

QUATERNARY GLACIATIONS AND CHRONOLOGY

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4.1 INTRODUCTION

Since the recognition in the mid-19th century that glaciers had been considerably more extensive than at present, the Quaternary (Pleistocene and Holocene series) has been synonymous with glaciation of the mid-latitudes. Today evidence from both the land and ocean-floor sediment sequences demonstrates that the major continental glaciations occurred repeatedly over what are now temperate regions of the Earth's surface. The climatic variation that characterizes the Earth's climate during the late Cenozoic, and indeed before, is controlled by variations of the Earth's orbit around the Sun that therefore controlled the receipt of solar energy at the Earth's surface. These Milankovitch variations, named after the Serbian mathematician, their discoverer, are responsible for the cyclic climate changes that characterize the Quaternary and indeed much of Earth's history. One of the most critical ways in which the Milankovitch climate cyclicity is expressed is through the development of 'Ice Ages' or periods when glaciation extended across large areas of the Earth (Ehlers et al., 2016).

4.2 PRELUDE TO THE QUATERNARY

The onset of the recent ice age dates back much further than the Quaternary. An overall, long-term trend of cooling global climates occurred throughout the Tertiary Period. The opening of the North Atlantic Ocean, accompanied by the simultaneous closure of the east–west-aligned Tethys Ocean, characterize this period. This development led to profound palaeogeographical rearrangements and initiated major changes in the circulation patterns of the ocean–atmosphere system. In particular, the drift of Antarctica to its position over the South Pole and the restriction and isolation of the Arctic Ocean are jointly responsible for the marked zonation of climate since transfer of energy to the polar regions became progressively more limited with the drifting of the continents and the associated uplift of young mountain belts, including the Alpine-Himalayan chain, through the period.

One of the consequences of the continent/ocean distribution on the Earth's surface was glaciation. Apart from some limited activity in the Eocene both in the Southern and Northern Hemisphere (Ingólfsson, 2004; Tripathi et al., 2005; Stickley et al., 2009), significant glaciation began in the late Oligocene (c. 35 Ma) in eastern Antarctica (Ingólfsson, 2004). It was followed by mountain glaciation

BOX 4.1 THE QUATERNARY PERIOD/SYSTEM AND ITS SUBDIVISION

The Quaternary Period/System, formally defined as beginning 2.58 million years ago, is represented in marine and terrestrial sediment sequences worldwide. Its base, and that of the Pleistocene Epoch/Series, were lowered in 2009 to the Global Boundary Stratotype Section and Point (GSSP) at Monte San Nicola, Sicily, Italy (Gibbard and Head, 2010; Gibbard et al., 2010; Head and Gibbard, 2015b), which had been established for the Gelasian Stage (formally the uppermost stage of the Pliocene) in 1996 (Rio et al., 1998). Thus the Gelasian now represents the lowest stage of the Quaternary, and is succeeded by the Calabrian Stage, which in 2011 was defined by the former Pleistocene GSSP at Vrica, Calabria, southern Italy, with an age of 1.80 Ma (Cita et al., 2012).

The only other GSSP currently ratified for the Quaternary defines the base of the Holocene Epoch/Series. It was established in 2008, and is defined in the North Greenland Ice Core Project's NGRIP2 ice core, with an age of 11,700 calendar years before 2000 (Walker et al., 2009).

That no other GSSP is located in an ice core emphasizes the creativity required to meet the high-precision correlations demanded of Quaternary stratigraphy. At the time of writing only three GSSPs have been defined for the Quaternary, leaving much work to be done for the future. In particular, the Lower (Early)-Middle Pleistocene and Middle-Upper (Late) Pleistocene subseries boundaries, will be formally defined in the near future (Fig. 4.1).

through the Miocene (23–5.3 Ma) in Alaska, Greenland, Iceland, and Patagonia, and later in the Pliocene (5.3–2.6 Ma) in the Alps, the Bolivian Andes, and possibly in Tasmania.

From the Neogene, glacially derived ice-rafted debris (IRD) is found in ocean-sediment cores from the North Atlantic region, including the Barents Sea (Moran et al., 2006), and areas adjacent to Norway, north and southeast Greenland, Iceland, and northern North America, and in the Southern Ocean off Antarctica (De Schepper et al., 2014), bearing witness to the early glaciations (Box 4.1).

4.3 GLACIATION DURING THE QUATERNARY

The beginning of the Quaternary, the Early Pleistocene (2.6–0.8 Ma) was characterized by climatic fluctuations dominated by the 41 ka orbital precession cycle, during which relatively few cold periods were sufficiently cold and long to allow the development of substantial ice sheets. Only 14 of the 41 cold stages of that period currently show evidence of major glaciation. They include the Plio-Pleistocene boundary events Marine Isotope Stage (MIS) 104, 100, and 98, together with Early Pleistocene MIS 82, 78, 68, 60, 58, 54, 52, 36, 34, 30, and 26 which reach $\delta^{18}\text{O}$ ocean water ratios of c. 4.6–5‰.

Extensive ice-rafting, an indication that glaciers had reached sea-level, is found from the earliest cold stage (2.6–2.4 Ma) in both the North Atlantic and North Pacific oceans (cf. Haug et al., 2005). Synchronous glaciation of the Greenland, Scandinavian, and North American regions began at 2.72 Ma (MIS G6), a dramatic increase in IRD at 2.7 Ma in the western sub-Arctic Pacific indicates apparent circum-Arctic expansion of ice sheets (Bartoli et al., 2005), potentially relate to the closure of the Panama Isthmus (Kameo and Sato, 2000). Further significant ice-sheet expansions in the Nordic Seas and North Atlantic between 2.7 and 2.4 Ma, correspond to cold MIS 104, 100, 98, and 96 (Kleiven et al., 2002). Changes also occurred in the Southern Hemisphere, where synchronous peaks in IRD at 2.9–2.7 Ma off the Antarctic Peninsula and southeast Greenland indicate the increase in intensity of glaciation was a global event (St. John, 2008).

It was not until the transition in dominant orbital cyclicity to the 100 ka cycles, that began c.1.2 Ma and was fully established by about 800 ka ('Early-Middle Pleistocene transition', EMPT), that the cold periods (glacials) were regularly cold and long enough to allow ice-sheet development on a continental scale, outside the polar regions (Head and Gibbard, 2005, 2015a). However, it was during MIS 22 (c. 870–880 ka) that the first of the 'major' cold events reached critical values of c. 5.5 or above $\delta^{18}\text{O}$ ocean‰ equivalent to substantial ice volumes that typify glaciations of the later Pleistocene (i.e., MIS 16, 12, 10, 6, 4–2). Potentially therefore, it is likely that there were a minimum of 20 periods during which extensive glaciation could have developed during the last 2.6 million years, with the most extensive (c. 5–6 periods) being limited to the last 900 ka (Fig. 4.1).

