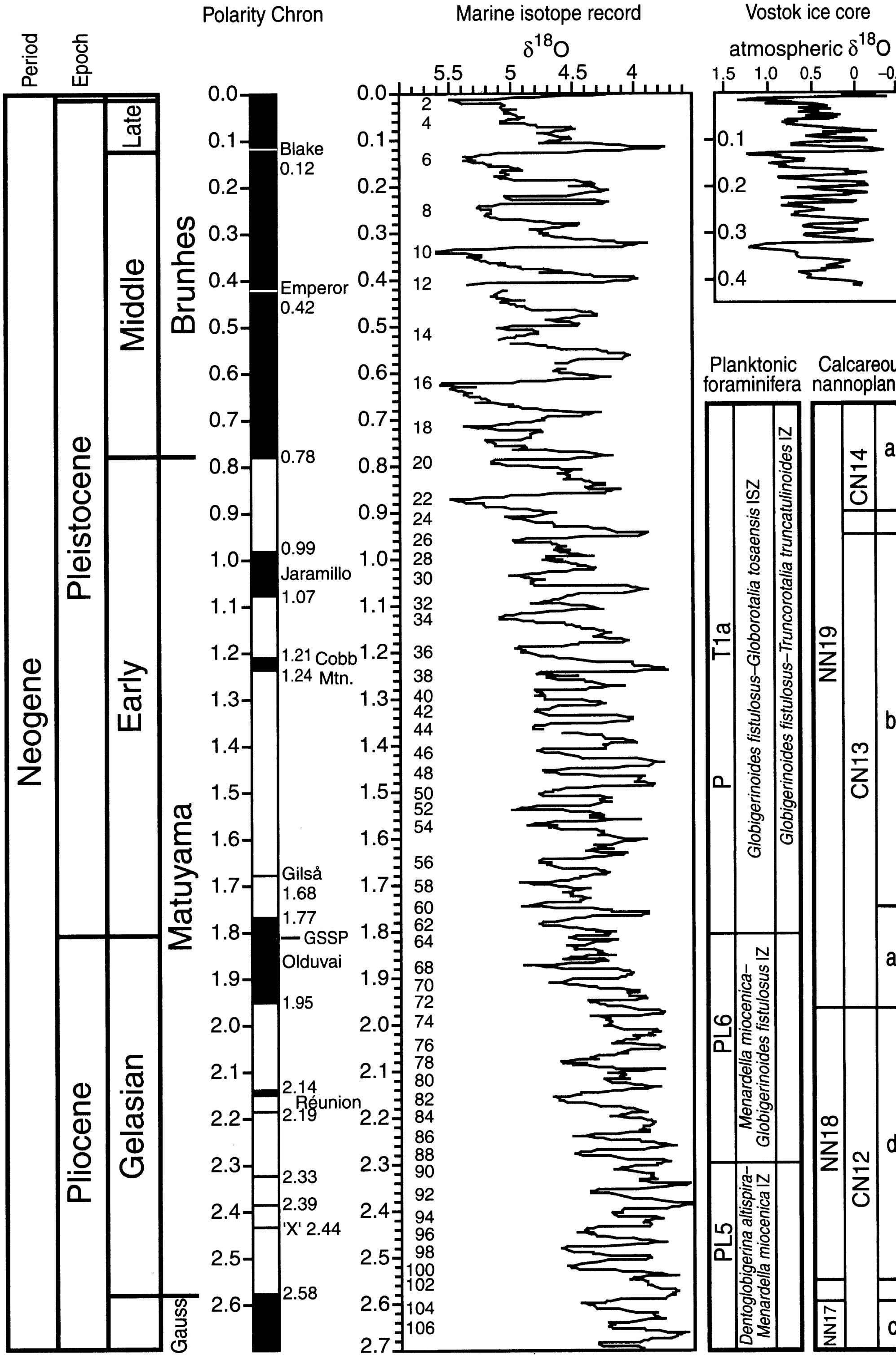


Figure 22.1 The Pleistocene-Holocene and upper Pliocene time scale. The Global Stratotype Section and Point (GSSP) for the base of the Pleistocene Epoch is indicated. The calibration of the geomagnetic polarity time scale is from oceanographic data collected and processed by S. J. Crowhurst (Delphi Project 2002) and modified from Funnell (1996). The composite marine $\delta^{18}O$ isotope sequence is from the Delphi Project (database at http://131.111.44.196 at Godwin Laboratory, University of Cambridge, UK). The micropaleontological zonations are from Berggren *et al.* (1995a). The atmospheric oxygen isotope curve from the Vostok ice coring is fro[m] Petit *et al.* (2001, Vostok Ice Core Data for 420, 000 Years, IGBP PAGES/World Data Center for Paleoclimatology Data Contribution Series #2001-076, at NOAA/NGDC Paleoclimatology Program, Boulder, CO, USA; original reference is Petit *et al.*, 1999

