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A global synthesis of the marine and terrestrial evidence for glaciation during the Pliocene Epoch

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ABSTRACT

The Pliocene climate is globally warm and characterised by high atmospheric carbon dioxide concentrations, yet the terrestrial and marine scientific communities have gathered considerable evidence for substantial glaciation events in both the Northern and Southern Hemisphere prior to the Quaternary. Evidence on land is fragmentary, but marine records of glaciation present a more complete history of Pliocene glaciation. Here we present a global compilation of the terrestrial and marine glacial evidence for the Pliocene and demonstrate four glaciation events that can be identified in the Southern and/or Northern Hemisphere prior to the latest Pliocene intensification of Northern Hemisphere glaciation. There are two globally recognisable glacial events in the early Pliocene (*c.* 4.9–4.8 Ma and *c.* 4.0 Ma), one event around the early/late Pliocene transition (*c.* 3.6 Ma), and one event during Marine Isotope Stage M2 (*c.* 3.3 Ma). Long-term climate cooling, decreasing carbon dioxide concentrations in the atmosphere and high climate sensitivity in the Pliocene probably facilitated each glaciation event, however the mechanisms behind the early Pliocene glacial events are unclear. The global glaciation at *c.* 3.3 Ma may be caused by changes in ocean gateways, whereas the decline in carbon dioxide concentrations is important for the latest Pliocene intensification of Northern Hemisphere glaciation.

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1. Introduction

The Pliocene Series or Epoch (5.33–2.58 Ma) is divided into two stages, the Zanclean (or early Pliocene Subepoch) with a base at 5.33 Ma (van Couvering et al., 2004), and the Piacenzian (or late Pliocene Subepoch), which begins at 3.60 Ma (Castradori et al., 1998). Since 2009, the end of the Pliocene has been defined by the base of the Pleistocene Series (Quaternary System) and the Gelasian Stage, which is dated at 2.58 Ma. This replaced the previous base of the Pleistocene, which was placed at 1.81 Ma, and caused the transfer of the Gelasian Stage (2.58–1.81 Ma) from the Pliocene into the early Pleistocene Subepoch (Gibbard et al., 2010).

The Pliocene Epoch spans a critical period in Earth's history during which global climate underwent a profound transition from relatively warm climates to the substantially cooler climates of the Pleistocene. The early Pliocene is considered to have had a globally warm climate (Ravelo et al., 2007; Brierley et al., 2009; Larivière et al., 2012), but is considerably less investigated than the mid-Piacenzian Warm Period (previously mid-Pliocene Warm Period, 3.29–2.97 Ma), which has been the focus of the Pliocene Research, Interpretation and Synoptic Mapping (PRISM) initiative (e.g. Dowsett et al., 2010) and the Pliocene Model Intercomparison Project (PlioMIP) (e.g. Haywood et al., 2013). Suggestions that the mid-Piacenzian Warm Period could serve as a true direct analogue for a globally warmed future have been questioned (Sarnthein et al., 2009; Haywood et al., 2011), but it nevertheless provides an ideal time interval to understand the climatic processes of a warm, high carbon dioxide world. In particular, the similarity of the late Pliocene palaeogeography to that of today, the occurrence of fossil assemblages similar to modern assemblages, and the well-preserved terrestrial and marine geological records, mean that, although problems remain, a relatively good insight into the ocean conditions and biosphere during a warm global late Pliocene climate can be assembled (Dowsett et al., 2012; Salzmann et al., 2013). Even within this globally warm Pliocene world, short-lived, episodic glaciation events and accompanying sea-level fluctuations have been recorded in benthic isotope records as well as sequence boundaries before the end-Pliocene climate deterioration. Such glacial events and sea-level fluctuations are rare in the Zanclean, but become progressively more common since 3.6 Ma (Lisiecki and Raymo, 2005; Miller et al., 2005, 2012). The global climate deteriorated severely only in the latest Pliocene, which led to the onset of widespread Northern Hemisphere glaciation at *c.* 2.75 Ma.

In this review, we compile both the terrestrial and marine evidence of glaciation during the entire Pliocene Epoch and present a global comparison of Arctic and Antarctic Pliocene (episodic) glaciation. Firstly, we aim to bridge the gap between the terrestrial and marine communities investigating Pliocene climate. Although there is ample evidence of episodic Pliocene glaciation in the marine and terrestrial realm, they have never been compared directly to identify synchrony or diachrony between the terrestrial and marine records. Secondly, by comparing all Pliocene records of glaciation on a global scale, we aim to identify the timing of Arctic and Antarctic ice sheet expansion. We also provide potential links between the glaciation events to large-scale, long- and short-term changes in oceanography and global climate. Finally, by comparing the Pliocene record with that from the Pleistocene, we attempt to reconstruct the World immediately before the impact of major Northern Hemisphere glaciation and present an assessment of how the Northern Hemisphere glaciations were initiated.

2. Pliocene global climate—characteristics

The chronology of the Pliocene Epoch has been finely tuned by orbital tuning or astrostratigraphy based upon marine oxygen isotope sequences (Shackleton et al., 1990; Hilgen, 1991; Tiedemann et al., 1994; Shackleton and Crowhurst, 1995; Shackleton et al., 1995). These approaches were further improved in the most recent Neogene timescales, where could be relied on more accurate numerical astronomical solutions, high-resolution studies of uplifted land-sections and more complete ocean drill records (Lourens et al., 2005; Hilgen et al., 2012). A highly detailed global benthic foraminifer oxygen isotope stack, of comparable resolution to those for the Pleistocene, has also been presented for the Pliocene demonstrating that the climate system during the period was controlled mainly by the *c.* 40 ka obliquity periodicity (Lisiecki and Raymo, 2005). Because the global stack, supported by a palaeomagnetic reversal chronology, provides a global standard for detailed chronostratigraphical division, key events in the Pliocene global climate history can be identified.

The evidence for climate oscillations through the Pliocene is seen in geological sequences throughout the world. They include ice-volume records from oxygen isotope successions, the occurrence of glaciomarine sediments or IRD (=-ice rafted debris/detritus), the variation of a range of geochemical or palynological proxies in marine and terrestrial sediment sequences (e.g. Willis et al., 1999; Knies et al., 2009;

Melles et al., 2012), and sea-level change in continental shelf marine sediments (Dowsett and Cronin, 1990; Miller et al., 2012).

