JOURNAL OF QUATERNARY SCIENCE (2009) **24**(1) 3–17

Published online 3 October 2008 in Wiley InterScience
(www.interscience.wiley.com) DOI: 10.1002/jqs.1227

Formal definition and dating of the GSSP (Global Stratotype Section and Point) for the base of the Holocene using the Greenland NGRIP ice core, and selected auxiliary records

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Walker, M., Johnsen, S., Rasmussen, S. O., Popp, T., Steffensen, J.-P., Gibbard, P., Hoek, W., Lowe, J., Andrews, J., Björck, S., Cwynar, L. C., Hughen, K., Kershaw, P., Kromer, B., Litt, T., Lowe, D. J., Nakagawa, T., Newnham, R., and Schwander, J. 2009. Formal definition and dating of the GSSP (Global Stratotype Section and Point) for the base of the Holocene using the Greenland NGRIP ice core, and selected auxiliary records. *J. Quaternary Sci.*, **Vol. 24** pp. 3–17. ISSN 0267-8179.

Received 18 June 2008; Accepted 26 June 2008

ABSTRACT: The Greenland ice core from NorthGRIP (NGRIP) contains a proxy climate record across the Pleistocene–Holocene boundary of unprecedented clarity and resolution. Analysis of an array of physical and chemical parameters within the ice enables the base of the Holocene, as reflected in the first signs of climatic warming at the end of the Younger Dryas/Greenland Stadial 1 cold phase, to be located with a high degree of precision. This climatic event is most clearly reflected in an abrupt shift in deuterium excess values, accompanied by more gradual changes in $\delta^{18}O$, dust concentration, a range of chemical species, and annual layer thickness. A timescale based on multi-parameter annual layer counting provides an age of 11 700 calendar yr b2 k (before AD 2000) for the base of the Holocene, with a maximum counting error of 99 yr. A proposal that an archived core from this unique sequence should constitute the Global Stratotype Section and Point (GSSP) for the base of the Holocene Series/Epoch (Quaternary System/Period) has been ratified by the International Union of Geological Sciences. Five auxiliary stratotypes for the Pleistocene–Holocene boundary have also been recognised. Copyright © 2008 John Wiley & Sons, Ltd.

Journal of Quaternary Science

KEYWORDS: Holocene boundary; Global Stratotype Section and Point; NGRIP ice core; auxiliary stratotypes.

Introduction

Holocene (meaning 'entirely recent') is the name given to the most recent interval of Earth history, which extends to and

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includes the present day. It is the second series or epoch of the Quaternary System/Period, and is perhaps the most intensively studied interval of recent geological time. The Holocene stratigraphic record contains a wealth of detail on such diverse phenomena as climate change, geomorphological and geophysical processes, sea-level rise, vegetational developments, faunal migrations and, not least of all, human evolution and activity. Despite this rich and varied palaeoenvironmental archive, and in spite of the fact that the Holocene has long been recognised as a stratigraphic unit of series status (see below), a definition of the base of the Holocene (the Pleistocene–Holocene boundary) has previously never been formally ratified by the International Union of Geological Sciences.

In 2004, a Joint Working Group of the North Atlantic INTIMATE programme (Integration of ice-core, marine and terrestrial records: INQUA project 0408) and the SQS (Subcommission on Quaternary Stratigraphy) was established to bring forward a proposal for a definition of the Pleistocene–Holocene boundary (Global Stratotype Section and Point, GSSP) based on the NorthGRIP (NGRIP) Greenland ice core; five auxiliary stratotypes were recommended to support the proposal. This proposal was reviewed and formally accepted by the Subcommission on Quaternary Stratigraphy and subsequently by the International Commission on Stratigraphy (ICS). Details of the voting can be found on the SQS website (http://www.quaternary.stratigraphy.org.uk). In May 2008, the Executive Committee of the International Union of Geological Sciences (IUGS) ratified the Holocene GSSP located at 1492.45 m depth within the NGRIP ice core, Greenland. A preliminary account of the new GSSP was published in a special issue of *Episode*s (Walker *et al*., 2008), but the full proposal, as submitted to the IUGS, is published for the first time in this paper.

Terminology

The origin of the term 'Holocene' is inextricably linked to the development of the nomenclature for ice-age time divisions. In the early to mid-19th century, two terms emerged independently to encompass near-surface, largely unconsolidated deposits: *Quaternary* (von Morlot, 1854; Desnoyers, 1829), and *Pleistocene* (Lyell, 1839), although the former considerably pre-dates this usage, having been employed by Giovanni Arduino as early as 1759 to describe the fourth stage or 'order' that he identified in the alluvial sediments of the River Po in northern Italy (Schneer, 1969; Rodiloco, 1970). The terms 'Quaternary' and 'Pleistocene' were initially applied to marine deposits, the former in the Paris Basin and the latter in eastern England. However, with the recognition that the extensive 'Drift' or 'Diluvial' deposits, until then considered to be marine, were the products of extensive recent glaciation, both soon became synonymous with the 'Ice Age' and with the evolution of humans. The Quaternary differed from the Pleistocene, however, in that it included Lyell's (1839) 'Recent' or Forbes' (1846) 'Post-glacial', a period that was subsequently termed 'Holocene' by Gervais (1867–69). The last-named term was formally adopted by the International Geological Congress (IGC) in 1885. Indeed, Parandier (1891) went further in that he considered the Holocene period to follow the Quaternary, and hence to constitute a fifth system ('Quinquennaire'), but this terminology was not subsequently adopted (Bourdier, 1957). In view of the near-temporal parallelism of the Quaternary and Pleistocene, numerous proposals have been made to discontinue usage of one or other of the terms, while alternatives such as 'Anthropogene' or 'Pleistogene' have been advocated, albeit with limited success (Gibbard and van Kolfschoten, 2005). In practice, both 'Quaternary' and 'Pleistocene' have been employed over the course of the past century, the former being ascribed the rank of period/system within the Cenozoic Era and containing two separate series/epochs: the Pleistocene and the Holocene with the latter, as noted above, extending to the present day (Gibbard *et al*., 2005).

Three terms continue to be used as alternatives to Holocene, however. *Recent* and *Post-glacial* (see above) are widespread in the literature, but both have no formal status and are therefore invalid in stratigraphic nomenclature (Gibbard and van Kolfschoten, 2005). A third term, *Flandrian*, derives from marine transgression sediments on the Flanders coast of Belgium (Heinzelin and Tavernier, 1957), and follows European practice in naming temperate stages of the Quaternary after their characteristic marine transgressions (e.g. Holsteinian, Eemian). The term has frequently been employed as a synonym for Holocene (e.g. Nilsson, 1983, p. 23), particularly by those who consider the present temperate episode to have the same stage status as earlier interglacial periods. In these circumstances, the Flandrian becomes a designated stage within the Pleistocene, in which case the latter extends to the present day (cf. West, 1968, 1977, 1979; Woldstedt, 1969; Hyvärinen, 1978). This interpretation has not found universal acceptance, however (e.g. Morrison, 1969), and the terminology has not been widely used outside the British Isles (Flint, 1971, p. 324; Lowe and Walker, 1997, p. 16). Accordingly, the term *Holocene Series/Epoch* is preferred in recognition not only of modern worldwide usage, but also to acknowledge the importance of the present interglacial in terms of its distinctive palaeoenvironmental and unique anthropological record.

It should also be noted that the Holocene as defined above does not begin at the termination of the last cold stage around 14 500 cal. yr BP. Rather, it begins at the end of the Younger Dryas/Greenland Stadial 1 (GS-1), an anomalous cooling episode that interrupts the postglacial warming. This short-lived cold event, widely dated by radiocarbon to between ca. 12 900 and 11 500 cal. yr BP, is most strongly recorded in proxy records from the northern mid- and high-latitude regions.