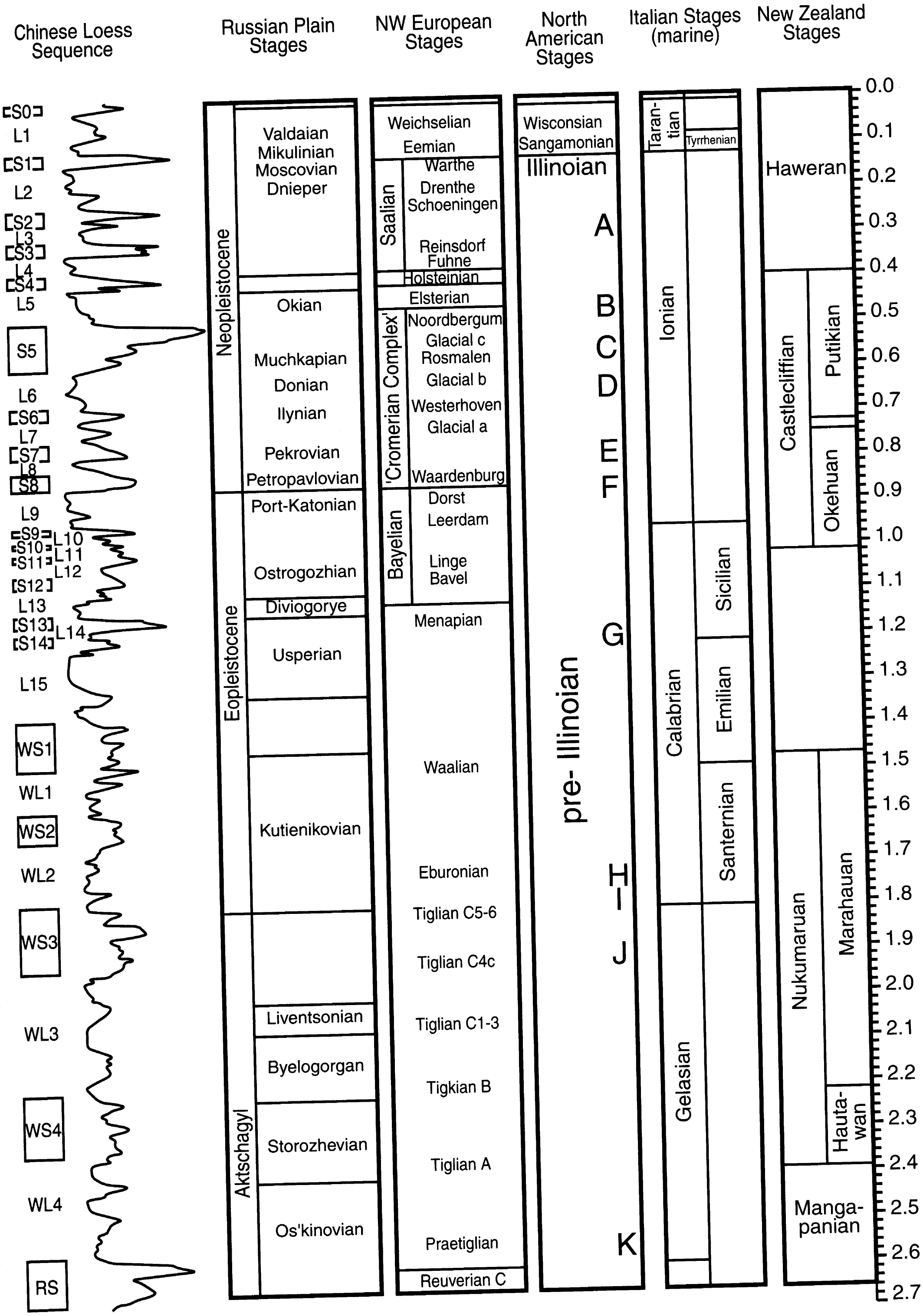


Figure 22.1 (*cont.*) This curve can be downloaded from www.ngdc.noaa.gov/paleo/icecore/antarctica/vostok/vostok_isotope.html. The Chinese loess sequence is a magnetic susceptibility signature from Luochuan (An Zhisheng *et al.*, 1990); S, soil horizon; L, loess interval; W and R, numbered older successions. The Quaternary continental successions were compiled from Zagwijn (1992), de Jong (1988), Tzedakis *et al.* (1997). European Russia succession were compiled from the Stratigraphy of the USSR: Quaternary System (Anonymous, 1982, 1984), Krasnenkov *et al.* (1997), and Shik *et al.* (2002). North America successions from Richmond (unpublished). The stage successions based on shallow-marine sequences of Italy (van Couvering, 1997) and New Zealand (Pillans, 1991) are included.

structures) or textures, soil development, and so on must be relied upon wherever possible to illuminate the origin and climatic affinities of a particular unit. Local and regional variability of climate complicates this approach in that sequences are the result of local climatic conditions, yet there remains the need to equate them with a global scale. For at least the first half of the twentieth century the preferred scale was that developed for the Alps at the turn of the century by Penck and Brückner (1909).