There is limited variation in the early Pliocene ice-volume records (Lisiecki and Raymo, 2005), and both data and model studies indicate expanded tropical warmth, reduced meridional and zonal sea-surface temperature gradients, mean global temperatures 3–4 °C warmer than today and the suppression of Northern Hemisphere glaciation (Hill et al., 2007; Lunt et al., 2009; Brierley and Fedorov, 2010; Salzmann et al., 2011b; Larivière et al., 2012). In the Southern Hemisphere, the East and West Antarctic Ice Sheets were already fully established, but showed a very dynamic behaviour during Pliocene warm periods (e.g. Naish et al., 2009; Dolan et al., 2011).

Northern Hemisphere glaciation was a gradual process that began around 3.6 Ma, at the beginning of the late Pliocene (Mudelsee and Raymo, 2005). This is reflected in the gradual North Atlantic sea-surface temperature decline since *c*. 3.6 Ma, which culminates in the glacial Marine Isotope Stage (MIS) M2 at *c*. 3.3 Ma (Lawrence et al., 2009; Naafs et al., 2010). There is early evidence for an ice sheet in the interior of North America at *c*. 3.5 Ma in the James Bay Lowland, Canada (Gao et al., 2012), but glaciation in the Northern Hemisphere was significantly reduced compared to the present, and probably also occurred occasionally in Greenland (Hill et al., 2007), Alaska (Lagoe and Zellers, 1996) and the Canadian Arctic Archipelago (de Vernal and Mudie, 1989). In general, the Pliocene Arctic remained warmer than today and the tree line was established considerably to the north (Salzmann et al., 2011b). For example, there is evidence for evergreen forests on Meighen Island, Ellesmere Island and the north-eastern Russian Arctic (Matthews and Ovenden, 1990; Tedford and Harington, 2003; Andreev et al., 2013). Multi-proxy evidence from the recently re-dated Beaver Pond peat on Ellesmere Island suggests that around the early/late Pliocene transition (i.e. the Zanclean/Piacenzian stage boundary) mean annual temperatures in this region were 18.3 ± 4.1 °C warmer than present (Ballantyne et al., 2010; Csank et al., 2011a,b; Rybczynski et al., 2013). Following the mid-Piacenzian Warm Period, when Earth's climate was 2–3 °C warmer than today (3.3–3.0 Ma), global climate again cooled further finally leading to the major glaciations at the transition from the Pliocene and to the early Pleistocene (MIS G6–96) (Funder et al., 2001; Bartoli et al., 2006; Lawrence et al., 2010).

The gradual process of ice build-up during the Pliocene suggests one or more tectonic mechanisms that operate over long time periods. One possibility is that uplift of the Tibetan Plateau led to weathering-induced atmospheric carbon dioxide removal (Raymo et al., 1988; Mudelsee and Raymo, 2005). Alternatively or additionally, tectonic processes at ocean gateways during the Pliocene might have led to the establishment of the modern ocean circulation: the Indonesian Seaway shoaled between 4–3 Ma (Cane and Molnar, 2001), the Central American Seaway closed (Haug and Tiedemann, 1998; Steph et al., 2006) and Pacific water flowed into the Arctic via the Bering Strait around 4.5 Ma (Marincovich and Gladenkov, 1999; Verhoeven et al., 2011). Recent modelling experiments indicate that the glaciation of Greenland during the late Pliocene is mainly controlled by a decrease in atmospheric carbon dioxide (Lunt et al., 2008a).

3. Nature of the evidence of glaciation and limitations

The evidence of ancient glaciations is generally preserved in two ways: on land, glacial diamicton (till, tillite) and related meltwater deposits occur in sedimentary sequences, whilst beneath the seas the main evidence occurs as waterlain diamictons or, more frequently and far-wider distributed, as IRD carried by floating ice. In volcanically active regions, such as the Antarctic Peninsula or Iceland, the study of glaciovolcanic sequences provides another powerful tool to reconstruct thickness, extension and thermal regime of past ice sheets (Smellie et al., 2008, 2009; Geirsdottír, 2011). Fig. 1 shows the localities where evidence of Pliocene glaciations is recorded.

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

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

The deposits arising from both sea-ice and iceberg melting are termed ice-rafted detritus or debris (IRD). Icebergs calved from glaciers that terminate in the sea (or lakes), especially those from temperate ice tongues, frequently include debris that is frozen onto or within the ice body. As the ice drifts with the ocean currents and wind, its included sediment can be transported over long distances that can exceed 1000–1500 km (Ruddiman, 1977; Bond et al., 1992; T. Williams et al., 2010) within and across ocean basins. According to Dowdeswell and Murray (1990), the volume of sediment deposited by iceberg rafting is dependent on several factors that are determined by the primary sediment source, the rate of glacier calving, the rate of iceberg melt, the ocean surface-water temperature and factors controlling the iceberg drift rate (cf. Benn and Evans, 2010). Icebergs can be produced in different ways, which T. Williams et al. (2010, p. 358–359) summarise as follows: "*IRD-bearing icebergs can be produced when an ice sheet reaches the edge of the continental shelf during glacial maxima* (Marshall and Koutnik, 2006), *during glacial advance* (McManus et al., 1999), *during glacial retreat* (McManus et al., 1999), *or due to instabilities in the ice sheet leading to ice rafting like the Heinrich events of the North Atlantic in the last glacial* (Hemming, 2004)." Because we review data of glaciation over a time scale of almost 3 million years (Pliocene, 5.33–2.58 Ma), we did not distinguish between these different production mechanisms, which operate on shorter time scales. We have simply taken the IRD records as evidence for an ice sheet that reaches the coastline.