The Pleistocene–Holocene boundary

The conventional approach to the subdivision of the Quaternary stratigraphic record is to employ evidence for contrasting climatic conditions to characterise individual stratigraphic units (*geologic-climatic units*). This follows recommendations by the American Commission on Stratigraphic Nomenclature (1961, 1970) that recognised a geologic-climatic unit as indicating 'an inferred widespread climatic episode defined from a subdivision of Quaternary rocks' (ACSN, 1970, p. 31). In subsequent formal stratigraphic codes, however (Hedberg, 1976; North American Commission on Stratigraphic Nomenclature, 1983; Salvador, 1994), the climatostratigraphic approach has not been employed because of the prevailing view that, for most of the geological record, 'inferences regarding climate are subjective and too tenuous a basis for the definition of formal geologic units' (North American Commission, 1983, p. 849). This position does not find favour with Quaternary scientists, however, who contend that, as climatic change is the hallmark of the past two million years, it is difficult to envisage a scheme of stratigraphic subdivision for recent earth history which does not specifically acknowledge that fact (Lowe and Walker, 1997). Indeed, it now appears that the principal driver of recent global climatic change, namely variations in the Earth's orbit and axis (Milankovitch forcing), is reflected not only in key stratigraphic horizons within the Quaternary record, but also can be detected in pre-Quaternary sequences as far back as the late Palaeocene (Lourens *et al.*, 2005). 'Astronomical pacing' provides both an index of the major climatic shifts and a basis for estimating the age of these events. As a consequence, major intervals of geological time within the Neogene and the Quaternary, which are evident in the climatic record, can be further defined as chronostratigraphic units (Zalasiewicz *et al.*, 2004). For these reasons, Quaternary scientists continue to employ geologic-climatic units based on proxy climatic indicators as the principal means of subdividing the Quaternary stratigraphic sequence, and hence the Pleistocene–Holocene boundary (the base of the Holocene Series/Epoch) is defined in this proposal on the basis of its clear climatic signature. Potential depositional contexts in which the GSSP for the base of the Holocene might be located are considered in the following sections.

Marine sediments

It is normal practice in stratigraphy to delimit system, stage and lesser chronostratigraphical and geochronological boundaries, as far as possible, in continuous or near-continuous sedimentary sequences in which multiple lines of evidence are preserved. Throughout the geological column, these requirements are best met in marine sediment successions where the assumed continuity of sedimentation allows the palaeontological, isotopic, chemical and palaeomagnetic changes to be determined, and hence any major boundary to be defined precisely within a multi-proxy context. During the Phanerozoic, this approach not only enables geological boundaries to be accurately determined, but it also provides a basis for correlation between the stratotypes and boundaries that have been established in other sequences.

In many geological systems or periods, these boundaries are often placed in what would be regarded by Quaternary geologists as relatively shallow-water or shelf-sediment sequences. However, the ubiquity of Quaternary sedimentation means that the period is represented in all marine sedimentary contexts, including ocean floor, continental slope and continental shelf, near-shore littoral and beach deposits. The Pleistocene–Holocene transition is therefore found in a wide spectrum of depositional contexts and, in theory, the boundary stratotype could be located in any of these. In practice, however, this is problematical. In deep-ocean sediments, for example, while the formally defined geologic-climatic units of the marine isotope stages (MIS) are pivotal to stratigraphic subdivision and correlation over most of the Quaternary time range (e.g. Shackleton *et al.*, 1990), in our view it would *not* be appropriate to define the base of the Holocene on the basis of the deep-ocean MIS record. There are two reasons for this: (1) many of the requirements of a GSSP, such as a sufficiently rapid rate of sedimentation, the assured continuity of sedimentation and a permanently fixed marker, are not met in deep-ocean sediments of Holocene age; and (2) there are considerable difficulties in determining the precise and accurate age of the Pleistocene–Holocene boundary in deep-ocean sediments, principally because radiocarbon dating, the most widely used method for the dating of late Quaternary marine sequences, is constrained by a range of methodological problems, most notably a spatially variable marine reservoir effect (Waelbroek *et al.*, 2001; Björck *et al.*, 2003; Hughen *et al.*, 2004b), and a 600 yr long 'radiocarbon age plateau' (discussed below, but also applicable to marine archives).

Difficulties also arise when marine shelf sediments are being considered as potential reference localities for the base of the Holocene. In the 1960s to 1980s, for example, attempts were made to define a Holocene basal boundary stratotype in a borehole sequence (Core B 873) from the Göteborg Botanical Garden in southwestern Sweden, where shallow-water marine deposits and glacio-lacustrine sediments, some of which are varved, occur above present sea level as a consequence of isostatic uplift at the end of the Late Pleistocene Weichselian Stage (Mörner, 1976). However, the poorly defined nature of the boundary in Core B 873, in terms of both lithostratigraphy and biostratigraphy (pollen and diatoms), the absence of radiocarbon dates from the core and problems associated with the interpretation of the palaeomagnetic record (Hyvärinen, 1976; Thompson and Berglund, 1976), led to the rejection of this sequence as a GSSP (Olausson, 1982). Similarly, two cores with marine sediments from a site to the north of Göteborg that had been proposed as a potential boundary stratotype and hyperstratotype for the Pleistocene–Holocene transition (Olausson, 1982) could not be accepted as such, again because of problems associated with radiocarbon dating and palaeoenvironmental interpretation. Elsewhere, shelf environments in general have proved to be unsuitable localities for defining Quaternary boundary stratotypes because glacio-eustatic lowstands during glacial maxima have exposed shelf areas to processes of subaerial erosion down to depths of $\sim$130 m below sea level. The combined effects of subsequent sea-level rise, neotectonics and/or isostatic crustal rebound during the Late Pleistocene and Holocene have resulted in shelf sediment sequences characterised by numerous marked erosional gaps, discontinuities, abrupt changes in sedimentation rate and complex facies variations. Collectively, therefore, these complicating factors mean that many Quaternary marine shelf successions are generally unsuitable for the siting of a GSSP locality.

A further difficulty in basing the Pleistocene–Holocene boundary on marine sediment records is that during the last deglaciation large quantities of cold fresh water, either from melting glacier margins or from lakes dammed by the ice, were released into the northern oceans (Clark *et al.*, 2001; Teller *et al.*, 2002). As a consequence, the global climatic warming signal in marine core records that reflects the onset of the Holocene may be masked by the regional or, in some cases, hemispherical-scale effects of the cooling of ocean waters that occurred during deglaciation.

Terrestrial sediments

For many years Quaternary scientists have sought a boundary stratotype for the Holocene in terrestrial sedimentary records. Various depositional contexts have been proposed, including boundaries of till units, palaeosols and littoral indicators of sea-level change but, for a variety of reasons, none of these has proved to be satisfactory (Morrison, 1969; Bowen, 1978). Particular attention has been directed towards depositional sequences in lakes as these frequently contain a record of continuous sediment accumulation across the Pleistocene–Holocene boundary, as well as a range of proxy climate indicators, including sedimentological/lithological, geochemical, isotopic and biological records. The latter provide the evidence for the climatic change that marks the beginning of the Holocene. A particularly widely used climate proxy has

DOI: 10.1002/jqs

been pollen data. In Scandinavia, for example, these have been employed to define the boundaries between European pollen zones III and IV; the boundary between the Younger Dryas and Preboreal chronozones; and the Lateglacial/Holocene boundary (Mangerud *et al.*, 1974; Mörner, 1976).

Although the climatic shift that marks the beginning of the Holocene can be readily identified in many limnic sedimentary sequences, no formally defined boundary stratotype based on these proxy climate records has yet been proposed. Indeed, in the absence of such a proposal, the Holocene Commission (meeting at the 1969 INQUA Congress in Paris) recommended that the Pleistocene–Holocene boundary should be defined *chronometrically* and placed at 10 000 ^{14}C yr BP (Hageman, 1969), and this remains the situation at the present day. As such, it was the first stratigraphic boundary later than the Proterozoic to be defined in this way (Harland *et al.*, 1989). It has been suggested, however, that by defining the base of the Holocene in terms of radiocarbon years, the INQUA Commission anticipated the discovery of a suitable stratotype (Bowen, 1978). The latter is clearly desirable for otherwise the Holocene, as a unit of geochronological time, would be unique in the geological record in having no chronostratigraphic standard as a basis for comparison.

