A stratigraphical basis for the Last Glacial Maximum (LGM)

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1. Introduction

The term ‘Last Glacial Maximum’ has attained iconic status in Quaternary Science. The often capitalised letters form the acronym LGM and suggest a formal stratigraphical status. However, the search for a common definition for the LGM is not as obvious as the acronym implies. Hughes et al. (2013) argued that the geomorphological record of glaciations does not provide a clear definition for the LGM since the imprint of glaciers in the geological and geomorphological records is asynchronous around the world. This is not only for mid-latitude mountain glaciers (e.g. Gillespie and Molnar, 1995) but also for some of the large continental ice sheets. The current status of the LGM is defined not by the terrestrial record of global environmental change, but by changes recorded in the ocean floor sediments’ oxygen isotope sequence.

In recognition of this paradox, Mix et al. (2001) used dated corals and marine oxygen isotope sequences to suggest the LGM as a chronozone. However, as noted by Walker et al. (2009) when defining the base of the Holocene Series, marine records do not offer the resolution to provide a formal stratigraphical basis for defining short-duration time divisions, exemplified by the LGM. Mix et al. (2001) argued that the LGM should be centred on the radiocarbon calibrated date of 21 cal. ka BP and should span the period 23–19 ka or 24–18 ka (all further ages in this paper are calibrated or in calendar kilo years, expressed as ka). However, the boundaries for these chronozones are not defined in a specific “body of rocks” (cf. Salvador, 1994) but are arbitrary time intervals...
spanning events within in a number of different records. Thus the chronozone status advocated by Mix et al. (2001) and others such as the MARGO Project (2009) can only be considered an informal labelling of the LGM interval. In addition, there is now evidence that the timings of maximum global ice volume and the lowest eustatic fall are at the older end of the interval 24–18 ka (cf. Thompson and Goldstein, 2006) with Peltier and Fairbanks (2006) suggesting that the LGM occurred as early as 26 ka.

On land, the evidence for an LGM climate signal is even more transient, leading researchers often to lean heavily on the marine records for correlation. However, a truly global LGM time division requires definition in both land-based and marine-based environmental proxies in order to provide a stratigraphical unit that can be used effectively and meaningfully in correlation. This paper provides a critique of the various Quaternary records, both marine and terrestrial. In particular, this paper seeks to establish the most appropriate record(s) and a suitable stratigraphical definition for this iconic label in the Quaternary geological succession. In this respect, it builds on suggestions proposed at the First International Conference on Stratigraphy (Hughes and Gibbard, 2014).

2. Current definitions of the LGM

The term ‘Last Glacial Maximum’ (abbreviated to LGM) refers to the maximum in global ice volume during the last glacial cycle. The LGM was originally described by CLIMAP Project Members (1976, 1981) as spanning the interval 23,000–14,000 14C ka BP, with a mid-point at 18,000 14C ka BP (Shackleton et al., 1977). It is marked by two independent proxies: in the marine isotope record and changes in global sea level; and, it is on this basis that the LGM was defined (Mix et al., 2001).

The δ18O signal in the marine record is known to lag global ice volume (Mix et al., 2001; Thompson and Goldstein, 2006) and, consequently, the global sea-level minimum is likely to be closer to the true global Last Glacial Maximum in terms of maximum ice volume. Based on evidence of global sea-level change from the continental margin of northern Australia, Yokoyama et al. (2000) concluded that the global land-based ice volume was at its maximum from at least 22–19 cal. ka BP. As noted earlier, the age of 21 ka is now widely used as a time marker for the acme of the global LGM (Mix et al., 2001; MARGO Project Members, 2009).

The definition of the LGM in terrestrial records depends on the criteria applied. Shakun and Carlson (2010) used 56 records to recognise a climate-defined LGM. They suggested that a global average age of 22.2 ± 4.0 ka best defines the LGM. However, they recognised that ‘there is considerable variation in the timing of these extreme climate states in different records with the LGM … spread over more than 10 kyr’ (Shakun and Carlson, 2010, p. 1802).

Shakun and Carlson (2010) found that the LGM within 56% of their records fell within the chronozone span of 23–19 ka, defined by Mix et al. (2001), and noted that this chronozone does not appear to capture the length or variability of the LGM. In their dataset the largest frequency of climate-defined ‘LGM’ events in the northern hemisphere are at 24 ka and 30 ka (Shakun and Carlson, 2010, their Fig. 4), although younger ‘LGM’ events between 23 and 16 ka result in a global average close to 22 ka.