In subpolar regions, the occurrence of coarse-grained (>63 μm) sediment and clasts in open ocean sediments is generally attributed to iceberg transport and melting, and provides indirect evidence of terrestrial glaciation (Andrews, 2000; St John, 2008; Polyak et al., 2010). In polar regions, melting of sea ice usually contributes fine grained sediment, but occasionally also coarse-grained sediment, to the geological record (Nürnberg et al., 1994; Müller and Knies, 2013). Fine-grained

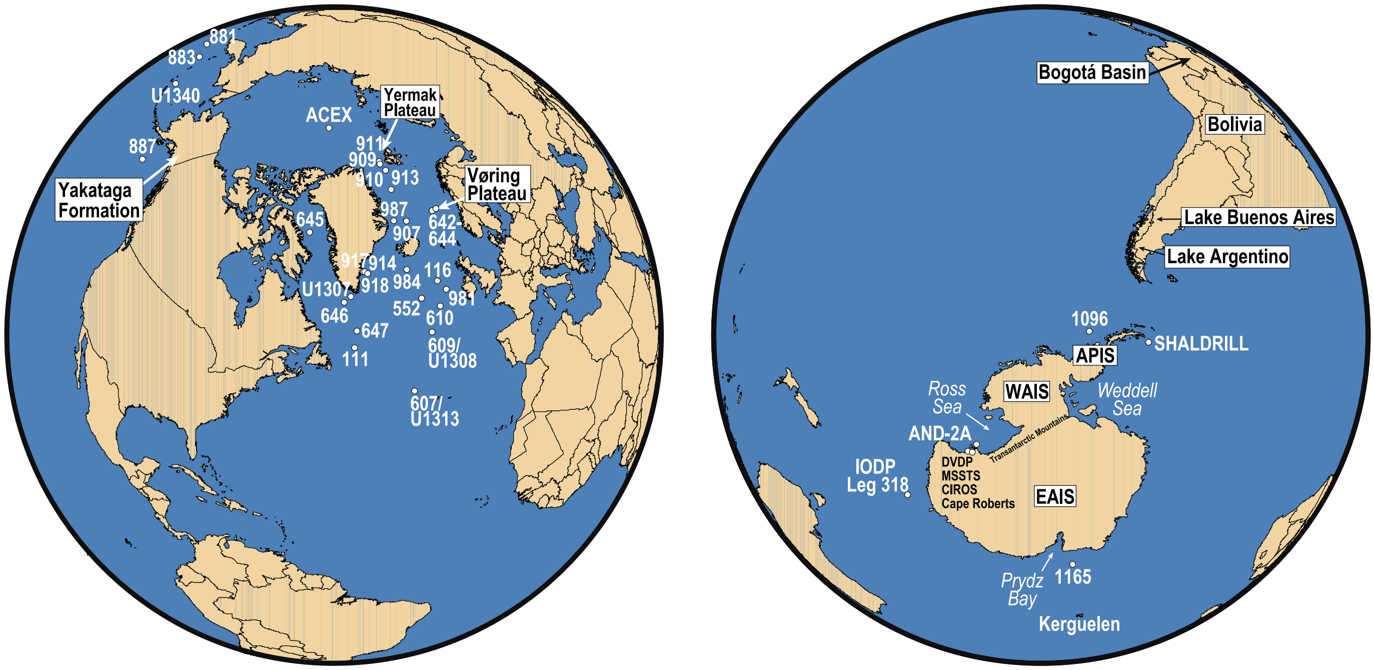


Fig. 1. Map of all localities with evidence for glaciation during the Pliocene Epoch. Ocean drill sites are from the Deep Sea Drilling Project (DSDP), Ocean Drilling Program (ODP) and Integrated Ocean Drilling Program (IODP).

sediment that occurs in suspension on shallow shelves can be incorporated in sea ice during ice-freeze-up (frazil-ice formation) (Kempema et al., 1989; Nürnberg et al., 1994; Polyak et al., 2010). In addition, anchor ice, which forms on the sea-floor during super-cooled conditions, can incorporate and distribute any sediment (Reimnitz et al., 1987; Polyak et al., 2010). The sediment transported by sea ice cannot be seen as direct evidence of glaciation on land, and makes the distinction between sea ice- and iceberg-transport processes challenging yet essential for palaeoenvironmental interpretations. Based on grain size alone, it is often not possible to attribute a small coarse-grained fraction in deep-sea sediments to either sea ice or iceberg transport in the geological record (St John, 2008). Using surface textures of sand-sized quartz grains, St John (2008) demonstrated the presence of sea ice and iceberg derived sediment in the Arctic Ocean since the Eocene. Chemical fingerprinting of iron oxide sand grains illustrated periods with perennial sea ice possibly already in the middle Eocene (*c.* 47.5 Ma; Darby, 2014) and in the late Miocene (Darby, 2008). The Pliocene IRD records reviewed below are mainly from subpolar regions (Fig. 1), and should therefore mostly present iceberg transport.

4. Northern Hemisphere

Early evidence of Northern Hemisphere glaciation—as isolated glaciers or small ice caps that reached the coastline on Greenland or Svalbard—is presented in glacigenic microstructures on Palaeocene quartz sand grains (Immonen, 2013), and in IRD records from the middle Eocene (45 Ma) of the Arctic Ocean (Moran et al., 2006) and from the middle Eocene (44 Ma) to early Oligocene (38–30 Ma) of the Greenland Sea (Eldrett et al., 2007; Tripati et al., 2008). It nevertheless takes until the middle to late Miocene and Pliocene that small-scale glaciations in the Arctic region occur more frequently (e.g. Knies and Gaina, 2008), and the late Pliocene marks the transition from local to extensive regional scale glaciations in the Northern Hemisphere (Fronval and Jansen, 1996; Kleiven et al., 2002; Matthiessen et al., 2009a; Bailey et al., 2013). All Northern Hemisphere records discussed below are summarised and displayed together with reconstructions of the major ice sheets in Fig. 2.