For the Alps, the sequence in increasing age is:

Würm Glacial (Würmian)
Riss–Würm Interglacial
Riss Glacial (Rissian)
Mindel–Riss Interglacial
Mindel Glacial (Mindelian)
Günz–Mindel Interglacial
Günz Glacial
Donau–Günz Interglacial
Donau Glacial
?Biber Glacial

For northern Europe, the sequence (with increasing age) is (see also Figs. 22.1 and 22.2):

Flandrian (i.e. Holocene), i.e. present interglacial, up to and including the present day.
Weichselian Glacial
Eemian Interglacial
Saalian Glacial Complex
Holsteinian Interglacial
Elsterian Glacial
Cromerian Complex
Bavelian Complex
Menapian Glacial
Waalian Interglacial
Eburonian Glacial
Tiglian Complex
Praetiglian Glacial

More recently, the northern European scheme tends to become replaced by the marine isotope record (Bowen, 1978). Today the burden of correlation lies in equating local, highly fragmentary, yet high-resolution terrestrial and shallow-marine sediments on the one hand, with the potentially continuous, yet comparatively lower resolution ocean isotope sequence on the other. Both are required, particularly because the ocean record sums a global situation while terrestrial sequences are dependent on local and regional conditions of climate. In addition, both need to be considered in relation to the extremely high-resolution (potentially annual) records of the ice-core sequences (see below).

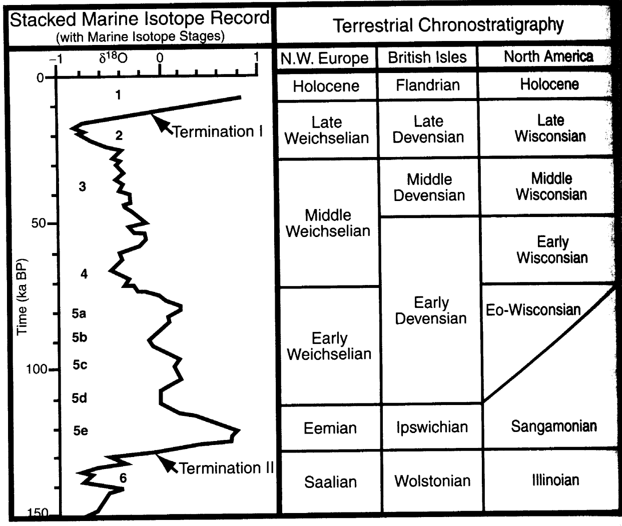


Figure 22.2 Marine and continental chronostratigraphy for the past 150 kyr. The stacked marine oxygen isotope sequence and associated stages are from Martinson *et al.* (1987), and the terrestrial climato- and chronostratigraphical divisions in northwest Europe and North America are modified from Lowe and Walker (1997).

Because stages are the fundamental working units in chronostratigraphy they are considered appropriate in scope and rank for practical intra-regional classification (Hedberg, 1976). However, the definition of stage-status chronostratigraphical units, with their time-parallel boundaries placed in continuous successions wherever possible, is a serious challenge especially in terrestrial Quaternary sequences. In these situations, boundaries in one region may be time-parallel but over greater distances problems may arise as a result of diachroneity. It is probably correct to say that only in continuous sequences which span entire interglacial–glacial–interglacial climatic cycles can an unequivocal basis for the establishment of stage events using climatic criteria be truly successfully achieved. There are the additional problems which accompany such a definition of a stage, including the question of diachroneity of climate changes themselves and the detectable responses to those changes. For example, it is well known that there are various "lag" times of geological responses to climatic stimuli. Thus, in short, climate-based units cannot be the direct equivalents of chronostratigraphical units because of the time-transgressive nature of the former. This distinction of a stage in a terrestrial sequence from that in a marine sequence should be remembered.

Before the impact of the ocean core isotope sequences, an attempt was made to formalize the climate-based stratigraphical terminology in the American Code of Stratigraphic Nomenclature (American Commission, 1961), where so-called geologic-climate units were proposed. Here a geologic-climate unit is based on an inferred widespread climatic episode defined from a subdivision of Quaternary rocks (American Commission, 1961). Several synonyms for this category of units have been suggested, the most recent being climatostratigraphical units (Mangerud *et al.*, 1974) in which an hierarchy of terms is proposed. These units are neither referred to in the standard stratigraphic codes by Hedberg (1976) nor Salvador (1994), and are not followed in New Zealand, but are included in the local Norwegian Code (Nystuen, 1986). Boundaries between geologic-climate units were to be placed at those of the stratigraphic units on which they were based.

The American Commission (1961) defines the fundamental units of the geologic-climate classification as follows:

> A *Glaciation* is a climatic episode during which extensive glaciers developed, attained a maximum extent, and receded. A *Stadial* ('Stade') is a climatic episode, representing a subdivision of a glaciation, during which a secondary advance of glaciers took place. An *Interstadial* ('Interstade') is a climatic episode within a glaciation during which a secondary recession or standstill of glaciers took place. An *Interglacial* ('Interglaciation') is an episode during which the climate was incompatible with the wide extent of glaciers that characterise a glaciation.

In Europe, following the work of Jessen and Milthers (1928), it is customary to use the terms interglacial and interstadial to define characteristic types of non-glacial climatic conditions indicated by vegetational changes (Table 22.1); interglacial to describe a temperate period with a climatic optimum at least as warm as the present interglacial (Holocene, Flandrian: see below) in the same region, and interstadial to describe a period that was either too short or too cold to allow the development of temperate deciduous forest or the equivalent of interglacial-type in the same region.