During the 1980s, it was assumed that the Pleistocene–Holocene boundary would be most satisfactorily defined using varved glacio-lacustrine sequences in Sweden, although more recent work suggests that the annually laminated lacustrine records from western Germany may offer a better prospect for a boundary stratotype (Litt *et al.*, 2001). While limnic sediment sequences *potentially* constitute a useful basis for defining the base of the Holocene and, indeed, apparent synchroneity can be demonstrated between climatic proxies in Swedish lake sediments and those in other archives (German tree rings and the GRIP δ^{18}O record: Björck *et al.*, 1996), more often the *key* climate signal (i.e. the first indication of early Holocene temperature rise) may be difficult to isolate. This is because the different rates of response of biotic and abiotic environmental systems to climate forcing will lead to temporal (and spatial) lags in the proxy records. Hence, time transgression, a phenomenon that can be readily measured in late Quaternary biostratigraphic records, presents major problems for defining the lower Holocene boundary using proxy climate data from limnic sedimentary sequences (Watson and Wright, 1980; Björck *et al.*, 1998). Moreover, the climate signal itself may be further compromised by the effects of sedimentary and taphonomic processes. Defining a GSSP for the base of the Holocene using lake sediment records is, therefore, not straightforward, although the lithostratigraphic approach to determining the boundary may, perhaps, present fewer difficulties than that based on biostratigraphic evidence (Björck *et al.*, 1996). The geochronology of the Pleistocene–Holocene transition is also problematic, for radiocarbon dating of lake sediments is adversely affected both by technical limitations (mineral carbon error, contamination, etc.) and by the presence of a 600-year long 'radiocarbon age plateau' (a period of constant radiocarbon age) at the beginning of the Holocene (Ammann and Lotter, 1989; Björck *et al.*, 1996; Lowe and Walker, 2000; Lowe *et al.*, 2001). As a consequence, there is frequently a spread of radiocarbon ages around the Pleistocene–Holocene boundary. To some extent this problem can be mitigated by 'wiggle-matching' a series of radiocarbon dates from a suitable sequence to the dendrochronologically based calibration curve (e.g. Björck *et al.*, 1996; Gulliksen *et al.*, 1998; Wohlfarth *et al.*, 2006; Lowe DJ *et al.*, 2008) and/or by employing annually laminated sediment sequences to provide an independent timescale (e.g. Hajdas *et al.*, 1995; Goslar *et al.*, 1995; Litt *et al.*, 2001, 2003). Nevertheless, as was the case with ocean cores, establishing an appropriate GSSP for the Holocene on the basis of a limnic sequence is likely to be compromised by the considerable uncertainties relating to chronology.

Glacier ice

One context in which many of the difficulties encountered with marine and terrestrial records can be overcome, and where there is considerable potential for defining a GSSP for the base of the Holocene, is the polar ice archive. The Greenland ice-sheet records extend back over a hundred thousand years (Hammer *et al.*, 1997; Alley, 2000a; Mayewski and White, 2002; North Greenland Ice Core Project Members, 2004), while in Antarctica the ice archive spans several glacial–interglacial cycles (Petit *et al.*, 1999), with the longest core record extending back almost one million years (EPICA community members, 2004). The ice sheets contain an unparalleled range of climatic proxies, including stable isotopes, trace gases, aerosols and particulates, and also preserve a continuous record of snow accumulation, usually in a very highly resolved and continuous stratigraphic sequence. Moreover, ice-core records can be dated by a number of independent methods, including visible ice-layer counting, oxygen isotope and chemical stratigraphy, ice-flow modelling and tephrochronology.

In Greenland, six major drilling programmes (Fig. 1) have been undertaken over the course of the last 40 yr (Johnsen *et al.*, 2001). The longest cores have been obtained from the summit of the ice sheet, where over 3 km of ice have accumulated. The GRIP (Greenland Ice Core Project) project, primarily a European consortium, drilled through the ice between 1989 and 1992 and reached bedrock at 3027 m, while the American GISP2 (Greenland Ice Sheet Project 2) core was drilled between 1989 and 1993 and reached bedrock at 3053 m (Hammer *et al.*, 1997). In 2003, a new core was drilled at NorthGRIP (NGRIP) (Figs. 2 and 3), which is located ~350 km NW of the GRIP/GISP2 sites. This is the deepest core so far recovered from Greenland (3085 m), and the base is dated to ca. 123 k yr BP (Dahl-Jensen *et al.*, 2002; North Greenland Ice Core Project Members, 2004). A combination of moderate accumulation rates (19 cm of ice equivalent per year at present) and bottom melting results in average annual layer thicknesses under glacial conditions that are greater than those in either the GRIP or GISP2 ice cores. In addition, the development of new high-resolution impurity measurement techniques makes the NGRIP core ideal for stratigraphic dating purposes. Based on data from NGRIP, the Copenhagen Ice Core Dating Initiative has developed a high-resolution stratigraphic timescale for the NGRIP and GRIP ice cores. This new timescale is known as the Greenland Ice Core Chronology 2005, or GICC05 (Andersen *et al.*, 2006; Rasmussen *et al.*, 2006; Vinther *et al.*, 2006; Svensson *et al.*, 2008). The NGRIP core also contains the most highly resolved stratigraphic record in any of the Greenland ice cores of the transition from the Pleistocene to the Holocene, and this is apparent in both the visual stratigraphy (Fig. 4) and in a range of chemical indicators (Fig. 5 and see below). Accordingly, we proposed that the boundary stratotype for the base of the Holocene (GSSP) should be defined on the basis of the stratigraphic record in the NGRIP ice core.

At first sight it may seem inappropriate to advocate a global geological stratotype on the basis of an ice-core sequence. There are, however, a number of sound reasons for such a proposal:

DOI: 10.1002/jqs

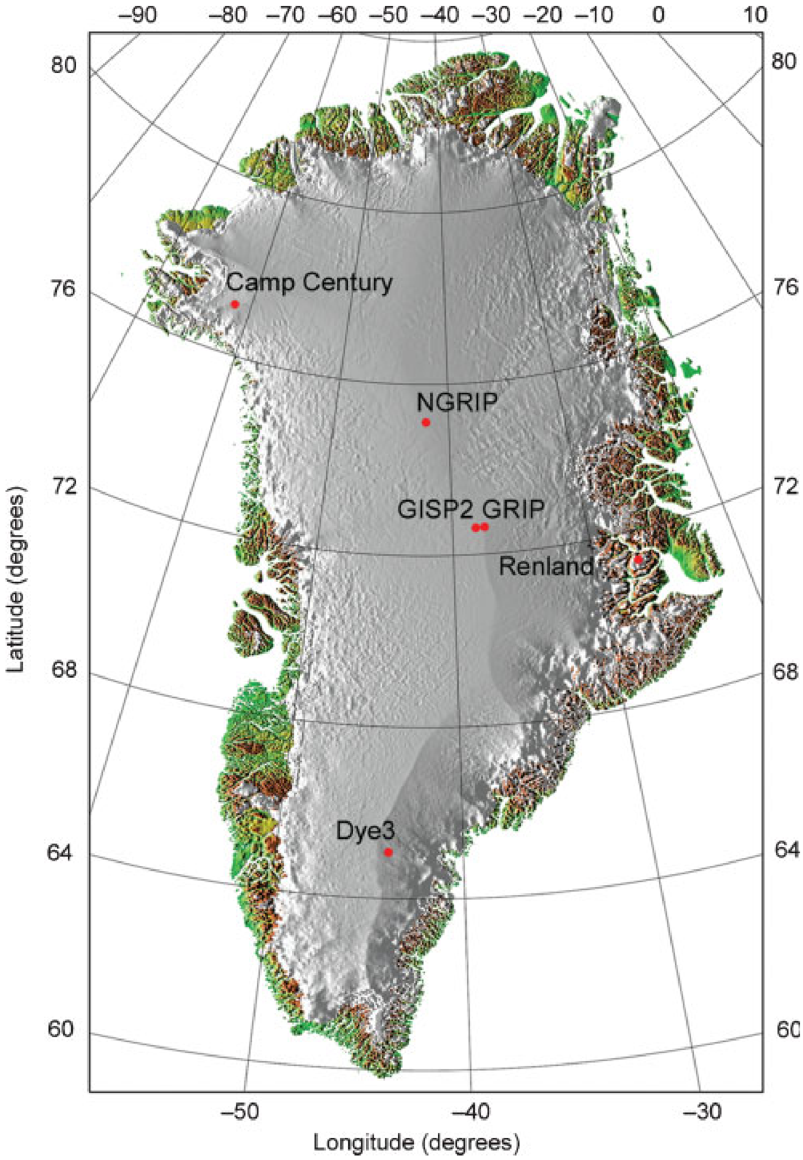

Figure 1 The locations of five deep drilling sites on the Greenland ice sheet: NGRIP (75.1° N, 42.3° W), GRIP (72.5° N, 37.3° W), GISP2 (72.5° N, 38.3° W), Dye-3 (65.2° N, 43.8° W) and Camp Century (77.2° N, 61.1° W). Also shown is the shallower site (324 m) of Renland (71.3° N, 26.7° W). At all of these sites, the ice-core record extends back to the last (Eemian) interglacial. Map by S. Ekholm, Danish Cadastre