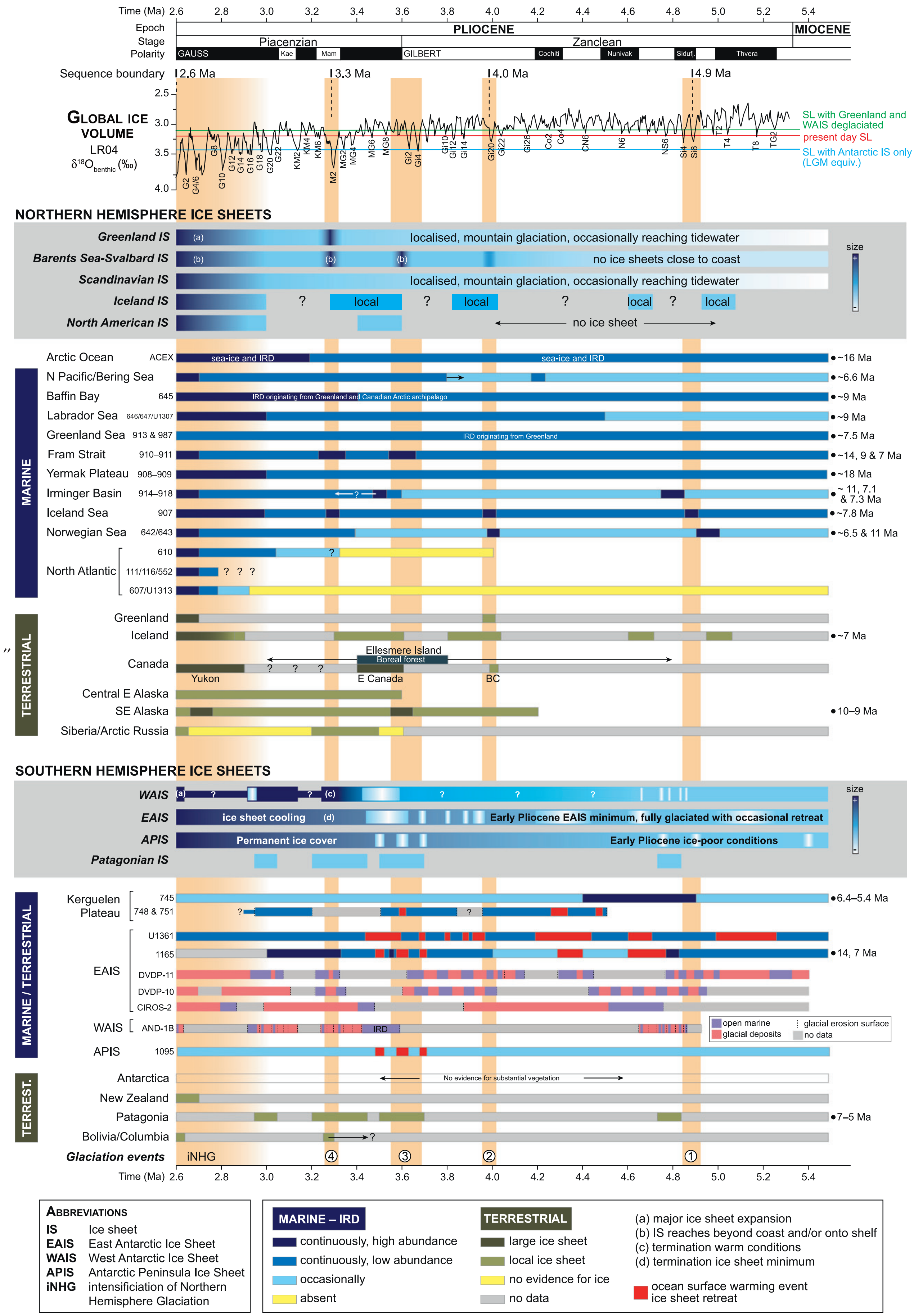
4.1. Direct evidence from the continents

4.1.1. North-western Canada and Alaska

In northern Canada and Alaska, the oldest till and accompanying IRD in adjacent marine settings dates from the early Miocene, with regionally widespread glaciation occurring in the Pliocene and regularly throughout the Pleistocene (cf. Barendregt and Duk-Rodkin, 2011).

In south-eastern Alaska the oldest evidence of glaciation is found in the Yakataga Formation, a sequence of glacially-influenced marine sediments. These sediments record IRD in the Gulf of Alaska. Originally dated to the late Miocene, from 5.91 to 5.50 Ma, by Lagoe et al. (1993), palaeomagnetic evidence indicates the onset of glaciation here to 4.2–2.27 Ma (Krissek, 1995; Rea and Snoeckx, 1995). According to Duk-Rodkin et al. (2004), mountain glaciers transported debris in the Alaska coastal ranges during this time. Terrestrial evidence from south-eastern Alaska indicates evidence for several glacial advances, where diamictites and associated deltaic deposits are interbedded with lava flows. The potassium–argon dates of the interbedded lavaflows indicate that the early tillites occur in the middle Miocene (*c.* 10–9 Ma), one tillite is about 3.6 Ma, and two are between 8.8 and 2.7 Ma (Denton and Armstrong, 1969). Although there is some disagreement over the genesis of the oldest materials, there is general acceptance that the younger units are indeed of glacial origin (cf. Duk-Rodkin et al., 2004). Both tidewater and the earliest Cordilleran glaciation are represented, but it appears that the glaciers were local and did not extend into the interior. Duk-Rodkin et al. (2004) note

Fig. 2. Terrestrial and marine evidence for glaciation during the Pliocene Epoch, combined into a summary of the major Northern and Southern Hemisphere ice sheets (grey insets). Four distinct glaciation events (① to ④) in the Northern and Southern Hemisphere have been identified, next to the latest Pliocene intensification of Northern Hemisphere Glaciation (iNHG). Sequence boundaries are from Miller et al. (2005), time scale and global ice volume is the LR04 global stack of Lisiecki and Raymo (2005) and estimated sea level (SL) is from Naish and Wilson (2009). Timing of glaciation events in the Miocene are also indicated.