Precisely where these glaciations occurred and how far they extended is very difficult to determine, given that the remnants of less extensive earlier glaciation tends to be obliterated and mostly removed by later, more extensive ice advances. Although this is so in all terrestrial areas, it is especially in mountain regions that the preservation potential of older sequences rapidly diminishes with time and subsequent glaciation.

Examination of the frequency of glaciation through the Cenozoic (Fig. 4.2) indicates that glaciation in the Southern Hemisphere was established first in Antarctica and southern South America, where it occurred continually from the early Neogene to the present day.

By contrast, Northern Hemisphere glaciation, although initially somewhat restricted, increased markedly at the beginning of the Quaternary, increasing again in frequency in the latest Early Pleistocene and reaching very high levels in the Middle-Late Pleistocene. Whilst this pattern is not unexpected, the striking increase in ice-sheet size through the Quaternary clearly emphasizes that, based on the distribution of land mass in higher latitude land mass, worldwide glaciation is in effect a northern-hemispheric phenomenon.

Examination of the evidence accumulated in the INQUA project *Extent and Chronology of Quaternary Glaciations* (Ehlers and Gibbard, 2004a,b,c; Ehlers et al., 2011a; Fig. 4.3), supported by other published sources, demonstrates the current state of knowledge. When examining the resulting tables (Tables 4.1–4.3) it is important to bear in mind that the stratigraphical control beyond the range of radiocarbon dating is weaker than might be desired. This is particularly a problem outside Europe where biostratigraphy is less well developed and the sheer extent of the unexplored regions makes future discoveries likely. Thus the presentation below can only be a first step toward a comprehensive overview. The correlations applied here are based on the Global Correlation Chart (Gibbard and Cohen, 2008) (Box 4.2).

4.4 PLIO-PLEISTOCENE GLACIATION

Evidence of glaciation is widespread from throughout the Quaternary and indeed the Neogene in the Northern Hemisphere. The longest sequences are restricted to Alaska, and the adjacent North-West Territories of Canada which, together with Greenland and the Rockies, preserve evidence of glaciation from the Neogene to the present. In northern Canada and Alaska, the oldest till and accompanying IRD in marine settings, dates from the Early Miocene, with regionally widespread glaciation occurring in the Pliocene and regularly throughout the Pleistocene (cf. Haug et al., 2005). In adjacent British Columbia a comparable sequence is found, particularly in the north. Similarly, in Greenland and Iceland glaciation began in the Miocene, occurring regularly through

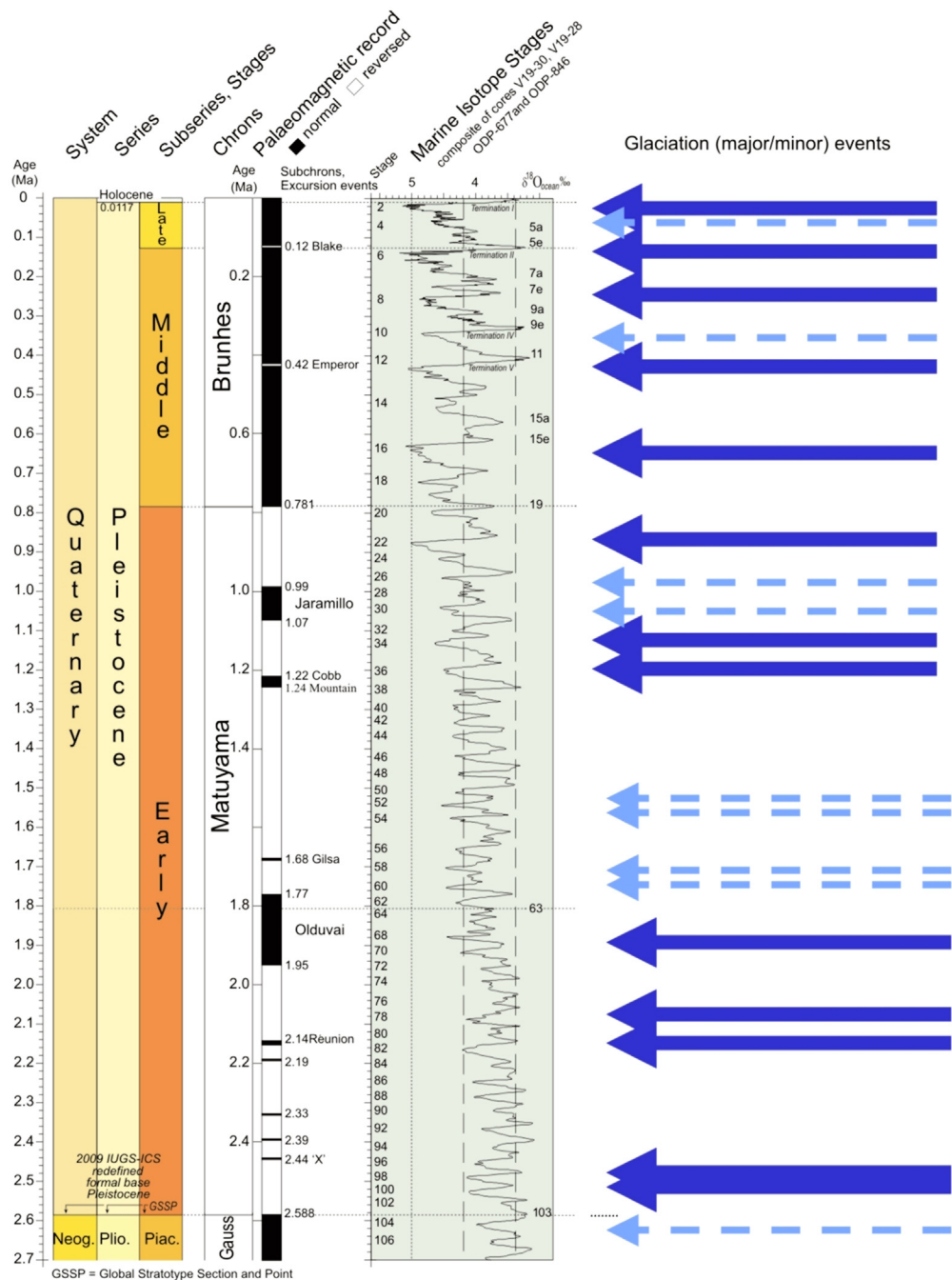


FIGURE 4.1

Distribution of major and minor glaciations through Quaternary time.