Whilst Mix et al. (2001) proposed that the LGM should be defined as a chronozone, the boundaries of such a unit, as determined from a particular type-section, remain elusive. Sea-level and ice-core evidence, which provided the basis for the LGM chronozone in Mix et al. (2001, their section 5.4), provides broad indications of the glacial maximum event. However, the bracketing ages do not conform to the strict formal requirements of a chronozone (cf. Hedberg, 1976; Salvador, 1994). Furthermore, Mix et al. (2001) noted at that time that there were some inconsistencies between the different ice-core chronologies and that ‘some puzzles remain to be solved regarding ice-core chronologies near the LGM’.

Since then, new ice-core records have been obtained and used to define events at the end of the last glacial cycle (e.g. Andersen et al., 2006; Rasmussen et al., 2006; Lowe et al., 2008; Walker et al., 2009) (Fig. 1).

3. Defining the LGM in marine records

3.1. Oxygen isotopes

Marine oxygen isotopes, determined from the tests of foraminifer from deep-sea floor sediments, have been viewed as a proxy for global ice volume since the 1960s (e.g. Shackleton, 1967). Deep-water temperatures also play a role and this means that δ18O variability in benthic foraminifera is not entirely driven by ice volume (Shackleton, 2000). Nevertheless, this effect can be accounted for (Shackleton, 2000) and the basic tenet relating variations δ18O to shifts in global ice volume stills holds. Since benthic δ18O variability is driven by global ice volume, it also provides a globally synchronous record of glacial/eustasy (Skinner and Shackleton, 2005). Whilst this is true when considering changes on longer timescales (>5 ka) and especially 100 ka glacial cycles (e.g. Shackleton, 2000; Waelbroeck et al., 2002), the marine isotope record does not offer sufficient resolution to differentiate environmental events at millennial timescales. This is, in part, a consequence of the slow sedimentation rates in the deep oceans and especially bioturbation, which “is a virtually universal source of degradation for deep-sea records” (Shackleton, 1987, p. 183; McCave et al., 1995). However, in addition to this, there are other significant reasons why marine isotope records such as benthic δ18O cannot provide a globally correlatable stratigraphical scheme for fine-resolution intervals such as the LGM (Gibbard, in press).

The oxygen isotope signal in the marine record is often assumed to be a proxy for global ice volume. However, it is not a straightforward as this and, as noted above, Shackleton (2000) highlighted that a substantial part of the 100 ka glacial climate cycle, recorded by δ18O in marine foraminiferal records, is a deep-water temperature rather than an ice-volume signal. Skinner and Shackleton (2005) showed that fluctuations in benthic δ18O and MIS boundaries from different hydrological settings may be significantly diachronous. The use of benthic δ18O as a proxy for global ice volume as established by Shackleton (1967) begins to “break down at millennial time-scales and in particular across glacial–interglacial transitions” (Skinner and Shackleton, 2005, p. 578). Thus, for relatively short intervals such as the LGM, the marine isotope record is inappropriate for defining its span.

The timescale of the deep ocean δ18O signal has been constructed by assuming that solar forcing paces variations in δ18O variations (Hays et al., 1976) and this provides the basis for the SPECMAP timescale. Thompson and Goldstein (2006) calibrated this timescale using radiometric dating and found significant discrepancies in the orbital tuning with observed U-series ages from corals. They found that for Marine Isotope Stage (MIS) 2, SPECMAP ages were too young, up to 5.2 ka too young at 17.9 ka, the original date assigned to the trough in δ18O for the last glacial cycle by Martinson et al. (1987, event 2.2, in their Fig. 18). The newly adjusted radiometric calibration of SPECMAP places the trough in δ18O at 23.1 ka, the lowest sea levels bracketed between 23.1 and 24.3 ka, with the latter age corresponding to the greatest sea-level lowering of ~132.1 m (Thompson and Goldstein, 2006). An offset also exists between the high-resolution coral-derived sea-level curve of Thompson and Goldstein (2006) and a global synthetic ‘stack’ of 57 marine oxygen isotope records compiled by...
Lisiecki and Raymo (2005) — (Fig. 2). In their ‘stack’, Lisiecki and Raymo (2005) correlated the interval before 22 ka with the radiocarbon-dated benthic $\delta^{18}O$ record of Waelbroeck et al. (2001) and the interval from 22 to 120 ka with the high-resolution benthic $\delta^{18}O$ record of Shackleton et al. (2000). In Lisiecki and Raymo’s (2005) LR04 ‘stack’ the $\delta^{18}O$ trough occurs c. 5 ka earlier (at 18 ka, as with SPECMAP) than the sea-level minimum of Thompson and Goldstein (2006). The lag of $\delta^{18}O$ behind sea-level minima was noted by Mix et al. (2001, p. 637) who concluded that “the highest value of benthic foraminiferal $\delta^{18}O$ is not precisely aligned with the lowest sea-level stand, and may even be offset in time by as much as a few thousand years”.