1. Because glacier ice is a sediment, defining the Holocene boundary stratotype in an ice core is as justified as basing a stratotype on hard or soft rock sequences. Moreover, while it might be argued that there is a potential impermanence about the Greenland ice sheet (i.e. it could melt with accelerated global warming), it could equally be argued that no conventional geological section or exposure has an assured permanence. Quarrying, flooding, erosion or instability/collapse are just four of the potential threats to the preservation of a conventional GSSP.
2. Ice sheets form through the annual incremental accumulation of snow. This means that there is a continuity of accumulation (sedimentation) across the Pleistocene–Holocene boundary. Moreover, the relatively heavy snowfall at the NGRIP site at the end of the last cold stage also means this key part of the record is very highly resolved – far more so than is the case with many terrestrial sedimentary contexts.
3. Because of its geographical location in the high-latitude North Atlantic, Greenland is a sensitive indicator of hemispherical-scale climate change. This is likely to have been even more the case at the Pleistocene–Holocene transition when the Greenland ice sheet lay mid-way between the wasting Eurasian and Laurentide ice masses. As noted above, a range of climate proxies is preserved within the Greenland ice, a number of which are immediately responsive to climate shifts. Accordingly, early signals of the marked climatic change that defines the onset of the Holocene Epoch will be registered very clearly in the Greenland ice-core record.
4. If defined on the basis of the NGRIP climatic record, the base of the Holocene can be very precisely dated by annual ice-layer counting (see below), and can be replicated in other Greenland ice-core records (see Greenland Ice Core Chronology 2005/GICC05, above). The boundary stratotype for the Holocene (GSSP) is therefore at a level of chronological precision that is unlikely to be attainable in any other terrestrial stratigraphic context.
5. The Greenland ice-core record has been proposed by an INQUA project group (INTIMATE) as the stratotype for the Late Pleistocene in the North Atlantic region (Walker *et al.*, 1999) and an 'event stratigraphy' was initially developed for the Last Termination based on the oxygen isotope record in the GRIP ice core (Björck *et al.*, 1998). More recently, the

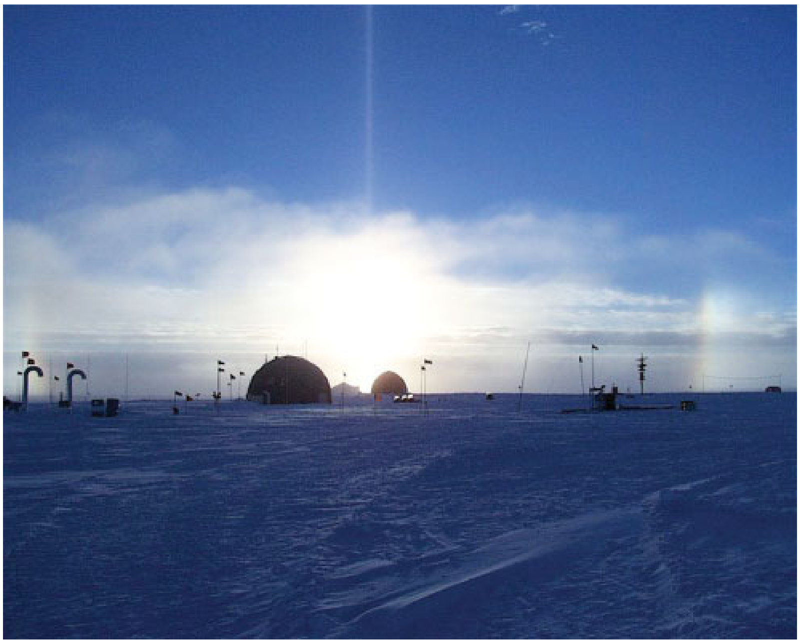

Figure 2 The NGRIP drilling camp on the summit of the Greenland ice sheet. The large structure is the camp main building ('Main Dome'); the two other domes are the workshop and storage buildings. The flags and pipes in the foreground mark the location of the subsurface drill and science trenches (Photo Centre for Ice and Climate: www.iceandclimate.dk)

Figure 3 Ice coring. (a) The drill being tilted into position. (b) An ice core in the drill prior to extrusion. (c) An extruded ice core (Photo Centre for Ice and Climate: www.iceandclimate.dk)

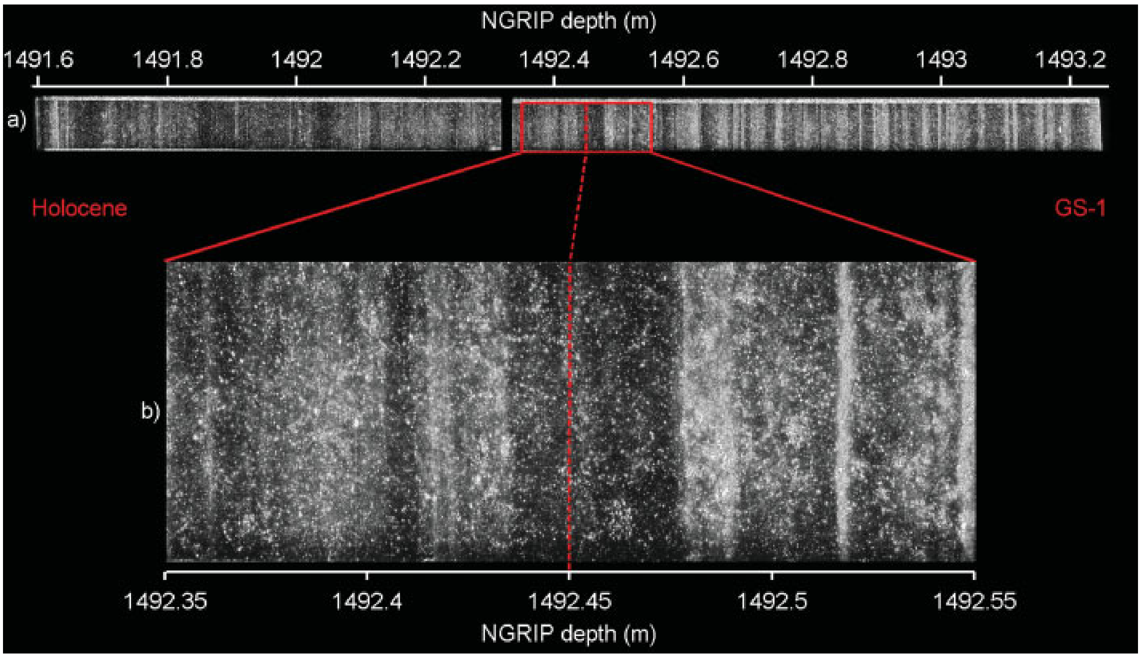

Figure 4 The visual stratigraphy of the NGRIP core between 1491.6 and 1493.25 m depth obtained using a digital line scanner (Svensson *et al.*, 2005). In this photograph, the image is 'reversed' so that clear ice shows up black, whereas the cloudy bands, which contain relatively large quantities of impurities, in particular micrometre-sized dust particles from dry areas in eastern Asia, appear white. The visual stratigraphy is essentially a seasonal signal and reveals annual banding in the ice. The location of the Pleistocene–Holocene boundary at 1492.45 m is shown in the enlarged lower image. A core break occurs at a depth of 1492.32 m. The ice core is complete and continuous across the core break, but the visual stratigraphy scanning image is disturbed by the break and has thus been masked out