Extract from Global Correlation table by Gibbard, P.L., Cohen, K.M., 2008. Global chronostratigraphical correlation table for the last 2.7 million years. Episodes 31, 243–247.

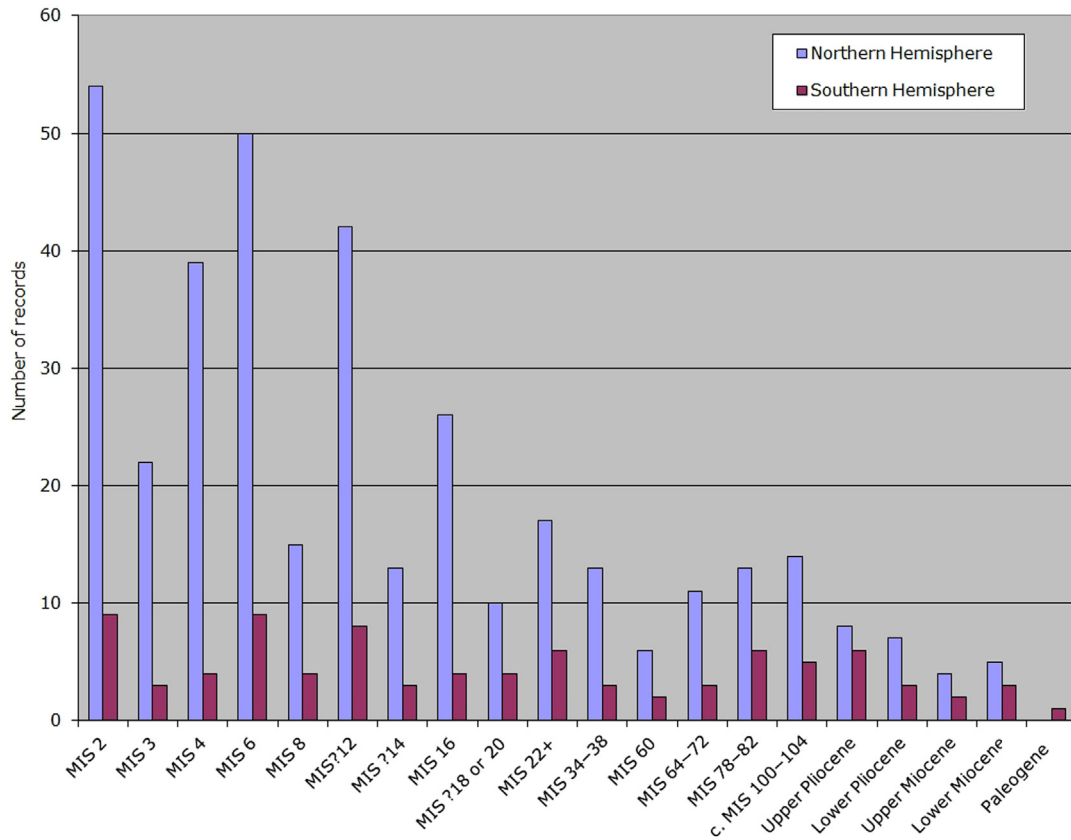
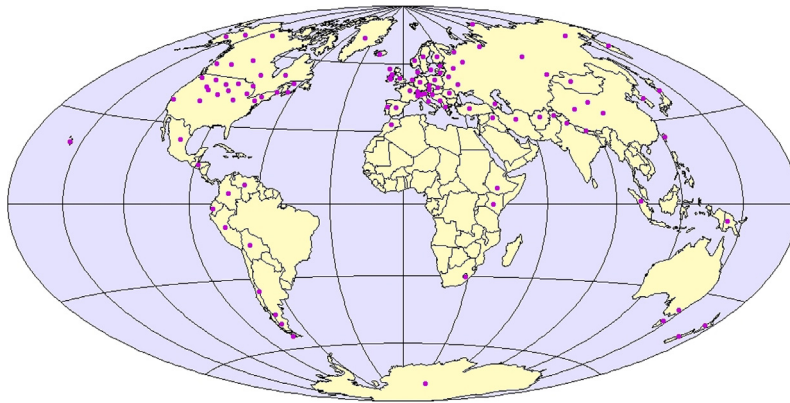


FIGURE 4.2

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 and Gibbard (2004a,b,c). MIS, Marine Isotope Stages.

the Pliocene and onwards to the present day in the mountains. Likewise, in Norway, its adjacent offshore and the neighbouring Barents Sea, glaciation is recorded from the Early Miocene, Early Pliocene, and Plio-Pleistocene.

A core from c. 75°N latitude in the Norwegian-Greenland Sea off East Greenland contains IRD with at least some likely from glaciers rather than sea ice, from c. 38 to 30 Ma (Late Eocene to Oligocene). Certain characteristics of this debris point to an East Greenland source and exclude Svalbard, the next-nearest land mass (Eldrett et al., 2007). This IRD may represent isolated mountain glaciers or more extensive ice-sheet cover (Alley et al., 2010). By the Late Miocene, inland ice sheets were periodically present in Greenland, especially in the mountainous east, with ice reaching the sea in SE Greenland, whilst contiguous ice sheets have occurred

**FIGURE 4.3**

Overview map showing the regions covered by the INQUA project '*Extent and Chronology of Glaciations*'. Each dot on the map represents one contribution to the project.

Adapted from Ehlers, J., Gibbard, P.L., 2003. Extent and chronology of glaciations. Quat. Sci. Rev. 22, 1561–1568.

since the earliest Pleistocene (c. 2.3 million years) (Funder, personal communication). In the eastern Rockies of the United States a much shorter glacial sequence occurs. Here Plio-Pleistocene-aged till is known from Montana, North Dakota, and California. On Mount Kilimanjaro, in East Africa the first glaciation is recognized at c. 2.0 million years (?c. MIS 68). In Europe glaciation before the Middle Pleistocene is represented only by ice-rafted material, outside the mountain regions (e.g., in the Netherlands, lowland Germany, European Russia, and Britain). Substantial glaciation of the Baltic region late in the Early Pleistocene is indicated by erratic materials in the Netherlands (1–1.2 Ma; ?MIS 34–36). Glaciation is also established in the Alps from the Plio-Pleistocene.