Fig. 1. The ice-core records from Greenland (NGRIP) and Antarctica (EPICA). 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.
4. Defining the LGM in terrestrial records

4.1. Glaciers

The term Last Glacial Maximum is often used to refer to the peak in global ice volume during the last glacial cycle. This is reflected in the marine oxygen isotope record and also global sea levels recorded in corals (Mix et al., 2001). On land, Clark et al. (2009) reviewed 5704 $^{14}$C, $^{10}$Be, and $^{3}$He ages that span the interval from 50 to 10 ka to constrain the timing of the LGM during MIS 2. They found that glaciers advanced between 33 and 26.5 ka and reached maximum positions between 26.5 and 20/19 ka, with rapid deglaciation occurring soon after. However, for the longer interval of the entire last glacial cycle the pattern of glacier advances, and in particular, the maximum extent of ice masses, was not synchronous. Hughes et al. (2013) noted that at high, mid- and low latitudes across the world, glaciers reached their maximum extent before MIS 2, in MIS 5, 4 and 3. It is well-established that mid-latitude mountain glaciers were asynchronous with the global record of ice volume (Gillespie and Molnar, 1995) but increasingly, new dating evidence has revealed that even some of the largest ice sheets were out-of-phase with the record of global ice volume determined from the marine isotope record (Hughes et al., 2013).

The East Antarctic Ice Sheet, which today represents the largest modern ice mass on Earth, retreated from its maximum position well before the global LGM during MIS 3 (Stolldorf et al., 2012) and in parts of the East Antarctic the ice-sheet thickness at the LGM was little different from that today (Mackintosh et al., 2007). In New Zealand, glaciers also retreated during MIS 3, with less extensive advances occurring in MIS 2, i.e. close to the global LGM (Putnam et al., 2013). In the Northern Hemisphere the Barents-Kara Ice Sheet, the third largest ice mass of this hemisphere, after the Laurentide and Fennoscandinavian, reached its maximum position early in the last glacial cycle, as early as 90 ka. In Asia, most glaciers reached their maximum before MIS 2, with the global LGM being recorded by a less extensive advance (e.g. Dortch et al., 2013; Owen and Dortch, 2014), or not recorded at all (Heyman et al., 2011; Stauch and Lehmkuhl, 2011).

The Laurentide Ice Sheet was by far the largest ice mass on Earth during the last glacial cycle. Evidence from its eastern and southern margins supports a maximum phase during MIS 2. At Martha’s Vineyard, in Massachusetts, boulders on a moraine marking the outer limit of the SE sector of the Laurentide Ice Sheet yielded cosmogenic nuclide exposure ages that cluster between 22 and 25 ka with a mean age of $23.2 \pm 0.5$ ka (Balco et al., 2002). New production rates make these ages older and Balco and Schäfer (2006) suggest that the ice front started to retreat from the Mar tha’s Vineyard moraine at c. 24 ka. These authors pointed out that this new age coincides with Heinrich Event 2, a major period of ice rafting in the North Atlantic. All later ice-front positions in New England and Connecticut indicate less extensive glaciation (Balco et al., 2002; Balco and Schäfer, 2006). There is also evidence from other parts of the Laurentide Ice Sheet margin which suggest an ice maximum before 23 ka, such as in Ohio (Lowell et al., 1999; Szabo et al., 2011), Illinois (Curry et al., 2011) and North Dakota (Manz, 2005). In NW Canada, Zazula et al. (2004) used radiocarbon to date pro-glacial lake sediments and suggest that the maximum extent of the Laurentide Ice Sheet in this area occurred between 35 and 22 ka followed by a less extensive readvance at 22--16 ka. These ages, from the outer margins of the last Laurentide Ice Sheet, correspond to global sea-level evidence compiled by Peltier and Fairbanks (2006) and Thompson and Goldstein (2006) who place...
the low-stands at 26 and 24.6 ka, respectively. Indeed, in their modelling of ice volume, Stokes et al. (2012) take 25 ka as the time of the Laurentide Ice-Sheet maximum (Fig. 4).