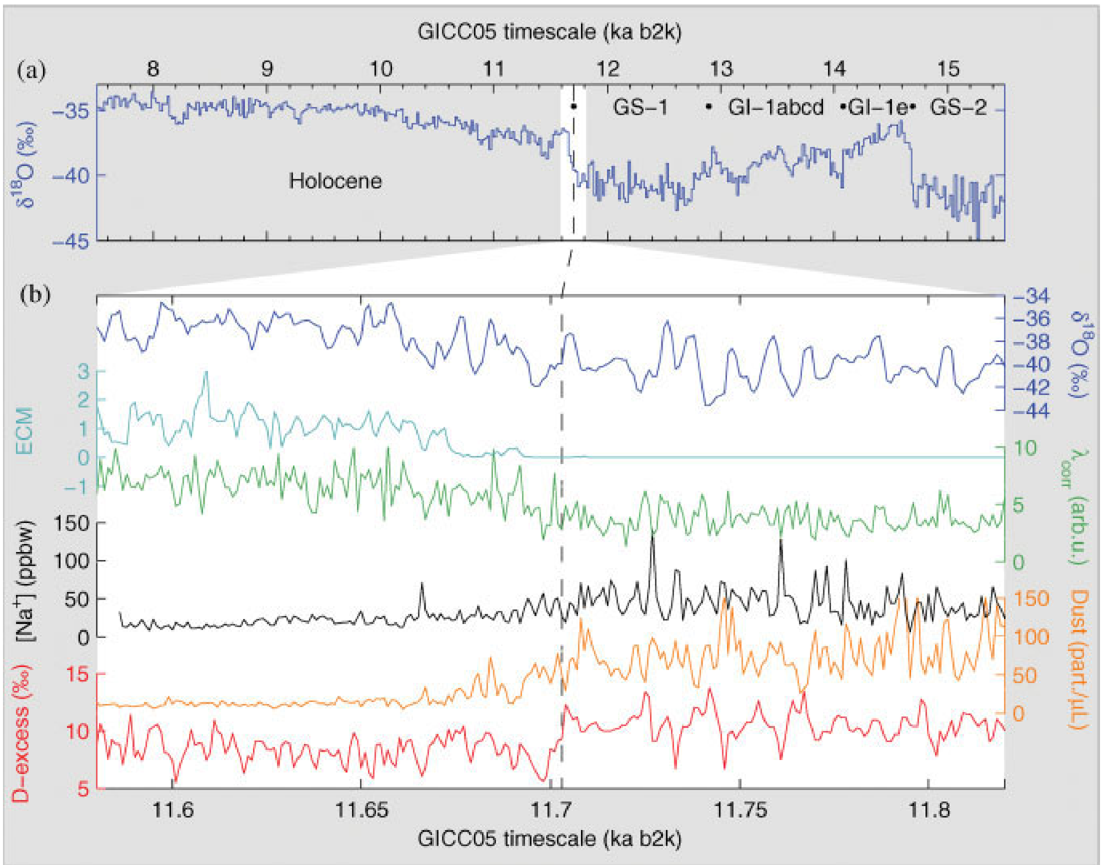

Figure 5 (a) The $\delta^{18}O$ record through the Last Glacial–Interglacial Transition, showing the position of the Pleistocene–Holocene boundary in the NGRIP core. (b) High-resolution multi-parameter record across the Pleistocene–Holocene boundary: $\delta^{18}O$, electrical conductivity (ECM), annual layer thicknesses corrected for flow-induced thinning (λ_{corr}) in arbitrary units, Na^{+} concentration, dust content, and deuterium excess

INTIMATE group has proposed that the new NGRIP isotopic record should replace GRIP as the stratotype, using the GICC05 chronology (Lowe JJ *et al.*, 2008). The establishment of a GSSP in the NGRIP ice core is entirely in keeping with these recent approaches to stratigraphic subdivision of the Late Pleistocene record based on the Greenland ice-core sequence (Lowe JJ *et al.*, 2008).

6. The role of a stratotype is to act as a fixed point for correlation, a procedure that is frequently (although not always) effected using biostratigraphic evidence. As there is little or no fossil evidence in Greenland ice, it could be argued that this would make an ice core unsuitable for a GSSP. As discussed above, however, the Pleistocene–Holocene boundary is not being defined on the lithostratigraphic or biological evidence *sensu stricto*, but rather on the *climatic signal* that is reflected in a range of proxy climatic indicators. Greenland ice constitutes a climatic archive which is unique in terms of its range, resolution and sensitivity, as well as the precision with which it can be dated and, as such, constitutes a climatostratigraphic reference point to which all other climatic proxies, from both the marine and terrestrial realms, can readily be correlated (e.g. Rohling *et al.*, 2002; van der Plicht et al., 2004). In addition, using variations in the time series of atmospheric gas records as indirect time markers, Greenland ice-core records can be synchronised with those from other regions, notably Antarctica (Blunier *et al.*, 2007).

Proposed GSSP for the base of the Holocene Series/Epoch

The details of the proposal can be found in the following section, which follows the ISC protocol for the recommendation of boundary stratotypes:

Name of the boundary

The base of the Holocene (Pleistocene–Holocene boundary) Series/Epoch.

GSSP definition

The base of the Holocene Series/Epoch is defined in the NGRIP ice-core record at the horizon which shows the clearest signal of climatic warming, an event that marks the end of the last cold episode (Younger Dryas Stadial/Greenland Stadial 1) of the Pleistocene (see details below and Fig. 5).

Stratigraphic Rank and Status of Boundary

Boundary stratotype: Global Stratotype Section and Point (GSSP).

Stratigraphic position and nature of the defined unit

In the Greenland ice cores, the Pleistocene–Holocene transition is chronologically constrained between two clearly defined tephra horizons: the Saksunarvatn tephra (1409.83 m depth) and the Vedde Ash (1506.14 m depth). These are dated at 10 347 yr b2 k (counting uncertainty 89 yr) and 12 171 yr (counting uncertainty 114 yr) b2 k, respectively. The term b2 k refers to the ice-core zero age of AD 2000; note that this is 50 years different from the zero yr for radiocarbon, which is AD 1950 (the derivation of the NGRIP timescale is explained below). Both tephras are found in terrestrial and marine records throughout the North Atlantic region (e.g. Davies *et al.*, 2002; Blockley *et al.*, 2007) and provide the basis for precise correlation between other sedimentary archives and the ice-core sequence. The transition to the Holocene is marked by a shift to 'heavier' oxygen isotope values ($\delta^{18}O$) between Greenland Stadial 1 (GS-1)/Younger Dryas ice and ice of early Holocene age; a decline in dust concentration from GS-1 to modern levels; a significant change in ice chemistry (e.g. reduction in sodium (sea-salt) values); and an increase in annual ice-layer thickness (Johnsen *et al.*, 2001; Steffensen, 2008; Fig. 5(b)). These various data sources reflect a marked change in atmospheric circulation regime accompanied by a temperature rise, probably of the order of $10 \pm 4^\circ C$, at the onset of the Holocene (Severinghaus *et al.*, 1998; Grachev and Severinghaus, 2005).

In the NGRIP core, this climatic shift is most clearly marked by a change in deuterium excess values (Fig. 5(b): red curve) which occurs before or during the interval over which the changes described above are recorded (Steffensen *et al.*, 2008). Deuterium (D) and ^{18}O are important isotopic tracers of precipitation, and the relative deviation of the isotopic ratios $^2H/^1H$ and $^{18}O/^{16}O$ in per mille (‰) from those in Standard Mean Ocean Water (SMOW) are indicated by δD and $\delta^{18}O$, respectively (Johnsen *et al.*, 1989). The relationship between δD and $\delta^{18}O$ is given by: $\delta D = 8.0\ \delta^{18}O + d$, d being the so-called deuterium excess, which is approximately 10‰. The deuterium excess in precipitation varies seasonally and, in general, anti-correlates with δD and $\delta^{18}O$ on climatic timescales. Deuterium excess in the Greenland ice-core records indicates changes in the physical conditions at the oceanic origins of arctic precipitation and, in particular, can be considered a proxy for past sea surface temperature in the moisture source regions of the oceans (Johnsen *et al.*, 1989; Masson-Delmotte *et al.*, 2005; Jouzel *et al.*, 2007). The NGRIP deuterium excess record shows a 2–3‰ decrease at the Pleistocene–Holocene transition, corresponding to an ocean surface temperature decline of 2–4°C. This decrease is interpreted as a change in the source of Greenland precipitation from the warmer mid-Atlantic during glacial times to colder higher latitudes in the early Holocene (Johnsen *et al.*, 1989; Masson-Delmotte *et al.*, 2005). The change reflects a sudden reorganisation of the Northern Hemisphere atmospheric circulation related to the rapid northward movement of the oceanic polar front at the end of the Younger Dryas Stadial/Greenland Stadial 1 (Ruddiman and Glover, 1975; Ruddiman and McIntyre, 1981). Hence the deuterium excess record is an excellent indicator of the abrupt and major climatic shift at the Pleistocene–Holocene boundary. Synchronised records from the Greenland Dye-3, GRIP, and GISP2 ice cores also show changes in deuterium excess, and these are fully consistent with the shift in NGRIP deuterium excess values (Steffensen *et al.*, 2008).