In North America, mainly in the USA, the term interglaciation is occasionally used for interglacial (cf. American Commission, 1961). Likewise, the terms stade and interstade may be used instead of stadial and interstadial, respectively (cf. American Commission, 1961). The origin of these terms is not certain but the latter almost certainly derive from the French word *stade* (m), which is unfortunate since in French *stade* means (chronostratigraphical) stage (cf. Michel *et al.*, 1997), e.g. *stade isotopique marin* = marine isotope stage.

It will be readily apparent that, although in longstanding usage, the glacially based terms are very difficult to apply outside glaciated regions, i.e. most of the world. Moreover, as Suggate and West (1969) recognized, the term Glaciation or Glacial is particularly inappropriate since modern knowledge indicates that cold rather than glacial climates have tended to characterize the periods intervening between interglacial events. They therefore proposed that the term "cold" stage (chronostratigraphy) be adopted for "glacial" or "glaciation." Likewise, they proposed the use of the term "warm" or "temperate" stage for interglacial, both being based on regional stratotypes. The local nature of these definitions indicates that they cannot necessarily be used across great distances or between different climatic provinces (Suggate and West, 1969; Suggate, 1974; West, 1968, 1977) or indeed across the terrestrial–marine facies boundary (see below). The use of mammalian biostratigraphic data, in particular the evolution of voles, offers the possibility of long-distance correlations between local assemblages. In addition, it is worth noting that the subdivision into glacial and interglacial is mainly ap plied to the Middle and Late Pleistocene.

Both interglacial and glacial, or temperate and cold stages, have been subdivided into substages and zones. This is achieved in interglacial stages using paleontological, particularly vegetational, assemblages. The cyclic pattern of interglacial vegetation that typifies all known temperate events in Europe was developed as a means of subdividing, comparing, and, therefore, characterizing temperate events by West (1968, 1977) and Turner and West (1968). In this scheme, temperate (interglacial) event sequences are subdivided into four substages: pre-temperate, early temperate, late temperate, and post-temperate. Finer-scale zonation schemes are also commonly in use throughout Europe and the former USSR (Table 22.1).

Late Middle- and Late-Pleistocene glacial stages have been divided on various bases, but in the northern hemisphere the division is based on a combination of vegetation, lithology, and occasionally pedological evidence, often resulting in an unfortunate intermixture of chrono-and climatostratigraphical terminology. The last glacial stage (Weichselian, Valdaian, Devensian, Wisconsinian) has particularly been divided into three or four substages (Early, Middle or Pleni-glacial, Late Weichselian, etc., and Late-glacial), using geochronology (^{14}C). Boundaries are defined at specific dates, especially in the last 30 ka (Table 22.1).

An independent record of Late Pleistocene and Holocene climatic changes has been derived from $\delta^{16}O/\delta^{18}O$ ratios in cores through the Greenland and Antarctic ice sheets (Johnsen *et al.*, 1972; Dansgaard *et al.*, 1993) and from other areas. In the

Table 22.1 *Examples of chronostratigraphical substage divisions of interglacial (temperate) stages and related cold (glacial) stages of the Middle and Late Pleistocene*[a,b]

	Chronostratigraphical substages[c]				Vegetational aspect	Characteristic vegetation
	Cromerian (~750ka)	Hoxnian (~350ka)	Ipswichian (~125ka)	Flandrian (post-10ka)		
Cold stage	e An	e Wo	e De		early glacial	herb-dominated
Temperate stage	Cr IV	Ho IV	Ip IV		post-temperate	birch–pine forest
Temperate stage	Cr III	Ho III	Ip III	Fl III	late-temperate	mixed deciduous–coniferous forest
Temperate stage	Cr II	Ho II	Ip II	Fl II	early-temperate	deciduous forest
Temperate stage	Cr I	Ho I	Ip I	Fl I	pre-temperate	birch–pine forest
Cold stage	l Be	l An	l Wo	l De	late-glacial	herb-dominated

[a] Modified after West (1968) and West & Turner (1968).
[b] For the Holocene (Flandrian), correlations with the zones of Godwin (1975) are also indicated.
[c] e, early; l, late; Be, Beestonian; An, Anglian; Wo, Wolstonian; De, Devensian.

past three decades, the drilling of cores into ice sheets in various parts of the world has revolutionized our records of detailed climatic change. Boreholes sunk particularly in Antarctica and in Greenland, and more recently into smaller ice shields in tropical mountain areas, have provided spectacularly unrivalled sequences which allow annual resolution of climatic events.

From a stratigraphical point of view, it is the recognition of patterns of a wide range of climatically controlled parameters that provide potentially very high-resolution correlation tools. Detailed patterns arise from determination of aerosol particles, dust, trace elements, spores, or pollen grains, etc., that have fallen onto the ice surface and become incorporated into the annual ice layers. They include, for example, dust from wind activity or volcanic eruptions. Trace gases such as carbon dioxide or methane can be trapped in air bubbles within ice crystals, etc. These gases can themselves be analyzed to determine their stable isotope content, particularly that of $\delta^{18}O$ and the sequences obtained can be compared to those from ocean sediments. In addition, naturally and artificially occurring radioactive isotopes present in the ice layers can be used to provide an independent chronology for dating the ice core sequences.