In the Southern Hemisphere glaciation is much longer established, as noted above. Here the ice already formed in the Late Eocene–Early Oligocene in east Antarctica and built-up in a step-like pattern through the Neogene. The present polar conditions were already established by the Early Pleistocene after 2.5 Ma. A similar history is known from the piedmont areas of Argentina and Chile, where substantial ice caps were established by 14 Ma. Widespread lowland glaciation between 2.05 and 1.86 Ma (?c. MIS 68–78), followed by the 'Great Patagonian Glaciation' that took place at 1.15–1.00 Ma (c. MIS 30–34) (Coronato and Rabassa, 2011). Further north, the earliest glaciation recorded in the Bolivian Andes dates from 3.27 Ma, with extensive events at 2.2 Ma (c. MIS 82). In Columbia the record also begins at 2.5 Ma. The earliest records in Australasia are found in Tasmania and New Zealand from the Plio-Pleistocene (2.6 Ma: MIS 98–104; Colhoun and Barrows, 2011). Only slightly younger is New Zealand's oldest known glacial event (the Porika Glaciation; Barrell, 2011).

Table 4.1 Occurrence of Glaciation in Europe Through the Cenozoic Based on Numbers of Observations Presented in Contributions to the INQUA Project *Extent and Chronology of Quaternary Glaciations* (Ehlers and Gibbard, 2004a)

[illegible]

Table 4.2 Occurrence of Glaciation in North America Through the Cenozoic Based on Numbers of Observations Presented in Contributions to the INQUA Project *Extent and Chronology of Quaternary Glaciations* (Ehlers and Gibbard, 2004b)

Stage	Weichselian			Saalian			Elsterian	Cromerian			Early Pleistocene				Plio-/Pleistocene		Pliocene		Miocene
Magnetostratigraphy	B M												M G						
MIS (approx.)	2	4	6	6	8	?12	?14	16	? 18 or 20	22+	34 - 38	60	64 - 72	?78 - 82	c.100 - 104				
N Canada and Alaska	~	~	~	~	~	~	~	~	~	~	~	~	~	~	~	~	?	~	
Alberta			?			?													
British Columbia	~	?	~		?						?				?	?			
California											?			?					
Great Basin							?	?											
Illinois						?		?											
Minnesota										?	?	?	?	?	?				
Montana & N Dakota		?				?					?								
New England			?																
New Brunswick		?		?															
New York			?																
Nova Scotia		?	?																
Ohio										~									
Ontario																			
Pennsylvania										?									
Québec				?															
Rocky Mountains		?		?										?					
Washington		?								?	?								
Wisconsin		?													~	~	~	~	
Greenland	~	~	~	~	~	~	~	~	~		~	~	~	~	~	~			

Table 4.3 Occurrence of Glaciation in the Rest of the World Through the Cenozoic Based on Numbers of Observations Presented in Contributions to the INQUA Project *Extent and Chronology of Quaternary Glaciations* (Ehlers and Gibbard, 2004c)

Stage	Weichselian				Saalian		Elsterian	Cromerian				Early Pleistocene				Plio-/Pleistocene		Pliocene		Miocene	Palaeogene						
Magnetostratigraphy	B M																										
MIS (approx.)	2	4	6	6	8	?12	?14	16	? 18 or 20	22+	34 - 38	60	64 - 72	?78 - 82	c.100 - 104												
Antarctica + S Ocean	~	~	~	~	~	~	~	~	~	~	~	~	~	~	~	~	~	~	~	~							
Argentina																											
Patagonia																											
Bolivia																											
Chile																											
China - Tibet								?																			
China - Tianshu								?							?												
Columbia										?																	
East African Mountains																											
Equador					?	?																					
Japan			?																								
Mexico		?	?			?																					
New Zealand																											
Tasmania								?									?										

4.5 THE 'GLACIAL' PLEISTOCENE

The 'glacial' Pleistocene effectively began with extensive glaciation of lowland areas, particularly around the North Atlantic region, and the intensification of global cold period (glacial) climates, in general. Till sheets of the major glaciations during the 'glacial Pleistocene' are found throughout Europe and Siberia. In northern Europe, till sheets characterize large areas of the lowlands and are also found on the floors of the adjacent seas (Gibbard and Clark, 2011; Karabanov and Matveyev, 2011; Johansson et al., 2011; Kalm et al., 2011; Lee et al., 2011). Recent investigations have demonstrated the extent to which the North Sea floor was sculpted by repeated glaciation (Graham et al., 2011; Sejrup et al., 2015). Further south, in central and southern Europe, glacial diamicton is

BOX 4.2 NATURE OF THE EVIDENCE OF GLACIATION AND LIMITATIONS

The evidence of ancient glaciations is generally preserved in two ways: on land, glacial diamicton (till, tillite) and related meltwater deposits occur in sedimentary sequences, whilst beneath the seas the main evidence occurs as waterlain diamictons or, more frequently and far-wider distributed, as IRD carried by floating ice. In volcanically active regions, such as the Antarctic Peninsula or Iceland, the study of glaciovolcanic sequences provides another powerful tool to reconstruct the thickness, extension, and thermal regime of past ice sheets (Smellie et al., 2008, 2009; Geirsdóttir, 2011).

By their nature, terrestrial glacial sedimentary sequences are fragmentary and are generally preserved where they have been protected by later accumulations or volcanic products. Precisely where these glaciations occurred and how far they extended is often very difficult to determine, given that the remnants of less extensive early glaciation tend to be obliterated and mostly removed by later, more extensive advances. Although this is so in all terrestrial areas, it is especially difficult in mountain regions where the preservation potential of older sequences rapidly diminishes with time and subsequent glaciation. This fragmentary pattern of preservation of terrestrial glacial sequences contrasts with that in the oceans where sequences, once deposited, remain largely unmodified for much longer periods. The only exceptions are those sequences affected by submarine debris flows or those that were laid down on the shelves or within fjords, etc. where later glaciation and sea-level changes can cause removal of preexisting evidence. Whilst all these points are true for glaciations at any time during the geological history, it is particularly a problem for the reconstruction of events during the late Neogene and Early Pleistocene. Then, glaciation was considerably less extensive than during the later Quaternary, implying that younger glaciations may have removed much of the evidence of previous glaciations. Nevertheless, by comparing the available Early Pleistocene glacial evidence on a global scale provides insights on the timing, location, and minimum extent of glaciation.

The transport of materials by floating ice is one of the principal processes by which coarse glacial sediment is dispersed in water bodies. The occurrence of clastic sediment in ocean-bottom sediments is generally interpreted as evidence that glaciers were sufficiently extensive to have reached the sea (i.e., tidewater glaciers). Although transport by calved icebergs is the most common way, it is also possible for materials to be transported in ocean basins by sea ice (Nürnberg et al., 1994; Andrews, 2000) or, although less commonly, by algae, driftwood, and the like (Gilbert, 1984; Vogt and Parish, 2012).