In the NGRIP core, sampling at 5 cm intervals (annual resolution or greater) across the Pleistocene–Holocene transition enables the abrupt decline in deuterium excess to be pinpointed with great precision. The data (Fig. 5(b)) show that this change occurred in a period of 3 yr or less. Over several decades, the $\delta^{18}O$ changes from glacial to interglacial values (reflecting the temperature-dependent nature of the fractionation of oxygen isotopes), and there is an order of magnitude drop in dust concentration, reflecting a reduction in dust flux to

DOI: 10.1002/jqs

the ice sheet from arid or poorly vegetated regions. The chemical content of the ice also changes significantly in the decades after the shift in deuterium excess, lower Na^+ values, for example, suggesting reduced transfer of sea-salt particles to the ice and hence less stormy conditions in the North Atlantic, while annual precipitation, as reflected in annual ice-layer thickness, increases by a factor of two during the first one or two decades immediately after the observed change in deuterium excess. The latter reflects the fact that climatic warming is accompanied by an increase in precipitation (and hence thicker ice layers) and is a feature seen in all polar ice cores at similar climatic transitions.

The base of the Holocene can therefore be defined on the basis of the marked change in deuterium excess values that occurred over an interval of ca. 3 yr, and the stratigraphic boundary is further underpinned by shifts in several other key proxies that occurred over subsequent decades (Steffensen *et al.*, 2008). As such, the NGRIP ice core constitutes a stratotype for the base of the Holocene of unparalleled detail and chronological precision.

Type locality of the GSSP

Borehole NGRIP2, located in the central Greenland ice sheet at 75.10° N, 42.32° W. The Pleistocene–Holocene boundary is at 1492.45 m depth.

Geological setting and geographic location

The central Greenland ice sheet comprises materials from the atmosphere which are deposited onto the ice-sheet surface by precipitation or dry fall-out, the majority of which is accumulated snow. Surface melting rarely occurs, and even where this does take place there is no loss of material because refreezing occurs *in situ*. The ice begins as fresh snow (density 100–200 kg m^{-3}) that is then compacted in the top 100 m by pressure from the accumulating overburden. The snow is progressively transformed into ice (density 917 kg m^{-3}) and undergoes gradual but regular modification as a result of the slow flow of glacier ice from the central part of the ice sheet to the margins. This modification leads to lateral stretching and vertical thinning of the strata, but the stratigraphic integrity of the accumulating sequence is not compromised in any part of the NGRIP record. This integrity is demonstrated by the stratigraphic coherence of six full Holocene ice-core records from different sectors of the Greenland ice sheet (Johnsen *et al.*, 2001). A similar pattern is evident in Antarctica, where broad similarities are apparent between proxy climate records from both coastal and inland ice cores (Brook *et al.*, 2005).

Lithology

The lithology is almost entirely glacier ice. Impurities, such as dust, are at the parts per billion (ppb) level.

Accessibility

In order to examine the GSSP in the field, a deep ice-coring operation is necessary. Permission must be sought from the Greenland Home Rule through the Danish Polar Centre. However, the NGRIP cores are archived at the University of Copenhagen, and access to these can be gained through the NGRIP curator via the NGRIP Steering Committee. Hence, although the GSSP for the base of the Holocene is in a remote region and cannot be easily examined in the field, free access to the NGRIP core, in which the GSSP is clearly defined, can be assured at the University of Copenhagen. This means that the boundary stratotype for the base of the Holocene is as accessible as any other designated GSSP.

Conservation

See 'Accessibility', above.

Identification in the field

There are no visible features in a newly drilled core, but the line-scan analysis of a polished section of core (Fig. 4) shows the Pleistocene–Holocene transition clearly. The glacial ice (Pleistocene) has many more visible layers than Holocene ice, and the change from one type of ice to the other can be pinpointed with a precision of 20 cm or so. Electrical conductivity analysis (ECM, Fig. 5(b)) also shows a marked drop in conductivity, reflecting a reduction in atmospheric dust flux to the ice at the Pleistocene–Holocene boundary.

Stratigraphic completeness of the section

The strata in the ice sheet (and hence in the ice core) are complete from the surface downwards. The only loss of material results from drilling and core handling and this is minimal. Statistical studies of the variance of annual layer thickness suggest that the probability of annual layers with more than double, or less than half, the average thickness is less than 1% (Rasmussen *et al.*, 2006). It is therefore extremely unlikely that ice layers are absent from the sequence through missing precipitation.

Best estimate of age

The age of the base of the Holocene is derived from annual ice-layer counting across the Pleistocene–Holocene boundary and through ice from the entire Holocene. This counting involves the analysis of a range of physical and chemical parameters, many of which vary seasonally, thereby enabling annual ice layers to be determined with a high degree of precision. They include dust concentration, conductivity of ice and melted samples, $\delta^{18}O$ and δD, and a range of chemical species including Ca^{2+}, NH_4^+, NO_3^-, Na^+ and SO_4^{2-} (Rasmussen *et al.*, 2006; Fig. 5). In the upper levels of a number of the Greenland ice cores, annual ice layers can be readily identified on the basis of the $\delta^{18}O$ and δD records and seasonal variations in ice chemistry (Hammer *et al.*, 1986; Meese *et al.*, 1997). However, because of the relatively low accumulation rate at the NGRIP drill site, and a relatively high sensitivity of the annual cycles in $\delta^{18}O$ and δD to diffusion, NGRIP $\delta^{18}O$ and δD data are not suitable for the identification of annual ice layers, and there are no continuous chemistry measurements with sufficiently high

DOI: 10.1002/jqs

resolution for determination of annual layers in the section of the NGRIP core back to ~1405 m depth (10 227 yr b2 k). In order to obtain a complete Holocene chronology for NGRIP, therefore, it is necessary to link the early Holocene record with that from other Greenland core sites, Dye-3 and GRIP. The former is located in southeastern Greenland and is critical in terms of dating because the higher ice accumulation rate has produced the best resolved of all the Greenland ice-core timescales for the mid and late Holocene (Vinther *et al.*, 2006). The Pleistocene–Holocene boundary cannot be accurately defined or dated in that core, however, because of progressive thinning due to the flow of the ice. Near the lower limit at which annual layers can be resolved in the Dye-3 core, there is a significant decline in $\delta^{18}O$ values to below normal Holocene levels that persists for a few decades. This marks the '8.2 k yr event', a short-lived episode of colder climate that registers in a range of proxy climatic archives from around the North Atlantic region, and which most probably reflects the chilling of ocean surface waters following rapid meltwater release from ice-dammed lakes in northern Canada (Alley and Ágústsdóttir, 2005; Kleiven *et al.*, 2007). The 8.2 k yr event is also clearly recorded in the various proxy climate indicators in the NGRIP core (Thomas *et al.*, 2007). In Dye-3, NGRIP, GRIP and, indeed, in other Greenland ice-core sequences, the $\delta^{18}O$ reduction marking the 8.2 k yr event is also accompanied by a prominent ECM double peak and a marked increase in fluoride content. This fluoride represents the fall-out from a volcanic eruption, almost certainly on Iceland. The location of the double ECM peak inside the $\delta^{18}O$ minimum around 8200 yr BP constitutes a unique time-parallel marker horizon for correlating all Greenland ice-core records.