Time (Ma) 2.6 2.8 3.0 3.2 3.4 3.6 3.8 4.0 4.2 4.4 4.6 4.8 5.0 5.2 5.4
Epoch PLIOCENE MIOCENE
Stage Piacenzian Zanclean
Polarity GAUSS Kae Mam GILBERT Cochiti Nunivak Sidufj. Thvera
Sequence boundary 2.6 Ma 3.3 Ma 4.0 Ma 4.9 Ma
GLOBAL ICE VOLUME
LR04
δ18O benthic (‰)
SL with Greenland and WAIS deglaciated
present day SL
SL with Antarctic IS only (LGM equiv.)
NORTHERN HEMISPHERE ICE SHEETS
Greenland IS
Barents Sea-Svalbard IS
Scandinavian IS
Iceland IS
North American IS
localised, mountain glaciation, occasionally reaching tidewater
no ice sheets close to coast
local
no ice sheet
size
MARINE
Arctic Ocean ACEX
N Pacific/Bering Sea
Baffin Bay 645
Labrador Sea 646/647/U1307
Greenland Sea 913 & 987
Fram Strait 910–911
Yermak Plateau 908–909
Irminger Basin 914–918
Iceland Sea 907
Norwegian Sea 642/643
North Atlantic 610 111/116/552 607/U1313
sea-ice and IRD
IRD originating from Greenland and Canadian Arctic archipelago
IRD originating from Greenland
~16 Ma
~6.6 Ma
~9 Ma
~9 Ma
~7.5 Ma
~14, 9 & 7 Ma
~18 Ma
~ 11, 7.1 & 7.3 Ma
~7.8 Ma
~6.5 & 11 Ma
TERRESTRIAL
Greenland
Iceland
Canada
Central E Alaska
SE Alaska
Siberia/Arctic Russia
Ellesmere Island
Boreal forest
Yukon
E Canada
BC
~7 Ma
10–9 Ma
SOUTHERN HEMISPHERE ICE SHEETS
WAIS
EAIS
APIS
Patagonian IS
ice sheet cooling
Early Pliocene EAIS minimum, fully glaciated with occasional retreat
Permanent ice cover
Early Pliocene ice-poor conditions
MARINE / TERRESTRIAL
Kerguelen Plateau 745 748 & 751
EAIS U1361 1165 DVDP-11 DVDP-10 CIROS-2
WAIS AND-1B
APIS 1095
IRD
6.4–5.4 Ma
14, 7 Ma
open marine
glacial deposits
glacial erosion surface
no data
TERREST.
Antarctica
New Zealand
Patagonia
Bolivia/Columbia
No evidence for substantial vegetation
7–5 Ma
Glaciation events iNHG 4 3 2 1
Time (Ma) 2.6 2.8 3.0 3.2 3.4 3.6 3.8 4.0 4.2 4.4 4.6 4.8 5.0 5.2 5.4
ABBREVIATIONS
IS Ice sheet
EAIS East Antarctic Ice Sheet
WAIS West Antarctic Ice Sheet
APIS Antarctic Peninsula Ice Sheet
iNHG intensificiation of Northern Hemisphere Glaciation
MARINE – IRD
continuously, high abundance
continuously, low abundance
occasionally
absent
TERRESTRIAL
large ice sheet
local ice sheet
no evidence for ice
no data
(a) major ice sheet expansion
(b) IS reaches beyond coast and/or onto shelf
(c) termination warm conditions
(d) termination ice sheet minimum
ocean surface warming event
ice sheet retreat

that the glaciation at 2.7 Ma appears to be coeval with the first Cordilleran glaciation of the interior Yukon that dates to 2.9–2.6 Ma during which a substantial ice cap developed. This event saw the first development of an ice cap, which covered much of the central Yukon region (Horton Ice Cap). Denton and Armstrong (1969) concluded that cold-climate conditions first occurred in the region at *c.* 10 Ma, and that following regional uplift at *c.* 4 Ma, cold climates returned towards the end of the Pliocene.

In central East Alaska, at least one glaciation is represented by Gauss-age (i.e. late Pliocene) normally magnetised diamicton, a similar sequence occurring in the west-central Alaskan Tintina Trench where the interaction of Cordilleran and local ice are recorded. A similar late Pliocene age is determined for a glaciation of the Smoking Hills region, the outwash from which entered the Mackenzie Delta area (Duk-Rodkin et al., 2004). Glacial deposits, possibly as old as 4 Ma, occur interbedded with volcanic rocks near Mount Edziza in adjacent northern British Columbia (Fulton et al., 2004).

4.1.2. Eastern Canada and United States

A till was deposited around 3.6–3.4 Ma by an early ice sheet in the James Bay Lowland, Canada, indicating an early glaciation of a magnitude comparable to Pleistocene glaciations (Gao et al., 2012). Other direct evidence of continental glaciation in North America is found in the Atlanta Till, which was deposited around 2.41 Ma (early Pleistocene) during an early, southerly expansion of the Laurentide Ice Sheet (Balco et al., 2005; Rovey and Balco, 2011).

4.1.3. Greenland

Japsen et al. (2006) suggest that mountain glaciers may have been present on Greenland at *c.* 4 Ma resulting from uplift in East and West Greenland. The late Miocene to Pliocene mountain formation in Greenland and uplift along the northwest European continental margin in the early Pliocene may have been essential for the initiation of large Northern Hemisphere ice sheets (Japsen et al., 2006; Solgaard et al., 2013; Knies et al., 2014). Nevertheless, ice sheet model studies demonstrate a large reduction of the Greenland Ice Sheet compared to the present during the mid-Piacenzian (Hill et al., 2010; Dolan et al., 2011; Koenig et al., 2011; Solgaard et al., 2013), but there are no direct accounts on Greenland of the presence of a Pliocene ice cap. Pliocene continental deposits are not available, and ice cores only extend back to *c.* 125,000 years ago (NEEM community members, 2013). The fauna and flora recovered from the late Pliocene Île de France Formation in north-east Greenland suggests seawater temperatures too high for perennial sea ice and air temperatures considerably higher than present (Bennike et al., 2002). The Kap København Formation in North Greenland consist mainly of shallow marine near-shore sediments which record (sub)arctic and boreal conditions (Funder et al., 2001). But with an estimated age of *c.* 2.4 Ma, this formation is currently placed in the earliest Pleistocene (Knies et al., 2009; Gibbard et al., 2010). The formation rests on and is overlain by till, which is one of the few direct lines of evidence for glaciation on Greenland around the intensification of Quaternary-style Northern Hemisphere glaciation. The major expansion in continental ice volume at *c.* 2.75–2.72 Ma possibly resulted in an ice-sheet advance onto the East Greenland Shelf south of Scoresby Sound (Vanneste et al., 1995). Most evidence for a Pliocene ice sheet on Greenland is found in the surrounding ocean basins in the form of IRD (see below).

4.1.4. Iceland

Glaciation of Iceland began in the Miocene (*c.* 7 Ma) (Fridleifsson, 1995), occurring regularly through the Pliocene and onwards to the present-day in the mountains (Geirsdottír, 2004, 2011). The first Pliocene glaciation has been tentatively identified near Vatnajökull, in SE Iceland where it is assigned an age of *c.* 5 Ma (Fridleifsson, 1995). In sections in the area at Skaftafell/Hafrafell, a second event possibly occurred at 4.7–4.6 Ma (Helgason and Duncan, 2001). However, the oldest undisputed glacial diamictites in these sections date from *c.* 4.0 Ma, based on palaeomagnetism and K/Ar analyses. Similarly in Fljótsdalur in eastern Iceland, the oldest tillite is interbedded with lavas for which an age of 4.0–3.8 Ma has been determined (Geirsdottír and Eiriksson, 1994; Geirsdottír, 2004; Geirsdottír et al., 2007). The same sequence yielded evidence for a second glacial diamictite, which is dated to *c.* 3.4 Ma based on palaeomagnetism and is potentially related to the major glaciation during MIS M2 at *c.* 3.30 Ma. Broadly penecontemporaneous glaciation early in the Gauss Chron, between 3.6 and 3.3 Ma has been identified from the Skaftafell/Hafrafell locality (Helgason and Duncan, 2001). The initial development of a major ice sheet on Iceland occurred at *c.* 2.9 Ma, and subsequent intensifications at *c.* 2.7 Ma and between 2.5 and 2.4 Ma (Geirsdottír, 2004; Geirsdottír et al., 2007).