The deposits arising from both sea-ice and iceberg melting are termed ice-rafted detritus or debris (IRD). Icebergs calved from glaciers that terminate in the sea (or lakes), especially those from temperate ice tongues, frequently include debris that is frozen onto or within the ice body. As the ice drifts with the ocean currents and wind, its included sediment can be transported over long distances that can exceed 1000–1500 km (Ruddiman, 1977; Bond et al., 1992; Williams et al., 2010) within and across ocean basins. According to Dowdeswell and Murray (1990), the volume of sediment deposited by iceberg rafting is dependent on several factors that are determined by the primary sediment source, the rate of glacier calving, the rate of iceberg melt, the ocean surface-water temperature and factors controlling the iceberg drift rate (cf. Benn and Evans, 2010). Icebergs can be produced in different ways, which Williams et al. (2010, p. 358–359) summarize as follows: ‘IRD-bearing icebergs can be produced when an ice sheet reaches the edge of the continental shelf during glacial maxima (Marshall and Koutnik, 2006), during glacial advance (McManus et al., 1999), during glacial retreat (McManus et al., 1999), or due to instabilities in the ice sheet leading to ice-rafting like the Heinrich events of the North Atlantic in the last glacial (Hemming, 2004)’. Because we review data of glaciation over a time scale of almost 3 million years, different production mechanisms are not differentiated, which operate on shorter time scales. Here the IRD records are simply taken as evidence for an ice sheet that reaches the coastline.

In subpolar regions, the occurrence of coarse-grained ($>63\ \mu\text{m}$) sediment and clasts in open-ocean sediments is generally attributed to iceberg transport and melting, and provides indirect evidence of terrestrial glaciation (Andrews, 2000; St. John, 2008; Polyak et al., 2010). In polar regions, melting of sea ice usually contributes fine-grained sediment, but occasionally also coarse-grained sediment, to the geological record (Nürnberg et al., 1994; Müller and Knies, 2013). Fine-grained sediment that occurs in suspension on shallow shelves can be incorporated in sea ice during ice-freeze-up (frazil-ice formation) (Kempema et al., 1989; Nürnberg et al., 1994; Polyak et al., 2010). In addition, anchor ice, which forms on the sea-floor during super-cooled conditions, can incorporate and distribute any sediment (Reimnitz et al., 1987; Polyak et al., 2010). Wind-blown sand also contributes to the sediments carried by sea ice and icebergs (Atkins and Dunbar, 2009). The sediment transported by sea ice cannot be seen as direct evidence of glaciation on land, and

(Continued)

BOX 4.2 NATURE OF THE EVIDENCE OF GLACIATION AND LIMITATIONS (CONTINUED)

makes the distinction between sea ice- and iceberg-transport processes challenging yet essential for palaeoenvironmental interpretations.

Based on grain size alone, it is often not possible to attribute a small coarse-grained fraction in deep-sea sediments to either sea ice or iceberg transport in the geological record (St. John, 2008). Using surface textures of sand-sized quartz grains, St. John (2008) demonstrated the presence of sea ice and iceberg-derived sediment in the Arctic Ocean since the Eocene. Chemical fingerprinting of iron oxide sand grains illustrated periods with perennial sea ice possibly already in the Middle Eocene (c. 47.5 Ma; [Darby, 2014](#)) and in the Late Miocene ([Darby, 2008](#)).

restricted to the mountains and piedmonts ([Calvet et al., 2011](#); [Giraudi, 2011](#); [Nývlt et al., 2011](#); [Urdea et al., 2011](#); [Van Husen, 2011](#); [Woodward and Hughes, 2011](#)).

In northern Europe and Siberia, widespread lowland glaciation began in the early Middle Pleistocene shortly after the Brunhes–Matuyama palaeomagnetic reversal (0.78 Ma). The phases represented include the Donian (Narevian, Sanian, MIS 16, 'Cromerian Complex' Stage), Elsterian (Okan, MIS 12), Saalian (Dniepr/Moscovian and Vologda, MIS 6, 8, and 10), and the Weichselian Stage (Valdaian, MIS 4–2) (e.g., [Astakhov et al., 2015](#)). More limited glaciation may have occurred in the circum-Baltic region during the latest Early Pleistocene (MIS 22 and 20). The evidence for Early Pleistocene glaciations is restricted in the region to Latvia ([Zelčs et al., 2011](#)), Poland ([Marks, 2011](#)), and possibly Lithuania ([Guobytė and Satkūnas, 2011](#)), although current research suggests pre-Cromerian glaciation could have extended into central Jylland, Denmark ([Houmark-Nielsen, 2011](#)). Although Early Pleistocene glaciation has been demonstrated from ice-rafting in the northern North Sea ([Ottesen et al., 2014](#)) and a c.1.1 Ma till from offshore Norway ([Sejrup et al., 1995](#)), early Middle Pleistocene glaciation has not yet been recorded from the North Atlantic and offshore Norway (where the earliest Middle Pleistocene glaciation occurred ~0.6 Ma, possibly during MIS 12; [Sejrup et al., 1995](#)), it is certainly present in Denmark, the Baltic region, and European Russia. The first evidence of major lowland expansion of the Fennoscandian ice sheet during the Menapian Stage potentially correlates with the 1.1 Ma Fedje Glaciation of [Sejrup et al. \(1995\)](#).

In the Italian Dolomites, glaciation became established in MIS 22 ([Muttoni et al., 2003](#)). Comparable evidence is also found north of the Alps in Switzerland and southern Germany ([Fiebig et al., 2011](#)). The oldest glaciation identified in the Pyrenees is of late Cromerian age (MIS 16 or 14) ([Calvet, 2004](#)). In North America, widespread lowland glaciation is first seen during MIS 22 or 20 ([Barendregt and Duk-Rodkin, 2011](#); [Duk-Rodkin and Barendregt, 2011](#)). Subsequently, major ice sheets repeatedly extended over large regions of the continent during the Middle Pleistocene pre-Illinoian events MIS 16, 12, 8, and 6 (Illinoian s.s.) and the Late Pleistocene MIS 4–2 (Wisconsinan). In Mexico, the oldest moraines on volcanoes have been dated to 205–175 ka, corresponding to advances during the Illinoian Stage (MIS 6) ([Vásquez-Selem and Heine, 2011](#)). Evidence from East Greenland suggests that the southern dome of the ice sheet may have almost disappeared during the Eemian Stage interglacial (c. MIS 5e). However, dating of the basal ice in the Dye 3 ice core has shown that glacial ice has been present over this locality since at least MIS 11 ([Funder et al., 2011](#)).