The southeastern sector of the Fennoscandinavian Ice Sheet reached its maximum extent in MIS 2. Here, moraines have yielded a mean exposure age of 19.0 ± 1.6 ka (Rinterknecht et al., 2006; using a production rate of 5.1 ± 0.3 atom/g/yr). These ages are likely to be slightly older (by up to 15%; Owen and Dortch, 2014) based on new production rates. These are as low as 3.77 atoms/g/yr for terrestrial cosmogenic nuclides that been determined for northern Norway (e.g. Fenton et al., 2011). This is similar to the rest of the world where production rates are now thought to be in the range of 3.7–4.5 atoms/g/yr (Balco et al., 2009; Putnam et al., 2010; Briner et al., 2012; Young et al., 2013). This means that the SE sector of the Fennoscandinavian Ice Sheet began retreating from its maximal position slightly earlier than 19 ka, possibly at c. 21 ka, though still slightly later than the Laurentide Ice Sheet. The situation in the SW sector of the Fennoscandinavian Ice Sheet was different, and there is evidence that the southwestern sector reached its maximum extent earlier, during MIS 3 (Houmark-Nielsen, 2011). In the British Isles, most datasets indicate a maximum glacial extent close to the LGM in the period 30–20 ka (Clark et al., 2012). The British–Irish Ice Sheet is now known to have been much larger during the last glacial cycle than previously thought, reaching the Atlantic shelf edge (Ballantyne, 2010). The timing of ice-sheet retreat is currently the focus of intensive research. The ice sheet expanded south into the Celtic Sea area between 34.0 and 25.3 ka, reaching maximum limits at 25.3–24.5 ka, based on several geochronological techniques (radiocarbon, cosmogenic and optically stimulated luminescence) (Chiverrell et al., 2013). In the west, the British–Irish Ice Sheet reached its maximum extent contemporaneous with Heinrich Event 2, when it reached the continental shelf and was followed by rapid retreat during Greenland Interstadial 2 (Scourse et al., 2009). This is very similar to the SE sector of the Laurentide Ice Sheet (see above). Likewise, Rolfe et al. (2012) found that cosmogenic exposure ages in the Bristol Channel area (Lundy) indicate an earlier exposure history (>35 ka), suggesting that the ice margins may have been very dynamic. However, in Wales, Glasser et al. (2012) obtained exposure ages showing that the summits of some of the highest mountain peaks were revealed by c. 20 ka. The Welsh Ice Cap was thinning at this time, following an ice maximum sometime before, and is consistent with the chronology for ice-sheet retreat in the neighbouring Irish Sea calculated by Chiverrell et al. (2013).

In the western Himalaya, glaciers reached their maximum extent well before the global LGM, and in parts of Tibet there appears to be no evidence of an LGM-equivalent advance at all (Heyman et al., 2011). Nevertheless, glaciers did advance in many areas at this time, albeit with more restricted extent. Dortch et al. (2013) and Murari et al. (2014) analysed >1700 cosmogenic ages and identified numerous moraine surfaces dating to 20 ± 2 ka and 22 ± 2 ka, respectively. However, Owen and Dortch (2014) noted that exposure ages could be 15–20% older, although geomagnetic corrections could nullify this effect. Until the production rates and geomagnetic corrections have been better constrained, cosmogenic exposure ages may not be suitable for precisely-dating fine-resolution events.

4.2. Vegetation

In vegetation records from long lake sequences, the LGM is often interpreted from the minimum frequency of arboreal pollen. This reflects both an aridity and cold temperature signal. In southern Europe, the Lake Ioannina sequence shows an arboreal pollen minimum frequency at 24.969 ± 331 cal. ka BP [based on a mid-point of 24.638–25.299 cal. ka BP recalibrated from original radiocarbon age of 20.760 ± 230 14C ka (Tzedakis et al., 2002) using the IntCal13 calibration (Reimer et al., 2013)]. However vegetation records are not uniform across regions in recording the most extreme cold and arid part of the last glacial cycle. For example, in Iberia the coldest

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and driest part of the Late Pleistocene occurred during the interval 18.5—14.6 ka (Moreno et al., 2012). Elsewhere in the Mediterranean, whilst vegetation records appear to match climatic events recorded in the Greenland Ice Sheet there is no obvious unified LGM event with vegetation shifts sometimes more pronounced during Heinrich stadials (cf. Fletcher et al., 2010).

The LGM is also poorly defined in other parts of the world in vegetation records. For example, in Central America, Hodell et al. (2008) found that the period around the LGM between 23 and 19 ka was characterised by temperate pine—oak forest. As in Iberia, the driest climate phase in lowland Central America occurred later between 18 and 14.7 ka and both Moreno et al. (2012) in Iberia and Hodell et al. (2008) in Central America relate this to the so-called ‘Mystery Interval’ – a particularly ill-defined and improperly named term. However, in both of these areas there is also a cold and arid phase in the period immediately pre-dating the period 23—19 ka. In the Southern Hemisphere tropics, the period around the LGM is complicated by a precession-driven insolation maximum which caused higher-than-present precipitation as monsoons strengthened (e.g. Baker et al., 2001a, 2001b; Cruz et al., 2005, 2009). Conversely, Whitney et al. (2011) found that vegetation records indicate that the LGM was drier than today in the Pantanal basin, Paraguay, highlighting the regional complexity of continental climates during the period around the LGM. Thus, it is clear from the small sample of vegetation records included here that pollen stratigraphy is unlikely to provide a unified signal of global climate change around the world that would facilitate defining a unified global LGM signal on land.