The Icelandic diamictites are interpreted to represent only local glaciation (Geirsdottír and Eiriksson, 1994; Geirsdottír et al., 2007), since they have only been recorded adjacent to the modern Vatnajökull ice cap. During the Neogene, this region was the most active volcanic zone of the island where volcanism gave rise to the development of a high-altitude landscape (cf. Harðarsson et al., 2008; Geirsdottír, 2011). The high precipitation in this area (3200 mm today) was, and indeed continues to be, particularly favourable for the establishment and nourishment of glaciers. The fortuitous product of this parallel development was the interbedding of lava flows and diamictites dating from between 3 and 2.5 Ma at Fljótsdalur, Jökuldalur and Skaftafell/Hafrafell. This evidence indicates the initiation of the progressive, step-like expansion and evolution of montane glaciers into ice caps in the south-east in Plio-Pleistocene time. That the ice-cap continued to expand during the earliest Pleistocene is further supported by the discovery of glacial deposits intercalated with lava flows in western Iceland and dated between 2.8 and 2.5 Ma (Geirsdottír and Eiriksson, 1994; Geirsdottír, 2011). On the basis of current knowledge, at least five glaciations are recorded in the period between 5 and 2.7 Ma (Geirsdottír et al., 2007; Geirsdottír, 2011).

4.1.5. Siberia/Arctic Russia

There are few accounts of evidence for glaciation in Siberia and Arctic Russia in the late Pliocene. Local glaciation may have occurred in the late Pliocene of the Caucasus (Milanovsky, 2008) and around the Pliocene–Pleistocene boundary in the Lake Baikal region, east Siberia (Prokopenko et al., 2001). Laukhin et al. (1999) reports two late Pliocene–early Pleistocene tills in the northeast Cukotka Peninsula, eastern Artic Russia: the Zhuravlinean Till dated at 3.5–3.2 Ma and the Okanaanean Till dated at 2.5–2.4 Ma.

The vegetation history recorded in Lake El'gygytgyn, northeast Arctic Russia, since *c.* 3.6 Ma indicates warmer than present conditions for most of the late Pliocene (Laukhin et al., 1999; Andreev et al., 2013; Brigham-Grette et al., 2013). The onset of permafrost conditions and the cooler, drier climate around *c.* 3.3 Ma (MIS M2) have been linked to the Zhuravlinean Till (Andreev et al., 2013; Wennrich et al., 2013), but climate conditions were not glacial and rather comparable to those of the Holocene (Brigham-Grette et al., 2013). Cooler and drier conditions are progressively recorded from *c.* 3.0 Ma in both the Lake El'gygytgyn (Andreev et al., 2013) and Lake Baikal (Demske et al., 2002) records.

4.1.6. Other records

In the Rocky Mountains of the USA, a Pliocene–Pleistocene–aged till is known from California (Gillespie and Clark, 2011). In mainland Europe there is no direct evidence of glaciation before the latest early to early middle Pleistocene, with the possible exception of northern Switzerland (Schlüchter, 2004; Ehlers and Gibbard, 2007).

4.2. Indirect evidence from the oceans

4.2.1. Arctic Ocean

During the Neogene, the Arctic Ocean may have been covered by perennial sea-ice and occasionally a seasonal sea-ice cover existed

(Moran et al., 2006; Darby, 2008; Krylov et al., 2008). Based on the occurrence of marine palynomorphs that require at least seasonally open waters, Matthiessen et al. (2009b) proposed a strong variability of the late Miocene sea-ice cover, with periods of seasonally open waters and periods of a perennial sea-ice cover (e.g. glacial–interglacial). From the middle Miocene onwards (*c.* 16 Ma), records of IRD from IODP Expedition 302 drill sites (e.g. the ACEX core) reflect transport via sea-ice and/or icebergs, calved from glaciers present in the Arctic Ocean. An important increase in icebergs and sea-ice is identified at *c.* 3.2 Ma in the late Pliocene (Moran et al., 2006), approximately coinciding with the major glaciation of MIS M2 at *c.* 3.3 Ma (De Schepper et al., 2009, 2013). In the central Arctic Ocean, sea-ice was probably the most dominant mode of transport, with a minor contribution of icebergs (Darby, 2008; Matthiessen et al., 2009a) until the middle to late Pleistocene when more icebergs reached the area (Spielhagen et al., 2004).

4.2.2. Barents Sea/Yermak Plateau/Fram Strait

In the Barents and Kara seas, glaciation is recorded from the early Miocene, early Pliocene and Plio-Pleistocene (Mangerud et al., 1996; Svendsen et al., 2004; Knies and Gaina, 2008; Knies et al., 2009; Mangerud et al., 2011; Vorren et al., 2011; Knies et al., 2014). The northern floor of the Barents Sea region was exposed sub-aerially in the Miocene to early Pliocene (Rasmussen and Fjeldskaar, 1996; Butt et al., 2002; Knies et al., 2009, 2014). Deposition of sand-rich material along the continental margin prior to onset of glacial erosion in the Svalbard–Barents Sea area has been related to (glacial)-fluvial erosion of Mesozoic and early Cenozoic weathering mantles (Dahlgren et al., 2005; Knies et al., 2014). Moderate to low IRD in ODP Hole 911A during the early Pliocene sequence excludes the presence of glacial ice close to the coastline, but changes in sedimentology indicate initial glacial ice build-up at around *c.* 4 Ma in the Svalbard–Barents Sea region (Knies et al., 2014). More intense glaciation in the region during the late Pliocene is indicated by frequent IRD strata from sediments on the Yermak Plateau (ODP Sites 910 and 911) between *c.* 3.6 and *c.* 2.4 Ma. Based on a new age model for ODP Hole 911A (Mattingsdal et al., 2013) and observed IRD records, Knies et al. (2009, 2014) illustrate that the northern Svalbard–Barents Sea Ice Sheet extends beyond the coastline, and probably even beyond the shelf edge, on three occasions during the late Pliocene: (1) during onset of Northern Hemisphere glaciation around 3.6 Ma sensu (Mudelsee and Raymo, 2005), (2) during MIS M2 around 3.3 Ma and (3) during glacial MIS G6/4 around 2.7 Ma. All events are severe glaciations with oxygen isotope values characteristic of early Pleistocene glaciations (Lisiecki and Raymo, 2005). At the same time, increased IRD pulses along the western Svalbard–Barents Sea margin support this interpretation (Knies et al., 2009). Simultaneous IRD events in the Fram Strait imply a coherent dynamic response of the circum-Arctic ice sheets to short-term glaciations during the latest Pliocene and early Pleistocene, i.e. at *c.* 2.7 and 2.52–2.43 Ma respectively. The low IRD frequencies along the western Svalbard margin imply deposition with little or no glacial influence from Svalbard before *c.* 2.3 Ma (Butt et al., 2000; Sejrup et al., 2005; Knies et al., 2014).