Most cores span the Holocene and provide an annually resolved sequence for the interglacial. However, the Vostok core in Antarctica spans a period of up to 400 kyr (Fig. 22.1; Petit *et al.*, 1999), but it is in Greenland that detailed cores from the Greenland Ice Core Project (GISP) and the Greenland Ice Sheet Project (GISP) have been obtained that provide a sequence that extends at least as far back as the Last Interglacial (Eemian). These sequences have revolutionized our understanding of patterns and rates of global climate changes, as well as the interlinking of the ocean–atmosphere–terrestrial systems (see Lowe and Walker, 1997, for a more detailed discussion).

22.3 OCEAN SEDIMENT SEQUENCES

Because the span of Quaternary time includes our own, a different order of discrimination is possible and different methods are rapidly developing. The principal development in the Pleistocene time scale depends on the regularity of the climatic cycle that was discovered around 1875 by Croll and developed, especially, by Milankovitch. This approach was not taken seriously by Quaternary geologists until Zeuner (1945), Emiliani (1955), and Evans (1971) were among those to recalculate and relate the astronomical parameters, testing, for example, 42 and 100 kyr cycles against other phenomena, such as the newly established oxygen isotope curve from the oceans. The first rigorous treatment using wide-ranging techniques was by Hays et al. (1976). Isotope studies from the bottom sediments of the world's oceans since then have indicated as many as 52 Late Cenozoic glacial ages and that the continental evidence is so incomplete as compared with the oceanic sequences that terrestrial glacial–interglacial stratigraphy in future must depend on the ocean record for chronological foundation as outlined in Chapter 21.

The marine oxygen isotope scale makes use of the fact that, when continental ice builds up as a result of global cooling and sea level is lowered, the ice is depleted in $\delta^{18}O$ relative to the ocean water, leaving the ocean water enriched in $\delta^{18}O$. The oxygen isotope composition of calcareous foraminifera and coccoliths, and of siliceous diatoms, varies in direct proportion to that of the water (cf. Shackleton and Opdyke, 1973, for discussion of the limitations of isotope stratigraphy). The

16 stages of Emiliani (1955, 1966) obtained from Caribbean and Atlantic sediment cores were extended to 22 by Shackleton and Opdyke (1973) after analysis of the V28–238 core from the equatorial Pacific. Subsequently another equatorial Pacific core, V28–239 (Shackleton and Opdyke, 1973), and an Atlantic core (Van Donk, 1976) extended the reconstruction of glacial–interglacial variability through the Pliocene–Pleistocene boundary. Later developments under the flag of the Deep Sea Drilling Program resulted in the extension of the isotope record into the Early Pleistocene and Pliocene (Shackleton and Hall, 1989; Ruddiman *et al.*, 1987; Raymo *et al.*, 1989). The sequence shown in Fig. 22.1 is a combination of measurements from cores V19–30, ODP677, and ODP846 (Crowhurst, 2002). The isotope stages recognized in core V28–238, from the eastern Pacific (Shackleton and Opdyke, 1973), are generally regarded as the "type" for the Late Pleistocene, while those defined in cores ODP 677 and 846 are those for the Middle Pleistocene and Pliocene, respectively (Shackleton and Hall, 1989; Shackleton *et al.*, 1995).

As regards nomenclature, the events differentiated in isotope sequences are termed marine isotope stages (MIS); this term is preferred by palaeoceanographers to the previously widely used oxygen isotope stages (OIS). This is because of the need to distinguish the isotope stages recognized from those identified from ice cores or speleothem sequences (Shackleton, pers. comm.). The stages are numbered from the present-day (MIS 1) backwards in time, such that cold-climate or glacial events are assigned even numbers and warm or interglacial (and interstadial) events are given odd numbers. Individual events or substages in marine isotope stages are indicated either by lower-case letters or in some cases by a decimal system, thus MIS 5 is divided into warm substages 5a, 5c, and 5e, and cold substages 5b and 5d, or 5.1, 5.3, 5.5, and 5.2 and 5.4, respectively, named from the top downwards. This apparently unconventional top-downwards nomenclature originates from Emiliani's (1955) original terminology and reflects the need to identify oscillations down cores from the ocean floor.

The biggest problem with climate-based nomenclature, like the marine isotope stratigraphy, is where the boundaries should be drawn. Ideally, the boundaries should be placed at a major climate change. However, this is problematic because of the multifactorial nature of climate. But since the events are only recognized through the responses they initiate in depositional systems and biota, a compromise must be agreed. Although there are many places at which boundaries could be drawn, in principle in ocean-sediment cores they are placed at midpoints between temperature maxima and minima. The boundary points thus defined in ocean sequences are assumed to be globally isochronous, although a drawback is that temperatures may be very locally influenced and may also show time lag. The extremely slow sedimentation rate of ocean-floor deposits and the relatively rapid mixing rate of oceanic waters argue in favour of the approach. Attempts to date these MIS boundaries are now well established (Martinson *et al.*, 1987).