4.3. Loess

The loess sequences of the world have long been regarded as some of the best preserved records of continental global change (e.g. Kukla and An, 1989). In the Chinese loess, An et al. (1991) suggested that the maximum dust flux of the last glacial cycle was attained during the global Last Glacial Maximum, close to 20 ka. They took the median value of the loess grain size and the aeolian dust flux (bulk density multiplied by a linear deposition rate), as well as CaCO3 and magnetic susceptibility variations, to infer the general state of the atmosphere. On this basis, and constrained by two thermoluminescence ages, An et al. (1991) then tuned the loess record to the SPECMAP oxygen isotope curve. Porter (2001) noted that “alternating loess—paleosol stratigraphy closely resembles the marine oxygen-isotope record, implying that episodic dust deposition and pedogenesis are in phase with global ice-volume fluctuations”. This approach led to numerous other
papers, which hailed the loess records as some of the best terrestrial records available (cf. review by Porter, 2001).

Lu et al. (2007) used OSL to date rates of dust accumulation in the Chinese loess sequences. They found that high dust accumulation rates occurred during MIS 3, whereas the colder intervals of MIS 4 and 2 had lower dust accumulation rates. Lu et al. (2007) also questioned the relationship between loess grain size and dust flux. Regional variability in dust accumulation and questions over the relationship between loess grain size and atmospheric dust flux suggest that loess records may not be appropriate for defining global climatic events such as the global LGM. This evidence also highlights the problems in tuning loess records to those from the marine isotope profiles, which has been widely practised.

4.4. Speleothems

Speleothems provide a subterranean record of cave water chemistry, which can be related to conditions at the surface and in the atmosphere. The advantage of speleothems is that they can be dated with high precision using U-series techniques. This means that speleothem datasets have an independent chronology which “offers opportunities to critically assess leads and lags in the climate system” (McDermott, 2004, p.901). Unlike marine records, this means that speleothem records do not need to rely on ‘tuning’ isotopic signals to fit pre-existing concepts of climate change drivers, such as the orbital theory (McDermott, 2004). It has been shown that the isotope signal in speleothems has the potential to reflect global climatic changes. For example, at Hulu cave in SE China, the oxygen isotope record from a speleothem closely matches that recorded in the Greenland ice-cores (Wang et al., 2001). However, not all speleothems are suitable for defining the global LGM. Water in cave systems is likely to have frozen during the coldest part of the last glacial cycle in the mid- to high-latitudes and at high altitudes in the tropics, and “calcite deposition ceases when an area is subjected to glacial conditions” (Gascoyne, 1992, p. 614). Thus, the global LGM is often recorded by an hiatus in speleothem sequences in caves. This issue is less of a problem at lower latitudes. However, at many mid-latitudes calcite precipitation ceased during the LGM (e.g. Serefiddin et al., 2004; Constantin et al., 2007), whilst in others it continued (e.g. Moreno et al., 2010; Rowe et al., 2012). This limits the application of speleothems for the definition of such an interval.

4.5. Ice cores

Ice-core records provide information on the state of the atmosphere through time. Oxygen isotopes are used as an approximate proxy for air temperatures at the time of snow fall (Johnsen et al., 2001), Deuterium and nitrogen isotopes improve the precision of temperature reconstructions, in the Antarctic and Greenland ice cores, respectively (Jouzel et al., 2007; Kindler et al., 2013). Other indicators, such as dust concentrations, can also provide insight into the state of atmosphere through time (e.g. Ruth et al., 2007). The ice-core records from Greenland have been used to develop an event stratigraphy for the last glacial cycle (Björck et al., 1998; Walker et al., 1999; Lowe et al., 2001, 2008; Blockley et al., 2012). This existing framework can be explored as one approach available for defining the LGM and this is the focus of the next section.