According to Chinese investigations, glaciation of Tibet and Tianshu is not recorded before the Middle Pleistocene, of which the MIS 12 glaciation was the most extensive. Four discrete Pleistocene glaciations have been identified on the Qinghai-Tibetan Plateau and the bordering mountains. These major Pleistocene glaciations are equated to MIS 18–16, 12, 6, and 4–2. This apparently delayed glaciation of the Himalayan chain might reflect late uplift of high Asia. The Kunlun Glaciation (MIS 18–16) was the most extensive. Subsequent glaciations have been successively more reduced, probably caused by increasingly arid climates resulting from the progressive Quaternary uplift of the plateau (Zhou Shangzhe et al., 2011). By contrast, Kuhle (2011) maintains that the last glaciation was most extensive since he envisages an ice sheet covered Tibet. However, this is an increasingly minority view, particularly the concept that the Qinghai-Tibetan Plateau was almost entirely covered by an ice sheet. The majority opinion is that Pleistocene glaciations over the Qinghai-Tibetan Plateau were much more restricted (Klinge and Lehmkuhl, 2004; Lehmkuhl and Owen, 2005). As these authors argue, it is now generally well established that a large ice sheet did not cover the Tibetan Plateau. This controversy serves to illustrate the fact that in several regions debate over the glacial history continues.

As in the Northern Hemisphere circum-Atlantic region, glaciation increased in intensity after 800 ka throughout the South American Andes. In the northern Andes there is evidence of glacial deposition from the Early Pleistocene in Venezuela and Colombia (e.g., Helmens, 2011). In the southern Andes, the Late Pleistocene saw glaciation that was less extensive than during the Early Pleistocene events (Coronato and Rabassa, 2011; Martínez et al., 2011; Harrison and Glasser, 2011). Both easternmost Tierra del Fuego and the Falkland Islands are thought by some to have remained largely unglaciated throughout the Pleistocene (Clapperton, 1993), though there are traces of glacial scouring (Ehlers and Gibbard, 2008). In Australasia, following a 1 Ma break, the glacial record resumes in MIS 12, followed by MIS 6, 4, and 3. In Tasmania, the earliest Pleistocene ice advances are thought to have been about 1 million years old but may be older. Middle Pleistocene ice advances occurred during MIS 10, 8, and 6 (Colhoun and Barrows, 2011). The presence of glaciations in this area during MIS 8, indicated by cosmogenic exposure dating, is interesting because a record of glaciation during this interval is absent from many other parts of the world. In Africa, glaciations were restricted to the East African mountains including Mount Kilimanjaro, Mount Kenya, and the High Atlas, where the major glacial events appear to have been broadly synchronous with those elsewhere beginning in MIS 12. In Asia, the mountains of Japan were glaciated (Sawagaki and Aoki, 2011), as well as the high mountains of Taiwan and Indonesia (Böse and Hebenstreit, 2011; Prentice et al., 2011).

The repeated records for the occurrence of Middle Pleistocene glaciations during MIS 16 and MIS 12 throughout the world are striking, whereas the absence of evidence for glaciations during MIS 18, and to a lesser extent MIS 14, implies that later glaciations were more extensive and therefore probably destroyed evidence for these events. However, it is also possible that some of these glacial deposits have been inadequately differentiated owing to the limitations of current dating techniques. Lack of precision for dating terrestrial Middle Pleistocene sequences leaves some uncertainty until better dating techniques become available. It is likely that the application of multiple methods of developing glacial geochronologies will be required in order to gain greater precision, and, crucially, to test the validity of the various dating techniques (cf. discussion in Heine, 2011a,b).

It therefore appears from direct evidence that until c. 1.2 Ma the cold periods (glacials) were not 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 (c. 5–6 intervals) began in the last 900 kiloyears (c. MIS 22 to present) in the Northern Hemisphere (Head and Gibbard, 2015a). In contrast, southern hemispheric glaciation is a much longer established phenomenon with substantial glaciation already a regular occurrence in the Tertiary.

The timing of the increased intensity of glaciation corresponds closely to observations around the world concerning the increase in relief generation. For example, it has been estimated that in the Swiss Alps, relief was increased 1250 ± 650 m in the last c.1 Ma (Häuselmann et al., 2007). Recent observations in the North Alpine Foreland indicate that thick glaciofluvial units, dating predominantly from the Penultimate Glacial Maximum (PGM, i.e., MIS 6a; cf. Box 4.3), overlie intensely eroded and overdeepened Miocene bedrock. This intense erosion probably occurred during Early Pleistocene major glaciations, based on luminescence ages (Salcher et al., 2015). Similar scale erosion occurred in southern Norway, where incision of ancient 'palaeic' surfaces to a depth of as much as 2 km occurred (Reusch, 1901; Lidmar-Bergström et al., 2000). This incision has

BOX 4.3 ASYNCHRONEITY OF GLOBAL GLACIATION

The term Last Glacial Maximum (LGM) refers to the period when global ice volume last reached its maximum and eustatic sea levels were at their lowest, between 27.540 and 23.340 ka (Hughes and Gibbard, 2015). However, the boundaries of the interval are ill-defined and therefore the term and acronym have no formal stratigraphical basis. As Hughes et al. (2013) have shown, increasingly evidence from terrestrial records demonstrates that the LGM is poorly represented since many sequences show a diachronous response to global climate changes during the last glacial cycle. For example, glaciers and ice sheets reached their maximum extents at widely differing times in different places. In fact, most terrestrial records display spatial variation in response to global climate fluctuations, and changes recorded on land are often diachronous, asynchronous, or both, leading to difficulties in global correlation (Hughes and Gibbard, 2015). For this reason it is recommended that the term LGM be considered an event-based—rather than a time-based—unit.

The LGM exemplifies the value of adopting an event-based classification for glacial sequences (i.e., a scheme in which boundaries can be diachronous in nature). However, in spite of this utility, glacial maximum (GM) events have not been formally recognized for earlier glaciations. Nevertheless, Margari et al. (2014) mentioned that for Marine Isotope Stage (MIS) 6 'eustatic sea level records and glacial geological evidence suggest that ice sheets expanded, with global ice-volume reaching its maximum extent toward the end of MIS 6', i.e., ~140–136 ka (within MIS 6a; 'PGM' or 'Penultimate Glacial Maximum', sensu Stirling et al., 1998); or c. 149–140 ka (Railsback et al., 2014, 2015b).

The penultimate glaciation (c. 135–185 ka) corresponds to MIS 6, the Illinoian glaciation in North America, the Saalian glaciation in Europe, and equivalents elsewhere. In Europe it was characterized by two major ice advances: the more extensive Drenthe, followed by the Warthe. No major interstadial warming occurs within the Saalian, and this is in concert with the lack of organic deposits and absence of palaeosols in northern Europe during the interval between the Drenthe and the Warthe Stadials (Ehlers et al., 2011b). After c. 150 ka, glacial geological evidence suggests that ice sheets expanded, with global ice volume reaching its maximum extent towards the end of MIS 6, reflecting the growth of the Late Illinoian ice sheet in North America (e.g., Curry et al., 2011; Syverson and Colgan, 2011). In Europe, the Warthe ice advances were less extensive than the Drenthe (Ehlers, 2011), but that may have been compensated for by ice expansion in Russia and Siberia (e.g., Astakhov, 2004). Compared to the Last Glacial Maximum (LGM), late MIS 6 (i.e., MIS 6a) was characterized by overall larger ice sheets (Margari et al. 2014), although modelling by Colleoni et al. (2016) suggests that the Laurentide Ice Sheet might have been smaller than during the PGM at 140 ka.

