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A stratigraphical basis for the Last Glacial Maximum (LGM)

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ABSTRACT

The Last Glacial Maximum (LGM) is widely used to refer to the episode when global ice volume last reached its maximum and associated sea levels were at their lowest. However, the boundaries of the interval are ill-defined and the term and acronym have no formal stratigraphical basis. This is despite a previous proposal to define it as a chronozone in the marine records on the basis of oxygen isotopes and sea levels, spanning the interval 23–19 or 24–18 ka and centred on 21 ka. In terrestrial records 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. However, variations in the global hydrological system during glacial cycles are recorded by atmospheric dust flux and this provides a signal of terrestrial changes. Whilst regional dust accumulation is recorded in loess deposits, *global* dust flux is best recorded in high-resolution polar ice-core records, providing an opportunity to define the LGM on land and establish a clear stratigraphical basis for its definition. On this basis, one option is to define the global LGM as an event between the top (end) of Greenland Interstadial 3 and the base (onset) of Greenland Interstadial 2, spanning the interval 27.540–23.340 ka (Greenland Stadial 3). This corresponds closely to the peak dust concentration in both the Greenland and Antarctic ice cores and to records of the global sea-level minima. This suggests that this definition includes not just the coldest and driest part of the last glacial cycle but also the peak in global ice volume. The later part of the LGM event is marked by Heinrich Event 2, which reflects the onset of the collapse of the Laurentide at c. 24 ka, together with other ice sheets in the North Atlantic region. A longer and later span for the LGM may be desirable, although defining this in chronostratigraphical terms is problematic. Whichever formal definition is chosen, this requires the contribution of the wider Quaternary community.

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

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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 ^{14}C ka BP, with a mid-point at 18,000 ^{14}C 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 δ^{18}O 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 foraminifera 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 δ^{18}O 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 δ^{18}O to shifts in global ice volume stills holds. Since benthic δ^{18}O variability is driven by global ice volume, it also provides a globally synchronous record of glacioeustasy (Skinner and Shackleton, 2005). Whilst this is true when considering changes over 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 isotopes such as benthic δ^{18}O cannot provide a globally correlative 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 δ^{18}O in marine foraminiferal records, is a deep-water temperature rather than an ice-volume signal. Skinner and Shackleton (2005) showed that fluctuations in benthic δ^{18}O and MIS boundaries from different hydrological settings may be significantly diachronous. The use of benthic δ^{18}O 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 δ^{18}O signal has been constructed by assuming that solar forcing paces variations in δ^{18}O 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 δ^{18}O 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 δ^{18}O at c. 23.1 ka, the lowest sea levels bracketed between 23.1 and 24.6 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