5. Defining the LGM in the ice-core stratigraphy

Gibbard and West (2000, 2014) have argued that marine and terrestrial records both need to be considered in relation to the high-resolution ice-core records. In the Northern Hemisphere the Late Pleistocene has been subdivided into events based on the $^18O$ signal in the Greenland ice-core profile (Andersen et al., 2006; Rasmussen et al., 2006; Svensson et al., 2006, 2008; Lowe et al., 2008) and this record is also used to define the base of the Holocene Series (Walker et al., 2009). It is logical, therefore, to consider the Greenland Ice-Core Chronology 2005 (Andersen et al., 2006) for the definition of the boundaries of the LGM interval (Figs. 3 and 4) building on the already well-established event stratigraphy (Björck et al., 1998; Walker et al., 1999; Lowe et al., 2001, 2008; Blockley et al., 2012). The main basis for defining boundaries in the NGRIP ice-core records rests on the counting of annual ice layers. The dating of the period spanning 14.7−41.8 ka is described by Andersen et al. (2006) who define the onset of various Greenland Interstadials in this interval. The annually-counted $^18O$ signal in the GICC05 record is simply used as a time reference with which boundaries can be defined and some other criterion is required in order to define the LGM. However, to determine a global reference interval that characterises a distinct environmental signal such as the global Last Glacial Maximum both hemispheres need to be considered. The ice-core records from the Northern and Southern Hemisphere polar ice sheets provide an obvious target for this purpose. The Antarctic ice cores can also be used to define events and correlations between Greenland and Antarctic ice-core records that have been made by Blunier et al. (1998) and Blunier and Brook (2001).

It has been suggested that the events preserved in the Antarctic and Greenland ice-cores are asynchronous, with climate changes in the former leading the latter by 1−2.5 ka over the period 27−23 ka (Blunier et al., 1998). However, Steig and Alley (2003) argue that whilst the Antarctic and Greenland climate records display anti-phase behaviour about 50% of the time, they are generally in-phase during cooling. A later paper by EPICA (2006) showed that there was a one-to-one coupling of climate variability between Greenland and Antarctica, with warm events in the latter corresponding with cold events in the former. Fig. 3 shows that in both Antarctica and Greenland the coldest parts of the period 30−20 ka occurred during Greenland Stadial 2, although the patterns leading up to this interval are different with a significant cold event occurring in Greenland during Heinrich Event 3. However, for Greenland Fig. 3 only presents the oxygen isotope curve (which reaches its lowest point during Greenland Stadial 3, –46.50 $^18O$ at 26,560 ka, Andersen et al., 2006). Temperature reconstructions using $^{14}N$ indicate the lowest temperatures at c. 33 and 44 ka (Kindler et al., 2013). Thus the temperatures over the polar sheets are not suitable criteria in isolation for defining the global LGM.

Given the asynchronous millennial-scale climate changes recorded in the Antarctic and Greenland ice cores, rather than relying solely on isotopes (such as $^18O$, $^18D$, $^15N$) to match these records, a better tool for global correlation is dust flux, especially for identifying a world-wide hydrological phenomenon such as the global LGM. As noted above, the amount of dust in the atmosphere is influenced by climate and the extent of dust sources, which is strongly influenced by vegetation cover (which itself is climatically-driven) (Harrison et al., 2001). Vegetation cover is influenced by atmospheric moisture supply, which is in turn influenced by air and ocean-surface temperatures.

The dust concentration in the NGRIP core can be used as the main criterion to define the LGM event in the GICC05 record. These data are provided by Ruth et al. (2007). The peak dust concentrations in the NGRIP core were chosen as a marker for the LGM. Peak values of >8000 µg kg$^{-1}$ occur between 25.7955 and 25.4598 ka (in the GICC05 timescale). This peak in dust occurs in Greenland Stadial 3 between Greenland Interstadials 3 and 2. Thus, based on the event stratigraphy of the NGRIP GICC05 record, the global LGM can be correlated with Greenland Stadial 3 which spans the interval 27,540−23,340 ka (Lowe et al., 2008) (Figs. 3 and 4). Given that it is
desirable for the LGM to be defined as global signal in both hemispheres, the dust concentrations from the NGRIP core were compared with the record from EPICA (Antarctica) (Figs. 1 and 3). In the EPICA ice core record the highest, and most sustained, dust flux of the last glacial cycle also occurs between 27 and 24 ka during a time interval equivalent to Greenland Stadial 3 (Figs. 2 and 3).

The global LGM can be defined in terms of global dust flux and recorded by dust concentration or dust flux records in Greenland and Antarctica. In these records this occurred within Greenland the last glacial cycle, as indicated by the maximum extents of both was associated with the largest accumulation of ice on Earth during (Bond et al., 2009). This corresponds with the fact that Heinrich Event 2 (Björck et al., 1998) between 28 and 20 ka, the traditional timing for the end of the LGM. Thus, correlation with Greenland Stadial 3 in the ice-core record. However, as noted in the previous section, there are leads and lags between the isotopic signals in ice cores between Greenland and Antarctica. Boundaries of chronozones are required to be isochronous, i.e. time-parallel (Björck et al., 1998). However, it would not be possible to define globally isochronous boundaries for a chronozone based on isotopic signals in either Greenland or Antarctica.