Previous glaciations also show GM peaks of similar duration, such as within MIS 12a (~437–430 ka) and MIS 16a (~640–625 ka), where marine isotope substage divisions follow Railsback et al. (2015a,b). As Hughes et al. (2013) noted, one of the major implications of the asynchronicity of maximum glacial extent apparent in the last glacial cycle is the possibility that this pattern of asynchronicity of GM extent might have been repeated during earlier glaciations, particularly during the Middle Pleistocene (Ehlers et al., 2011a,b). If this occurred it could explain why some glaciations are represented in one region of the world, whilst apparently being absent elsewhere.

been attributed to the opening of the Atlantic Ocean and accompanying crustal unloading. However, [Medvedev and Hartz \(2015\)](#) have concluded that 0.8 km uplift of rocks and up to a 0.5 km rise of local topography resulted from fluvial and glacial erosion as a consequence of climatic deterioration during the Quaternary. [Champagnac et al. \(2014\)](#) also consider that cold climates, accompanied by glacial and periglacial erosion, are more likely to have caused increase and/or decrease in relief. They attribute this to the related physical erosional processes being highly nonuniform in space and time and, equally, the sensitivity of these processes to minor climatic changes. They conclude that the great intensity of erosion under glacial and periglacial conditions suggests the overall observation that cold and variable Quaternary climate has resulted in an increase in relief at the global scale.

The application of numerical dating is increasingly demonstrating the potential asynchronicity of glacial extent on a global scale ([Box 4.3](#)). Whilst the differing timing of ice lobes at a local scale has been known for decades, evidence that the maximum extents of major ice sheet margins may differ by as much as tens of thousands of years is beginning to be established. Contrasts between different hemispheres, and within them, different environmental settings, appear to occur. Judging from the evidence from the Last Glaciation (see [Section 4.6](#)), such differences appear to result from the differing regional geographical situation, variations in the supply of precipitation, regional temperature variation, differing ocean and atmospheric circulation, etc. (cf. [Margari et al., 2014](#)).

4.6 LAST GLACIATION

The term Last Glacial Maximum or LGM is widely accepted as referring to the maximum global ice volume during the last glacial cycle corresponding with the trough in the marine isotope record centred on c. 18 ^{14}C ka BP ([Martinson et al., 1987](#)) and the associated global eustatic sea-level low also dated to 18 ^{14}C ka BP ([Yokoyama et al., 2000](#)). It has also been assigned chronozone status (23–19 or 24–18 ka cal BP dependent on the dating applied) by [Mix et al. \(2001\)](#) who consider the event should be centred on the calibrated date at 21 ka cal BP (i.e., LGM *sensu stricto*). However, other research on global sea-level minima places the global ice maximum slightly earlier at between 26 and 21 ka ([Peltier and Fairbanks, 2006](#)). A recent paper by [Hughes and Gibbard \(2015\)](#) combined the evidence from marine cores with that from ice cores on land to define the LGM. Unlike marine sediment records which accumulate slowly and can be affected by bioturbation, ice-core records have a much higher resolution and provide evidence of climate changes on a year-by-year basis. [Hughes and Gibbard \(2015\)](#) identified the period between 27,540 and 23,340 years ago corresponding with Greenland Stadial 3 in the Greenland ice-core record, as the best candidate interval for its formal stratigraphical definition ([Fig. 4.4](#)). This interval was characterized by the largest global ice volume, the lowest global sea levels and the greatest atmospheric dust flux in both hemispheres and some of the coldest temperatures of the last ice age. The end of the LGM event is marked by evidence for a large discharge of icebergs into the North Atlantic Ocean (Heinrich Event 2) which coincided with the onset of the collapse of the Laurentide Ice Sheet at c. 24 ka, together with other ice sheets in the North Atlantic region.

During the LGM, the extent of the glaciation of the Southern Hemisphere differed very little from that of the Pleistocene glacial maximum (GM). Glaciers in Antarctica still reached the shelf