22.4 LAND–SEA CORRELATION

In recent years it has become common to correlate terrestrial sequences directly with those in the oceans. This arises from the need felt to correlate local sequences to a regional or global time scale, mentioned above, occasioned by the fragmentary and highly variable nature of terrestrial sequences. The realization that more events are represented in the deep-sea, and indeed ice-core, sequences than were recognized on land, together with the growth in geochronology, has often led to the replacement of locally established terrestrial scales. Instead, direct correlations of terrestrial sequences to the global isotope scale are advanced, as advocated, for example, by Kukla (1977). The temptation to do this is understandable, but there are serious practical limitations to this approach (cf. Schlüchter, 1992).

In reality, there are very few means of directly and reliably correlating between the ocean and terrestrial sediment sequences. Direct correlation can be achieved using markers that are preserved in both rock sequences, such as magnetic reversals, radiometric dating, or tephra layers, and, rarely, fossil assemblages (particularly pollen). However, this is normally impossible over most of the record and in most geographical areas. Thus these correlations must rely totally on direct dating or less reliably on the technique of "curve matching;" a widely used approach in the Quaternary. The latter can only reliably be achieved where long, continuous terrestrial sequences are available, such as long lake profiles (e.g. Tzedakis *et al.*, 1997), but even here it is not always straightforward (e.g. Watts *et al.*, 1995) because of overprinting by local factors. Moreover, the possibility of failure to identify "leads-and-lags" in timing by the matching of curves is very real. Loess–soil sequences, such as those in China (An Zhisheng *et al.*, 1990; Fig. 22.1), have also provided very important and locally reliable correlative sequences, but they are also restricted by the need to have continuous or at least quasi-continuous sedimentation without subsequent disturbance. In discontinuous sequences, which typify land and shelf environments, correlations with ocean-basin sequences are potentially unreliable, in the absence of fossil groups distributed across the facies boundaries or potentially useful markers.

In recent years, the growth of stratigraphy recognized from short-duration, often highly characteristic, events has led to attempts to use these features as a basis for correlation. This event stratigraphy (e.g. Lowe *et al.*, 1999), typically deposition of a tephra layer or magnetic reversals, can also include geological records of other potentially significant-type events such as floods, tectonic movements, changes of sea level, climatic oscillations or rhythms, and the like. Such occurrences, often termed "sub-Milankovitch events," may be preserved in a variety of environmental settings and thus offer important potential tools for high- to very high-resolution cross-correlation.

Of particular importance are the so-called "Heinrich Layers" which represent major iceberg-rafting events in the North Atlantic Ocean (Heinrich, 1988; Bond *et al.*, 1992; Bond and Lotti, 1995). These detritus bands *can potentially* provide important lithostratigraphical markers for intercore correlation in ocean sediments and the impact of their accompanying sudden coolings (Heinrich Events) *may be* recognizable in certain sensitive terrestrial sequences (see summary in Lowe and Walker, 1997). Similarly, the so-called essentially time-parallel periods of abrupt climate change termed "terminations" (Broecker and van Donk, 1970), seen in oxygen isotope profiles, can also be recognized on land as sharp changes in pollen assemblage composition or other parameters, for example, where sufficiently long and detailed sequences are available, such as in long lake cores (cf. Tzedakis *et at.*, 1997). However, their value for correlation may be limited in high-sedimentation-rate sequences because these "terminations" are not instantaneous but have durations of several thousand years (Broecker and Henderson, 1998). These matters essentially concern questions of resolution and scale.

Of greater concern for the development of a high-resolution terrestrial stratigraphy is the precise recognition and timing of boundaries or events from the marine isotope stages on land, and indeed vice versa. Until very recently this has not been perceived as a problem since it has generally been assumed that boundaries identified using a variety of proxies on land are precisely coeval with those seen in ocean sediments. Yet we know that different proxies respond at different rates and in different ways to climate changes and these changes themselves may be time-transgressive. Moreover, changes in ocean currents, sea level, wind patterns, tectonics, and so on, may further complicate local responses reflected in coastal regions. This has been forceably demonstrated recently by work off Portugal by Sanchez-Goñi *et al.* (1999) and Shackleton *et al.* (2002) where the MIS 6/5 boundary has been shown to have not been coeval with the Saalian–Eemian Stage boundary on land, as previously assumed. The same point concerns the MIS 1–2 boundary, which pre-dates the Holocene–Pleistocene (Flandrian–Weichselian) boundary by some 2000–4000 years. Thus high-resolution land sequences and low-resolution marine sequences must be correlated with an eye to the detail since it cannot be assumed that the boundaries recognized in different situations are indeed coeval.

A different, yet equally relevant, example is the situation that occurs during MIS 3. This period was generally interpreted by ocean sediment workers as being of an "interstadial" character because it showed a decrease in $\delta^{18}O$ relative to the preceding and following MISs. Moreover, today it is known to include considerable climatic variability of a lower amplitude cyclic character (e.g. Bond cycles: Bond *et al.*, 1992; Bond and Lotti, 1995). However, on land, particularly in northwest Europe, this period is not wholly interstadial (*sensu* Jessen and Milthers, 1928; see above), but is characterized by a variable, predominantly cold, climate that is interrupted by short minor climatic ameliorations (interstadials), as evidenced by the occurrence of frozen-ground features and of biota indicating warmer and/or arid conditions (e.g. Guthrie and van Kolfschoten, in press).