Event stratigraphy offers a suitable alternative, especially since it is already applied for the Greenland ice-core record by Björck et al. (1998). In event stratigraphy the boundaries between events are not specifically designated and problems of time transgression that have arisen in applications of the terrestrial chronostratigraphy are no longer encountered (Björck et al., 1998, p. 289). In effect, this recognises that ‘events’ can be diachronous. An example of this is where event stratigraphy allows for eustasy transgressions and regressions (Whittaker et al., 1991). This is important because it is apparent from ice-core records from the two hemispheres that climate change is not coeval, especially at millennial timescales, which include distinct leads and lags. Whittaker et al. (1991, p. 820) noted that “perhaps the best prospects for high resolution event stratigraphy come from geochemical techniques, for instance by the use of stable isotopes such as oxygen, carbon, sulphur and strontium”. In this respect either marine or ice-core isotope curves can be used as a basis for event stratigraphy. However, the important difference between the ice-core and the marine isotopic records is that the former has much greater resolution and is most suited to defining fine-resolution stratigraphical intervals for the late Quaternary. As such, the base of the LGM is defined above using the Greenland ice-core stratigraphy and is based on the mid-point isotopic transition between Greenland Interstadial 3 and Stadial 3, in a similar way to the practice adopted for the subdivision of marine isotope records.

6. Event versus chronozone for the global LGM

The ice-core records provide a useful basis for defining the global LGM within an existing event stratigraphy developed for Greenland. However, there are limitations to relying solely on event stratigraphy for global correlation. For example, the global LGM until now was not clearly defined in the ice-core records. Relying on Greenland interstadials or stadials for global correlation could lead to the same erroneous correlations that have arisen from attempted equation with marine isotope stages. Environmental signals defined by isotopes, as atmospheric temperatures in the case of ice cores or ice volume in the case of marine sediments, may not translate as an environmental signal in other types of sequences. Hughes et al. (2013) made this point for glacier records. Despite these reservations regarding the use of isotopic records for global correlations, there is evidence that the Greenland ice-core isotopic signal is also represented in tropical cave speleothems, such as in Hulu Cave (Wang et al., 2001). Nevertheless, caution is required to ensure that environmental events in other types of record are not simply forced into line with the Greenland ice-core isotopic signal. Correctly defined chronostratigraphical units for the subdivision of the Late Pleistocene would avoid this potential pitfall. Previous definitions of the LGM have attempted to define the interval as a chronostratigraphical unit of unspecified rank, a chronozone spanning the intervals 24–18 or 23–19 ka (Mix et al., 2001). However, as discussed in Section 3 above, it is now questionable whether the criteria originally chosen by Mix et al. still fall within their suggested chronozone time span. Applying a chronozone status to the LGM using ice-core records is equally problematic. According to Salvador (1994, p. 83), “the time span of a chronozone is the time span of a previously designated stratigraphic unit or interval”. This previously designated stratigraphic unit or interval could be an event such as Greenland Stadial 3 in the ice-core record. However, as noted in the previous section, there are leads and lags between the isotopic signals in ice cores between Greenland and Antarctica. Boundaries of chronozones are required to be isochronous, i.e. time-parallel (Björck et al., 1998). However, it would not be possible to define globally isochronous boundaries for a chronozone based on isotopic signals in either Greenland or Antarctica.

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Pleistocene Subseries) and should not replace locally-defined terrestrial chronostratigraphical frameworks. Indeed, Lowe et al. (2008, p. 7) stated that “INTIMATE recommended that all site records should initially be designated using the appropriate local terminology, and that the timing and duration of local or regional climatic/environmental events be established independently of the ice-core record”. However, in the search for a definition of the global LGM this has to be made with reference to some globally recognisable stratigraphical sequence. If not, then the term ‘global LGM’ or even ‘LGM’ should be abandoned. For stratigraphical purists this may seem the best course of action, but a more pragmatic approach recognises that this will simply be ignored by the majority of the Quaternary community. Thus, the compromise is to lean on the ice-core event stratigraphy developed for Greenland and define the global LGM as coeval with Greenland Stadial 3. Whilst this itself is a regional environmental signal with evidence of complexity within this time interval in other types of record (Austin et al., 2012), it does allow for cross-correlation with similar records (i.e. dust flux) in Antarctica and thus provide an inter-hemispheric framework for defining the global LGM. Other local and regional climatic/environmental events should not be expected to fall within the narrow time span of a global LGM defined by Greenland Stadial 3. In fact, it should be expected that some local and regional environmental signals recorded by different types of Quaternary records will not coincide with the global LGM (i.e. Hughes et al., 2013).