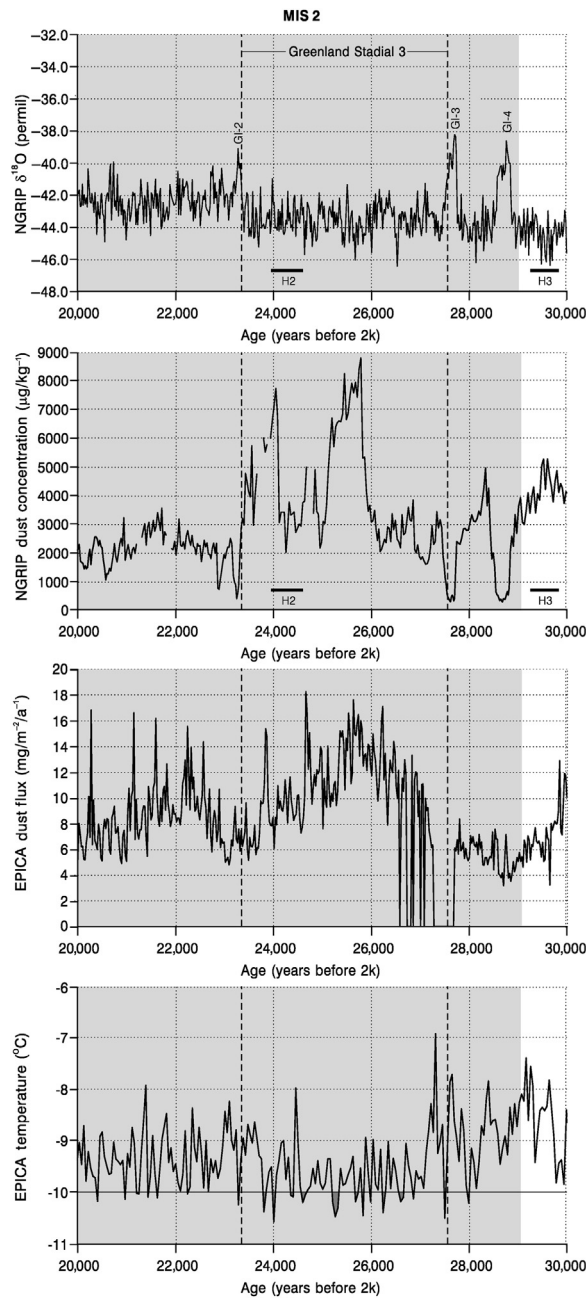


FIGURE 4.4

The proposed definition of the global Last Glacial Maximum (LGM) within Greenland Stadial 3 by [Hughes and Gibbard \(2015\)](#). The top two diagrams are the oxygen isotope ([Andersen et al., 2006](#)) and dust concentration ([Ruth et al., 2007](#)) records from the NGRIP core. The NGRIP core is on the GICC05 age model. The bottom two diagrams are the dust flux ([Lambert et al., 2012](#)) and deuterium-derived temperature ([Jouzel et al., 2007](#)) records from EPICA, Antarctica. The EPICA records are both on the EDC3 age model. H2 and H3 are Heinrich Events when large icebergs were discharged into the North Atlantic Ocean.

edge, and on New Zealand, Tasmania, and in South America the glacier tongues were only slightly smaller than during earlier events (Andersen et al., 1999; Coronato et al., 2004a,b). On mainland Australia, local mountain glaciation occurred (Barrows et al., 2001, 2002). It seems that the LGM in the Southern Hemisphere began earlier than in the Northern Hemisphere, probably at around 27 ka (Suggate and Almond, 2005). The high mountains of East Africa were glaciated (Osmaston, 2004). There is only evidence of minor glaciers in South Africa (Hall, 1994; Mills et al., 2009). It must be borne in mind that South Africa is located relatively close to the equator to the North. Were it in the Northern Hemisphere, Cape Town would be situated at the same latitude as Atlanta in Georgia, USA, or, if placed relative to the European ice sheet, it would be south of Tunis.

During the LGM, ice in many parts of the Northern Hemisphere reached an extent very similar to the Quaternary GM. In North America, the differences are very small. Again, most parts of Canada were ice-covered, including the shelf areas (Dyke, 2004). It is the same in Greenland (Funder et al., 2004; Weidick et al., 2004).

In Europe, however, the situation was different. New evidence suggests that the North Sea was not fully glaciated during the Weichselian GM, but slightly earlier (Sejrup et al., 1994, 2009; Carr, 2004). An ice sheet covered the Barents Sea and extended well into the Kara Sea but hardly touched the Russian mainland (Svendsen et al., 1999, 2004a,b). The ice sheet was markedly smaller than during the Quaternary GM. It seems that the northward drainage of the Ob and Yenisei rivers was not impeded (Lambeck et al., 2006). In glaciated mountain areas outside the major ice sheets, such as in Italy and Greece, the maximum Middle Pleistocene glaciations were much greater than the local last glacial maxima (Giraudi, 2004, 2011; Woodward et al., 2004; Woodward and Hughes, 2011). This is attributed to a change in equilibrium-line altitude. Features on the floor of the Arctic Ocean seem to suggest that the Arctic basin was once covered by a one-kilometre-thick, flowing ice shelf derived from large ice sheets in eastern Siberia, Arctic Canada, and the Barents Sea (Polyak et al., 2001; Domack, 2016).

There were major ice sheets or mountain glacier systems in the Siberian mountains further to the east. Because of the lack of detailed investigations to date, in most cases it is not possible to differentiate between the extent of the LGM and earlier glaciations (Brigham-Grette et al., 2003; Glushkova, 2011). Likewise, the age and extent of the glaciers in Iran and the mountain ranges and high plateaux further to the east, and especially the extent of glaciation on the Tibetan Plateau, are still matters of debate (Owen et al., 2003; Klinge and Lehmkuhl, 2004; Kuhle, 2004; Lehmkuhl and Owen, 2005; Abramowski et al., 2006). In contrast, the small mountain glaciers of Japan are well mapped and dated and, unlike those of the adjoining Asian mainland, the maximum ice advance of the last glaciation occurred in MIS 4 (Sawagaki et al., 2004). The same date for the maximum extent is found in New Zealand (Suggate, 1990; Preusser et al., 2005).

Whilst the LGM was a major ice advance around the world, it is now clear that glaciers reached their true *maximum* Late Pleistocene extent at different times during the last glacial cycle (Hughes et al., 2013). Whilst this finding is not new for mid-latitude mountain glaciers (cf. Gillespie and Molnar, 1995), the degree of complexity and variability both spatially and temporally revealed is striking. In the global review by Hughes et al. (2013), it is clear that in many areas glaciers reached their maximum extent early in the last glacial cycle in MIS 3, 4, or even 5, well before the global LGM in MIS 2. Furthermore, this is not only a phenomenon that is restricted to mid-latitude mountain glaciers but it is also seen in certain sectors of the large continental ice sheets.

The Laurentide Ice Sheet maximum corresponds to the global LGM in MIS 2 and dominates the marine oxygen isotope and sea-level records. In contrast, the Barents-Kara ice sheet in northern

Eurasia reached its maximum extent during MIS 4, as did the western sectors of the Fennoscandinavian Ice Sheet and the northern sector of the Cordilleran Ice Sheet. In Asia, the vast majority of mountain glaciers reached their maximum positions before MIS 2, in MIS 5, 4, and 3. The same appears to be valid for New Zealand and southeast Australia and parts of South America. The East Antarctic Ice Sheet also appears to have been out-of-synchronization with global ice volume, retreating from its maximum extent before MIS 2. Thus the concept of the global LGM needs to be carefully considered in terms of meaning and significance (Hughes et al., 2013; Hughes and Gibbard, 2015). The findings from the last glacial cycle are also likely to be true for all other glacial cycles and further research is required to decipher the relationship between global glacier activity and other records of global change such as the marine isotope record.

4.7 SUMMARY

The Quaternary is synonymous with extensive glaciation of Earth's mid- and high-latitudes. Although there were local precursors, significant glaciation began in the Oligocene in eastern Antarctica. It was followed by glaciation in mountain areas through the Miocene (in Alaska, Greenland, Iceland, and Patagonia), later in the Pliocene (e.g., in the Alps, the Bolivian Andes, and possibly in Tasmania) and in the earliest Pleistocene (New Zealand, Iceland, and Greenland). Today evidence from both the land and the ocean floors demonstrates that the major continental glaciations, outside the polar regions, rather than occurring throughout the 2.6 Ma of the Quaternary, were markedly restricted to the last 1 Ma—800 ka or less. MIS 22 (c. 870–880 Ma) included the first of the 'major' worldwide events with substantial ice volumes that typify the later Pleistocene glaciations (i.e., MIS 16, 12, 10, 6, 4–2).

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