Fram Strait ODP Site 909 records nearly continuous ice-rafting with a probable Eurasian source since 18 Ma, with one distinct pulse of IRD deposition in the middle Miocene indicating large-scale glaciation in the northern Barents Sea and one in the late Pliocene (3–2.5 Ma) (Thiede et al., 1998; Winkler et al., 2002; Knies and Gaina, 2008; Knies et al., 2009; Thiede et al., 2011).

4.2.3. Greenland Sea

East Greenland Margin ODP Site 913 records the earliest known isolated IRD in the Eocene and Oligocene of the Nordic Seas (Eldrett et al., 2007; Tripati et al., 2008). IRD becomes more frequent in the Miocene to Quaternary, with highest abundance during the Pliocene and early Pleistocene when the most likely source was Greenland (Thiede and Myhre, 1995; Thiede et al., 2011). On the northeastern flank of the Scoresby Sound, the East Greenland continental margin, ODP Site 987 dropstones derived from Greenland were recorded from the Miocene (~7.5 Ma) to Pleistocene (Shipboard Scientific Party, 1996; Butt et al., 2001; Thiede et al., 2011).

Seismic identification of six glacial units south of Scoresby Sound suggest ice-sheet advance and grounding on the shelf during the Pliocene–Pleistocene, with the oldest grounding event possibly corresponding with the late Pliocene intensification of Northern Hemisphere glaciation (Vanneste et al., 1995).

4.2.4. Norwegian Sea

As elsewhere in the region, the record of IRD begins in the late Miocene at 11 Ma on the Vøring Plateau, off the north-western Norwegian coast (ODP Sites 642–644). This debris-influx continues as a series of pulse-like peaks through the latest Miocene to late Pliocene, with significant peaks around 6.5 Ma, 5.0–4.9 Ma and *c.* 4.0 Ma (Jansen and Sjøholm, 1991—updated to Geological Time Scale 2012, Hilgen et al., 2012). These IRD pulses could have been produced by Scandinavian glaciers that reached sea level. The strongest increase in both frequency and continuity occurs in parallel with an increased sedimentation rate in the late Pliocene. This input almost certainly indicates the initiation of extensive glaciation of the Scandinavian mountains, the increased intensity of ice-rafting reflecting the fact that glaciers were reaching the sea (Mangerud et al., 1996). Before this time mountain, valley or fjord glaciers predominated in the region (Jansen and Sjøholm, 1991; Jansen et al., 2000; Kleiven et al., 2002; Svendsen et al., 2004; Sejrup et al., 2005). A major increase in IRD occurs around 2.72 Ma, MIS G6 (Jansen and Sjøholm, 1991; Bailey et al., 2013).

4.2.5. Irminger Basin/Iceland Sea

By the late Miocene (at 11, 7.3 and 7.1 Ma), glaciation reached sea-level in south-eastern Greenland, indicated by IRD in the Irminger Basin (ODP Sites 914–918; Larsen et al., 1994; Helland and Holmes, 1997; St John and Krissek, 2002). ODP Site 907 in the Iceland Sea records IRD since *c.* 7.5 Ma (Fronval and Jansen, 1996). There was limited IRD input until the early Pliocene, when a peak in IRD flux is recorded at *c.* 4.9–4.8 Ma in the Irminger Basin (St John and Krissek, 2002) and Iceland Sea (Fronval and Jansen, 1996). A second early Pliocene increase in IRD in the Iceland Sea ODP Site 907 is recorded at *c.* 4 Ma (Fronval and Jansen, 1996). In the late Pliocene, an increased flux of IRD is recorded around 3.5 Ma, which St John and Krissek (2002) tentatively link to IRD deposited in the Iceland Sea at *c.* 3.3 Ma (ODP 907; Jansen et al., 2000). An important IRD increase is recorded in the Iceland Sea at *c.* 3 Ma (Fronval and Jansen, 1996), but the major increase in IRD occurs in the Iceland Sea around 2.72 Ma, MIS G6 (Jansen and Sjøholm, 1991; Bailey et al., 2013).

4.2.6. North Atlantic

There is no evidence for IRD in the early Pliocene of the North Atlantic. The late Pliocene increase in IRD frequency in the circum-Arctic oceans at *c.* 3.3–3.2 Ma is thought to have been derived mainly from Greenland (Jansen et al., 2000; Kleiven et al., 2002; St John and Krissek, 2002). DSDP Site 610 is the southernmost North Atlantic site where IRD was recorded during MIS M2 at *c.* 3.3 Ma (Kleiven et al., 2002). However high-resolution studies of this particular glacial did not confirm the presence of IRD (De Schepper et al., 2009, 2013).