It is now becoming clear that in many Quaternary records the global LGM was not always the most significant time interval marked by geomorphological, biological, climatological or other signals during the last glacial cycle. Recent advances in geochronological techniques means that Quaternary sediments and landforms can be dated with greater accuracy and precision. This technological advance has led to fundamental assumptions of time-equivalence being undermined. For example, in the glacier records what were assumed to be LGM moraines are now known to be much older in many parts of the world and in some cases (albeit only a few) evidence of an LGM advance is absent altogether (Hughes et al., 2013).

Given the statements above, it is tempting to assume too much focus on geochronology. To do this over and above a robust stratigraphical framework is likely to lead to problems similar to those which geochronological techniques have exposed. Geochronology should be tied to an independent stratigraphical framework that is both internally consistent (within the same types of record) and also externally consistent (allowing time-equivalent correlations between different types of record). It is becoming apparent that the previous basis for global Quaternary stratigraphy, the marine oxygen isotope record, was neither. This has led to erroneous correlations between many different types of records. There is a risk that future advances in Quaternary Science are driven by geochronology with stratigraphical considerations deemed obsolete. However, advances in geochronological techniques are reliant on robust stratigraphical frameworks. A good example is the current issue surrounding production rates in terrestrial cosmogenic nuclide exposure dating. In order to constrain production rates in surface rocks, surfaces of known age are required. In order to have faith in the geochronological technique the time-equivalence of the sample must be independently determined with confidence prior to calibration. Of equal geochronological/stratigraphical importance is that the geomagnetic field instabilities associated with both reversals and excursions lead to enhanced production of cosmogenic isotopes. Therefore the application of cosmogenic isotopes for surface exposure dating or as chronostratigraphical markers requires a complete understanding of geodynamic instabilities (Singer et al., 2009; Singer, 2014). Independent age confirmation is required for calibrating all dating techniques. Since all dating techniques require independent calibration, it would be erroneous to assume that one dating technique can simply be tested against another to ensure validation.

As outlined in this paper, definition of the LGM as an event within the Greenland ice-core record has the advantage of building on the existing event stratigraphy for the last glacial cycle that is already in place. However, it is only one possibility. Some may consider that the definition of the LGM as an equivalent within Greenland Stadial 3 as too narrow. This is perhaps true when considering the results of Shakun and Carlson (2010) who argued that the climatically-defined LGM should be placed at 22.2 ± 4.0 ka. Thus, broader definitions may be more appropriate. Whichever formal definition is chosen, this requires the contribution of the wider Quaternary community. This paper does not seek to impose any definition, simply to raise awareness that the current status of the LGM is insufficient and in need of revision and definition. Any definition will require the integration of marine, ice-core and terrestrial records and the collaboration of scientists with a wide range of specialisms.

7. Conclusions

The global LGM is not formally defined and this poses problems for correlation of global climatic events that are recorded in different types of sediments and landforms. Previous definitions proposed the LGM as a chronozone spanning the interval 23–19 ka. However, the fact that the LGM (and isotope boundaries in general) cannot be easily defined by isochronous boundaries within a sediment or ice-core sequence, means that it is difficult to define as a chronostratigraphical unit (Salvador, 1994, p. 78–85; Ehlers et al., 2011). Nevertheless, the global LGM can be defined within event stratigraphy comparable to that already adopted for the Greenland ice-core sequence (cf. Björck et al., 1998; Lowe et al., 2008). The global LGM can be defined as an event coeval with Greenland Stadial 3 (27.540–23.340 ka) – as already defined in the NGRIP Greenland ice-core stratigraphy. This event encompasses several key environmental global markers including the global sea-level lowstand (Peltier and Fairbanks, 2006; Thompson and Goldstein, 2006) and peaks in dust concentrations or dust flux recorded in the Greenland and Antarctic ice-core records (Ruth et al., 2007; Lambert et al., 2012). These environmental effects also coincide with the largest expansion of the Laurentide Ice Sheet to its maximum (Fig. 4) and other major ice masses, such as that over the British Isles and Ireland, which reached maximum extents at c. 24–25 ka (Balco and Schäfer, 2006; Stokes et al., 2012; Chiverrell et al., 2013).

This discussion demonstrates the difficulties and potential pitfalls that can arise when attempting to define a significant global event such as the LGM, an event which many would consider as one of the most strongly defined markers of the Late Pleistocene. The clear message is that problems comparable to those outlined in this article are very likely to hold for most, if not all, potentially global events that might be identified in the future. The implication, therefore, is that in the majority of cases, it cannot be assumed that a climate signal is likely to have a simple record across the globe.

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