In the original Dye-3 core, the annual layer situated in the middle of the ECM double peak was dated at 8214 yr BP with an uncertainty of 150 yr (Hammer *et al.*, 1986). Subsequent high-resolution analysis of the Dye-3 stable isotopes has enabled this age estimate to be considerably refined and it is now dated to 8236 yr b2 k with a maximum counting error[1] of only 47 yr (Vinther *et al.*, 2006). Multi-parameter annual layer counting down from the 8236 yr double ECM horizon within the $\delta^{18}O$ minimum in the GRIP and NGRIP cores gives an age for the base of the Holocene, as determined by the shift in deuterium excess values, of 11 703 yr b2 k with a maximum counting error of 52 yr (Rasmussen *et al.*, 2006). The total maximum counting uncertainty (Dye-3 plus GRIP and NGRIP) associated with the age of the Pleistocene–Holocene boundary in NGRIP is therefore 99 yr, which is here interpreted as a 2σ uncertainty.[1] Indeed, the dating error of the deuterium excess transition in relation to the Saksunarvatn and Vedde Ash horizons is only one or two decades. In view of the 99 yr uncertainty, however, it seems appropriate to assign an age to the boundary of 11 700 yr b2 k. This figure is very close to that of 11 690 yr b2 k for the Pleistocene–Holocene boundary in the GISP2 core, where the counting error was estimated to be between 1% and 2% (Meese *et al.*, 1997). Moreover, the GICC05 and GISP2 timescales have approximately the same number of ice-core years in the 9500–11 500 yr b2 k sections of the cores and agree within a few years on the age of the Pleistocene–Holocene boundary, when the transition depth is defined using deuterium excess data (Rasmussen *et al.*, 2006). However, the more highly resolved stratigraphic record in the NGRIP ice core, and the multi-parameter annual layer counting approach that has been adopted, enable the climatic shift marking the onset of the Holocene to be located and dated with much greater precision. Accordingly, we recommend that the GSSP for the base of the Holocene be defined at a depth of 1492.45 m in the Greenland NGRIP ice core.

Global auxiliary stratotypes

In most cases, GSSPs are defined in the geological record on the basis of biological (fossil) evidence. The Holocene boundary as defined in this proposal is unusual, therefore, in that it is based on climatically driven physical and chemical parameters within the ice-core sequence. Accordingly, the Working Group has proposed a number of auxiliary stratotypes where the climatic signal that reflects the onset of the Holocene is identified on the more conventional basis of biostratigraphic evidence, or by integrating the GSSP age with biostratigraphic context. These proxy climate records can then be linked (correlated) directly with the NorthGRIP GSSP. Five auxiliary stratotypes have been initially proposed, and details of these are presented below. In due course, it is anticipated that additional global auxiliary stratotypes will be defined, *inter alia*, from Africa, from South America and perhaps from Antarctica.

Europe: Eifelmaar Lakes, Germany (Thomas Litt)

Lakes Holzmaar (HZM: 50° 7′ N, 6° 53′ E; 425 m above sea level (a.s.l.)) and Meerfelder Maar (MFM, 50° 6′ N, 6° 45′ E, 336.5 m a.s.l.) are located within the Westeifel Volcanic Field, less than 10 km apart. The present lake surface of MFM, which is the largest maar of the Westeifel Volcanic field, is 0.248 km^2, compared to 0.058 km^2 of HZM. The maximum water depth of both lakes is nearly the same (17–18 m).

Sediment successions of annual laminations in HZM and MFM form the basis for long varve chronologies based on microstratigraphical thin-section analyses. The chronology of HZM relates to varve counting of core B/C, which has been cross-checked by counting additional profiles (cores E/F/H; see Zolitschka, 1998). On the basis of comparison with calibrated accelerator mass spectrometry (AMS) ^{14}C dates (Hajdas *et al.*, 1995), the varve chronology has been corrected by adding 346 yr between 3500 and 4500 cal. yr BP. The HZM timescale extends to the present, whereas the MFM chronology is floating because varves have not been preserved continuously during the last ca. 1500 yr (core MFM 6). However, this unlaminated part of the MFM timescale is based on five dendro-calibrated AMS ^{14}C dates obtained from terrestrial macrofossils. The MFM chronology has been linked to the calendar-year timescale by accepting the Holzmaar age of 11 000 varve yr BP for the Ulmener Maar Tephra, an isochron of local importance (Zolitschka *et al.*, 1995) and present in both HZM and MFM. A comparison of the Holocene part of both chronologies demonstrates a good agreement, particularly for the Lateglacial–Holocene transition, which is clearly marked by distinct changes in deposition and in the vegetation (pollen) signal in both lakes, and dated to 11 600 varve yr BP (before AD 1950) in HZM (Zolitschka, 1998) and to 11 590 varve yr BP (before AD 1950) in MFM (Brauer *et al.*, 1999, 2001; Litt and Stebich, 1999; Litt *et al.*, 2001, 2003). Thin sections of cores HZM B/C and MFM 6 are available in the GeoForschungsZentrum Potsdam (GFZ), Germany.

Eastern North America: Splan Pond (= Basswood Road Lake), Canada (Les Cwynar)

Splan Pond (45° 15′ 20″ N, 67° 19′ 50″ W; 106 m a.s.l.; ~4 ha in area; 10.8 m maximum depth) lies in a basin underlain by Palaeozoic metasedimentary rocks. The lake is in the temperate zone and is within the Acadian Forest Region (Rowe, 1972), which is a mixed forest of various conifers and deciduous, broad-leaved trees. Sediment cores ranging from 580 to 644 cm in length have been analysed by a number of different investigators for pollen, plant macrofossils or organic content (Mott, 1975; Mott *et al.*, 1986; Levesque *et al.*, 1993; Mayle and Cwynar, 1995), diatoms (Rawlence, 1988), and chironomids, the latter providing a basis for palaeotemperature estimates (Walker *et al.*, 1991; Levesque *et al.*, 1997). All of the cores contain a thick (55–80 cm), grey clay layer that represents the Younger Dryas event. The clay ends abruptly at the Pleistocene–Holocene boundary with a switch from grey clay to dark-brown gyttja occurring within 5 cm, and loss-on-ignition values (organic carbon) rapidly increasing over that increment from <5% to >30%; the Pleistocene–Holocene boundary is therefore visually and lithologically striking. Correspondingly, the pollen record for this interval shows a sharp decline in herb pollen from 30% to 15% and a large increase in tree taxa, particularly *Picea*, which rises from 7% to 40% across this boundary. Diatom frustules are too rare in the Younger Dryas clay to yield usable counts, but increase abruptly in number at the Pleistocene–Holocene boundary (Rawlence, 1988). Chironomid analysis (Walker *et al.*, 1991; Levesque *et al.*, 1996) indicates abrupt changes in the composition of chironomid assemblages across the Pleistocene–Holocene boundary, with cold types that dominate the Younger Dryas clays, most notably *Heterotrissocladius*, giving way to warm types, particularly *Dicrotendipes*. A chironomid-based inference model provides an estimate of a 10°C increase in temperature across the Pleistocene–Holocene boundary (Levesque *et al.*, 1997). An AMS radiocarbon date of 10 090 ± 70 ^{14}C yr BP was obtained on bract, seed, twig and leaf fragments. This provides a 2σ age range (at 97.5% probability) of 11 385–11 981 cal. yr BP (Reimer *et al.*, 2004) for the Younger Dryas–Holocene boundary.

Little remains of the various cores that have been previously studied. Although on private property, the lake is directly accessible by car, and the owners have been readily obliging in allowing access.

East Asia: Lake Suigestu, Japan (Takeshi Nakagawa)

Lake Suigetsu (35° 35′ 08″ N, 135° 52′ 57″ E, 0 m a.s.l.; 2 km in both N–S and E–W diameters; 34 m deep) is a tectonic lake located on the Sea of Japan coast of central Japan. Although the site lies in the warm temperate zone, it is also very close to the boundary with the cool temperate forest zone, making the vegetation around the site an ecotonal boundary that is very sensitive to climate change. The lake contains 73.5 m of thick lacustrine sediment, and the top 40 m (0 to ca. 50 000 yr BP) of the sediment sequence is annually laminated. Two sets of cores were recovered from near the centre of the lake in 1993 and 2006 (SG93 and SG06 cores, respectively). The biostratigraphically defined Pleistocene–Holocene boundary occurs at 13.51 m below the SG93 core top (and will be replicated in core SG06 by the end of 2009), where the temperature curve, quantitatively reconstructed from pollen data, shows an abrupt rise (Nakagawa *et al.*, 2003, 2005, 2006). This horizon is characterised by an abrupt fall in percentages of *Fagus* pollen (from >40% to ~20%), and it occurs 124 cm below the base of the widespread U-Oki tephra (this stratigraphical relationship will be refined in core SG06 which, unlike SG93, does not have any gaps in the sediment sequence). The SG93 core has been intensively dated by AMS radiocarbon dating of terrestrial plant macro-remains (Kitagawa and van der Plicht, 1998a,b, 2000). Eighteen AMS radiocarbon ages are available from a 1000 varve yr long section spanning the Pleistocene–Holocene boundary. Wiggle matching of this dataset to the IntCal04 tree ring chronology using a Bayesian approach (Bronk Ramsey *et al.*, 2001) dates the Pleistocene–Holocene boundary at Lake Suigetsu to 11 552 ± 88 cal. yr BP (at 2σ).