Notwithstanding these problems of detail, which will no doubt be further resolved as new evidence becomes available, it is now generally possible to relate the onshore–offshore sequences fairly reliably at a coarse scale at least for the last glacial–interglacial cycle (Upper Pleistocene). This was first proposed by Woillard (1978), but is now well established (e.g. Tzedakis *et al.*, 1997). Beyond, things are very much more complicated. Witness, for example, the longstanding disagreements over the nature and duration of the northwest European Saalian Stage, already referred to above (Litt and Turner, 1993). Questions of whether the Holsteinian–Hoxnian temperate (interglacial) stage relates to MIS 9 or 11, and thus the immediately preceding Elsterian–Anglian glacial stage (cf. Zagwijn, 1992) to MIS 10 or 12 (Turner, 1998; de Beaulieu and Reille, 1995) or even MIS 8, and precisely how many interglacial-type events occur within the Saalian, leave much potential for inaccuracy that cannot be resolved by "counting-backwards" methods. In the absence of reliable dating these correlations represent little more than a matter of belief. The situation becomes significantly more difficult in the early Middle Pleistocene (Turner, 1998), in spite of the fact that there is the important marker of the Brunhes–Matuyama magnetic reversal event with which to correlate. In the Early Pleistocene, where the dominant cyclicity of the climate signal is the

42 kyr periodicity, not the 100 kyr periodicity of the later Quaternary, ocean–terrestrial sequence correlation is virtually impossible at present, except close to the major magnetic reversal boundaries.

To add further to these problems the phenomenon of delayed preservation of the magnetic signal that has been detected in some terrestrial sediments, in particular in Chinese loess (Zhou and Shackleton, 1999). This therefore suggests that it is questionable whether magnetic reversals can be used as reliable markers for inter-regional correlation in high-resolution, high sedimentation-rate sequences.

Nevertheles, dating through astronomical (and subastronomical) cycles is clearly a geochronological tool of considerable future potential, already realized in respect of the ocean and ice-core sequences (e.g. Björck *et al.*, 1998), and of singular importance to understanding rates of process operation on land once the problems of cross-facies correlation have been overcome. Perhaps the way forward should be to date fixed events – probably magnetic reversals or major climatic events – as accurately as possible, then use the astronomical cyclicity to provide a finer-scale chronology. In future, it is important that this scheme be phased-in to run in parallel and perhaps eventually to replace the fundamentally palaeontologically tuned scheme that has served stratigraphical geology so well in the past.

22.5 PLEISTOCENE–HOLOCENE BOUNDARY

In the previous edition of this book it was stated that "this boundary was thought to correspond to a climatic event around 10 000 radiocarbon years before present (BP)." At the time, the boundary was considered likely to be standardized in a varved lacustrine sequence in Sweden (cf. Mörner, 1976). It was originally proposed at the Eighth INQUA Congress in Paris in 1969 and was subsequently accepted by the INQUA Holocene Commission in 1982 (Olausson, 1982). The climatic amelioration on which this boundary is identified is well established in a variety of sediments, particularly in northern Europe and North America. In Scandinavia, it corresponds to the following boundaries: European Pollen Zones III–IV, the Younger Dryas–pre-Boreal and Late Glacial–postglacial (Mörner, 1976; Mangerud *et al.*, 1974). However, this boundary definition was not formally ratified by the ICS. If it is finally defined precisely at 10 000 (^{14}C yr BP), it would be the first stratigraphic boundary later than the Proterozoic to be defined chronometrically. This statement remains broadly accurate, albeit seen in the light of the abundant ice core evidence, now available from both hemispheres (see below), it appears highly likely that the boundary should actually occur at 11.5 kyr (ice-accumulation years: Dansgaard *et al.*, 1993). In spite of the potential accuracy of less than five years that can now be achieved in the annually laminated ice cores, it is possible that the basal boundary stratotype of the Holocene (or potentially a parastratotype) will be placed in an annually laminated lacustrine sequence in western Germany (Litt *et al.*, 2001). This is because here it can be identified to the nearest year and can be precisely radiocarbon-dated. Moreover, the sequence has yielded an excellent, easily correlated fossil record, principally pollen and other freshwater microfossils, which facilitates regional biostratigraphical correlation.

22.6 HOLOCENE SERIES

Holocene is the name for the most recent interval of Earth history and includes the present day (see Section 21.2.4). It is generally regarded as having begun 10 000 radiocarbon years, or the last 11 500 calibrated (i.e. calender) years, before present (i.e. 1950). The term "Recent" as an alternative to Holocene is invalid and should not be used. Sediments accumulating or processes operating at present should be referred to as "modern" or by similar synonyms.

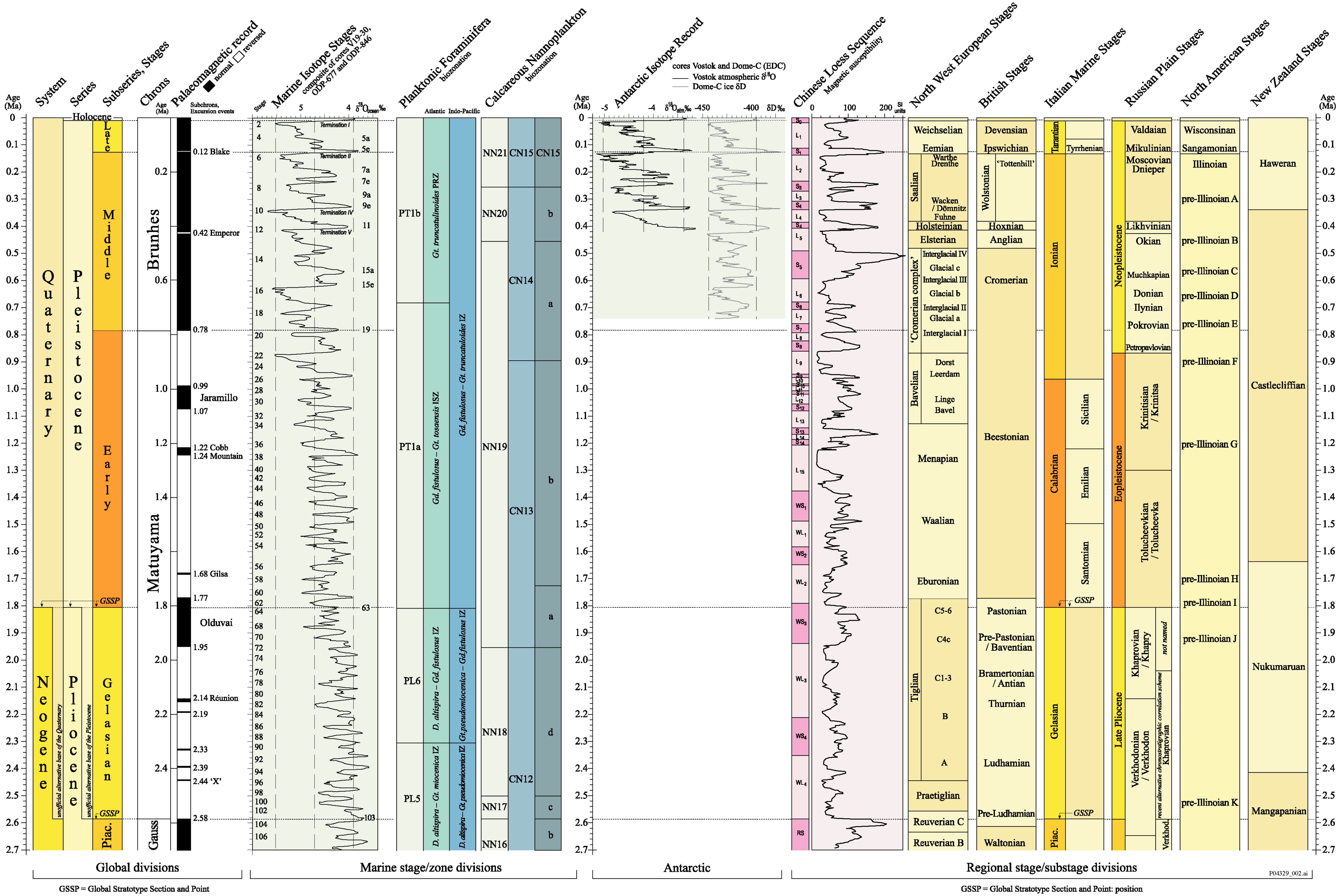
The term Flandrian, derived from marine transgression sediments on the Flanders coast of Belgium (Heinzelin and Tavernier, 1957), has often been used as a synonym for Holocene (Fig. 22.2). It has been adopted by authors who consider that the last 10 000 years should have the same stage status as previous interglacial events and thus be included in the Pleistocene. In this case, the latter would thus extend to the present day (cf. West, 1968, 1977, 1979; Hyvärinen, 1978). This usage, although advocated particularly in Europe, has been losing ground in the last two decades (cf. Lowe and Walker, 1997, p. 16).

Various zonation schemes have been proposed for the Holocene (Flandrian) Epoch. The most established is that of Blytt and Sernander (cf. Lowe and Walker, 1997), which was developed for peat bogs in Scandinavia in the late nineteenth to earliest twentieth centuries. Their terminology, based on interpreted climatic changes, comprised, in chronological order, the pre-Boreal, Boreal, Atlantic, sub-Boreal, and sub-Atlantic. This scheme was refined by the Swede von Post and others, using pollen analysis throughout Europe. Today this terminology remains in use in northern Europe, although it has been largely displaced by absolute chronology, particularly ^{14}C. Dating has shown that the biostratigraphically defined zone boundaries are diachronous (cf. Godwin, 1975). An attempt to fix these boundaries to precise dates was proposed for northern Europe by Mangerud *et al.* (1974).

In prehistoric times, as well as later, climatic events have largely served to identify the divisions elaborated by modern ^{14}C, other dating techniques, tephrachronology, and dendrochronology as well as successively by archeology and human history. Using these techniques Holocene time be divided into ultra-high-resolution divisions. For examp recent developments indicate that cyclic patterns of clim change of durations as short as 200 yr can be differentia and potentially used for demonstrating equivalence in p sequences.

Global chronostratigraphical correlation table for the last 2.7 million years

v. 2004



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