As at Splan Pond, little now remains of the original core material from SG93, but a complete set of SG06 cores (consisting of cores from four bore holes) is archived at the University of Newcastle Upon Tyne (UK), and can be accessed for research purposes on request. Should new cores be required, permission to core the lake is relatively easy to obtain, and local coring companies have the expertise to do the fieldwork. Access to the lake is very good. A few tourist boats capable of carrying about 100 passengers operate regularly on the lake all year round, and coaches can reach the port on the lake shore without difficulty. It is about five hours drive from the nearest international airport (Osaka).

Australasia: Lake Maratoto, New Zealand (Rewi Newnham, David J. Lowe and Peter Kershaw)

Lake Maratoto (37° 53′ S, 175° 18′ E; 52 m a.s.l.) is one of more than 30 small lakes in the Waikato lowlands formed by aggradation of the ancestral Waikato River around 20 000 calendar years ago in northern North Island, New Zealand (Green and Lowe, 1985; Lowe and Green, 1992; Selby and Lowe, 1992). It preserves a continuous sedimentary record from the present back to the Last Glacial Maximum (or soon thereafter) and, in addition to well-developed pollen and tephrochronological records (Green and Lowe, 1985; Lowe, 1988; McGlone, 2001), chironomid, Cladocera, sedimentary pigments, geochemical and stable isotope (δ^{13}C) data have also been obtained from the sequence (Green, 1979; Boubée, 1983; Etheridge, 1983; McCabe, 1983). An extensive multi-core (33 lake cores) stratigraphic survey, including use of ground radar, has enabled reconstruction of the lake's origin and development in detail (Lowe *et al.*, 1980; Lowe, 1985; Green and Lowe, 1985).

In assigning an auxiliary stratotype from Australasia and other parts of the Southern Hemisphere, it is important to recognise that Lateglacial (Pleistocene–Holocene transition) records from southern mid–high latitudes in particular appear to be broadly anti-phased with climate trends from the North Atlantic region. As a consequence, they typically show a progressive warming often commencing before 11 700 yr b2 k, rather than an abrupt warming step at that time (Alloway *et al.*, 2007). The position of the Pleistocene–Holocene boundary in Lake Maratoto can be pinpointed by using tephrostratigraphy in combination with palynostratigraphy. Full Holocene warmth, as reflected in the pollen sequence (McGlone, 2001; Wilmshurst *et al.*, 2007), is attained at close to the time of deposition of the andesitic Konini Tephra, derived from the Egmont/Taranaki volcano and recently dated at 11 720 ± 220 cal. yr BP (Hajdas *et al.*, 2006; Lowe DJ *et al.*, 2008). In a 3 m long core taken from the northern part of Lake Maratoto (April 1979), and referred to variously as core Mo A/1 (Green, 1979; Lowe *et al.*, 1980), core 4,1a (Green and Lowe, 1985), and core X79/1 (McGlone, 2001), the Konini Tephra (Eg-11) is preserved as a pale-grey fine ash layer 2–3 mm in thickness at a depth of 1.50 m below the Taupo Tephra (Lowe, 1988). Its stratigraphic position is constrained by two easily recognised tephra marker beds: a distinctive greyish-black coarse ash, the Mangamate Tephra, lies above it at 1.40–1.45 m depth, whereas the white and cream, fine and medium-bedded ash of Waiohau Tephra lies below it at 1.67–1.70 m depth below the Taupo Tephra. Eg-11 has been provenanced and correlated with the Konini Tephra (as defined by Alloway *et al.*,

1995) through its ferromagnesian mineralogical assemblage and by electron microprobe analysis of glass shards (Lowe, 1988; Lowe DJ *et al.*, 1999, 2008). Thirty-four ^{14}C dates in total have been obtained from the Lake Maratoto sequence. The peaty sediments containing Eg-11 were dated at 10 100 $\pm$ 100 ^{14}C yr BP (Wk-519; Hogg *et al.*, 1987; Lowe, 1988), which gives a 2σ calibrated age (at 99.6% probability) of 12 049–11 305 cal. yr BP (Reimer *et al.*, 2004). The attainment of full climatic warmth associated with the onset of the Holocene at around the time of deposition of Konini Tephra has also been reported from other pollen sites across the North Island (Newnham *et al.*, 1989; Newnham and Lowe, 2000; Turney *et al.*, 2003; Alloway *et al.*, 2007).

Lake Maratoto is located 12.5 km south of Hamilton city centre, 3 km south-west of Hamilton International Airport and 1.5 km due west of State Highway 3. The lake, though on private land owned by two adjacent landowners, is protected from any development in perpetuity by a covenant under the New Zealand Government's Queen Elizabeth II National Trust Act of 1977. There is easy access by vehicle to the northern shore of the lake on a well-surfaced farm road.

Deep Oceans: the Cariaco Basin, Venezuela (Konrad Hughen)

The Cariaco Basin (10° 41′ N, 65° W) is an anoxic marine basin off the northern coast of Venezuela. Restricted deep circulation and high surface productivity in the basin create an anoxic water column below 300 m. The climatic cycle of a dry, windy season with coastal upwelling, followed by a non-windy, rainy season, results in distinctly laminated sediment couplets. It has been demonstrated previously that the laminae couplets are annually deposited varves and that light laminae thickness and sediment reflectance (grey scale) are sensitive proxies for surface productivity, upwelling and trade wind strength (Hughen *et al.*, 1996). Nearly identical patterns, timing and duration of abrupt changes in Cariaco Basin climate proxies (including laminae thickness (upwelling), sediment reflectance (productivity), bulk sediment elemental abundances (run-off), foraminiferal elemental abundances (SST) and molecular isotopes (vegetation)) compared with Greenland ice-core records at 10 yr resolution during the last deglaciation (Hughen *et al.*, 1996, 2000, 2004a; Haug *et al.*, 2001; Lea *et al.*, 2003) provide evidence that rapid climate shifts in the two regions were synchronous. A likely mechanism for this linkage is the response of North Atlantic trade winds to the equator–pole temperature gradient forced by changes in high-latitude North Atlantic temperature. The highest-resolution record (sediment reflectance) measured on the highest-deposition-rate Cariaco sediment core (PL07-58PC) showed the Pleistocene–Holocene transition occurring over approximately 6 varve yr. Core PL07-58PC has been intensively dated by AMS radiocarbon dating of planktonic foraminifera (Hughen *et al.*, 2000). 197 AMS radiocarbon ages are available from a 1915 varve yr long section spanning the Pleistocene–Holocene boundary. Wiggle matching of these dates to the IntCal04 tree ring chronology shows strong agreement ($r = 0.99$) and indicates that the age of the Pleistocene–Holocene boundary mid-point in the Cariaco region is 11 578 $\pm$ 32 cal. yr BP (2σ).

In order to examine the Cariaco Basin Holocene boundary in the field, a deep-ocean sediment coring operation is necessary. Permission must be sought from the Venezuelan government in order for foreign ships to enter the area. However, Cariaco Basin sediment cores are archived at the core repository at Lamont-Doherty Earth Observatory of Columbia University, NY, and access to these can be gained through the LDEO core repository curator. In addition, a section of Cariaco sediment core containing the Pleistocene–Holocene boundary is on permanent display at the Museum of Natural History in New York City, USA.

Note [1]The uncertainty estimate of the GICC05 timescale is derived from the number of potential annual layers that the investigators found difficult to interpret. These layers were counted as ½ ± ½ yr, and the so-called maximum counting error (mce) is defined as one half times the number of these features. At the base of the Holocene, the mce is 99 yr. Strictly speaking, the value of the mce cannot be interpreted as a standard Gaussian uncertainty estimate, but it is estimated that the true age of the base of the Holocene is within 99 yr of 11 703 yr b2 k with more than 95% probability. For further discussion see Andersen *et al.* (2006).

Acknowledgements We would like to express our thanks to Jim Ogg (Secretary of the ISC), Thijs von Kolfschoten (Secretary of the SQS) and other members of the Subcommission on Quaternary Stratigraphy for their advice, support and encouragement during the preparation of the proposal, and its subsequent passage through the various stages of the ratification process. We also thank Dr Mebus Geyh for his helpful comments on the manuscript.

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