It is not certain whether a coherent Greenland Ice Sheet was established prior to the latest late Pliocene, and therefore whether the debris was transported to the sea by an ice sheet or coastal-montane glaciers. The major expansion in continental ice volume at *c.* 2.7 Ma resulted in increased delivery of IRD to several Nordic Seas sites, and also to the North Atlantic: DSDP 552 (Shackleton et al., 1984), DSDP 111 and 116 (Berggren, 1972; Backman, 1979); ODP 984 (Bartoli et al., 2006); IODP U1308 (Bailey et al., 2010); IODP U1313 (Bolton et al., 2010); DSDP 607, 610 and ODP 981 (Kleiven et al., 2002; Becker et al., 2006) all record a major increase in IRD. These data from the late Pliocene allowed (Kleiven et al., 2002; Darby, 2008) to propose a somewhat coeval development of major ice sheets around the Arctic Basin. However, a geochemical provenance study of the IRD at DSDP Site 611 suggests that North Atlantic ice-rafting prior to 2.64 Ma (MIS G2) was dominated by melting of icebergs originating only from Greenland and other circum-Nordic Seas landmasses, but not North America (Bailey et al., 2013). The same authors hypothesise that a North American Ice Sheet did not expand to the North Atlantic coastline until 2.64 Ma, thus showing a delayed expansion relative to the glacial expansion in Greenland and Europe. From around *c.* 2.6 Ma, the British Ice Sheet also supplied IRD to the North Atlantic (Porcupine Basin, IODP U1317) (Thierens et al., 2012).

4.2.7. Baffin Bay and Labrador Sea

IRD, originating from northeastern Canada and Greenland, was recorded first in the Miocene (*c.* 9 Ma) and became more prominent around 3.4 Ma at ODP Site 645 in Baffin Bay (Korstgård and Nielsen, 1989; Stein, 1991; Thiede et al., 2011). Dropstones at ODP Site 646 in the Labrador Sea almost exclusively indicate a Greenland source, whereas ODP Site 647 received IRD from Greenland and occasionally also the Canadian Arctic Archipelago (Korstgård and Nielsen, 1989; Thiede et al., 1998). The records from ODP Site 646 demonstrate fluctuating delivery of IRD in low amounts between 9 and 4.5 Ma and an increase in IRD since *c.* 3 Ma (Wolf and Thiede, 1991). IRD is reported in the early late Pliocene (*c.* 3.6–3.0 Ma) of IODP U1307 (Sarnthein et al., 2009), but because of the difficulties with establishing an age model in that part of the section, it is unclear when exactly peak fluxes were recorded. The major increase in IRD occurs around 2.72 Ma (Sarnthein et al., 2009; Bailey et al., 2013).

4.2.8. North Pacific and Bering Sea

The record of glaciation on the Alaskan and neighbouring British Columbian landmass is complemented by that of ice-rafting and terrigenous sediment increases in the North Pacific. Here ice-rafting began in the late Miocene (6.6 Ma) south of Kamchatka and in the Gulf of Alaska (Krissek, 1995; Rea and Snoeckx, 1995). IRD recorded occurrences between 6.6 and 4.2 Ma in the NW Pacific and Gulf of Alaska can be related to the glaciomarine Yakataga Formation deposits of the Alaskan coastal area (Lagoe et al., 1993; Krissek, 1995; Rea and Snoeckx, 1995). In the Bering Sea, pebbles interpreted as IRD are found at >3.8 Ma, indicating the formation of sea-ice or iceberg transportation to the Bowers Ridge IODP Site U1340 (Takahashi et al., 2011). Ice-rafting increased substantially at 2.73 Ma (Krissek, 1995; Haug et al., 2005). Until this marked increase, the dropstone frequency is thought to result from montane glaciation in the surrounding landmasses, but from 2.73 Ma onwards, substantial continental-scale glaciation occurred (Prueher and Rea, 2001; Haug et al., 2005), comparable to that seen in the North Atlantic region.

5. Southern Hemisphere

Antarctic Ice Sheets began building up at the Eocene/Oligocene transition, approximately 34 Ma ago, and significantly expanded during the Miocene (Lear et al., 2000; Francis et al., 2008). Combined results from ODP drillings, seismic records, terrestrial stratigraphical evidence and model experiments indicate a step-like development from local ice caps to a large ice sheet covering the entire Antarctic continent during the Miocene (Anderson, 1999; Hambrey et al., 1991) around 13.9 Ma when climate conditions (atmospheric carbon dioxide and/or insolation) passed a threshold (Holbourn et al., 2005; Shevenell et al., 2008; Langebroek et al., 2009). The Pliocene deep-sea record from around Antarctica suggests significant sea-surface temperature fluctuations, and the oxygen isotope and eustatic records indicate that considerable ice-volume fluctuations may have occurred (e.g. Prentice and Fastook, 1990; Denton et al., 1991; Flemming and Barron, 1996). The continental seismic shelf records from the Ross Sea region and Antarctic Peninsula continental shelf show a number of shelf-wide unconformities bounding till sheets, indicating high-frequency grounding events, taken to represent waxing and waning of continental glaciers during the Pliocene (Bart, 2001). In the following we summarise the available literature describing the extent and response of the East Antarctic, West Antarctic, and Antarctic Peninsula Ice Sheets to Pliocene climate change (see also reviews in Haywood et al., 2009; Smellie et al., 2009; Davies et al., 2012; Clark et al., 2013). All Southern Hemisphere records discussed below are summarised and displayed together with reconstructions of the major ice sheets in Fig. 2.

5.1. East Antarctic Ice Sheet (EAIS)

Of particular significance to the Pliocene record is the question when the EAIS switched from a polythermal and dynamic ice sheet to a cold-based and stable ice sheet, relatively inert to short-term climate fluctuations. The Pliocene continental record in Antarctica is fragmentary and difficult to interpret in terms of climate and glaciation development. Consequently, it has led to different opinions as to the Miocene–Pliocene–Pleistocene stability of the Antarctic Ice Sheet (Wilson, 1995), and the nature, timing, extent and evolution of the ice sheets. Reconstructions of the history of the EAIS range from drastic ice-volume fluctuations, with a dynamic ice sheet fluctuating between extensive collapse and very extensive build-up and mountain overriding (e.g. Barrett et al., 1992), to suggestions that it has existed close to its present configuration for the past *c.* 14 million years (e.g. Shackleton and Kennett, 1975; Denton et al., 1993; Sugden et al., 1993; Kennett and Hodell, 1995). Much of the controversy between 'dynamicists' and 'stabilists' focuses on the interpretation of the Sirius Group glacial deposits in the Transantarctic Mountains. Fossil plant remains of the tundra shrub *Nothofagus beardmorensis* (Francis and Hill, 1996; Hill et al., 1996) and palaeosols (Retallack et al., 2001) suggest that the Sirius Group diamictons have been deposited in a significantly warmer environment than today. It is still unclear, however, whether the Pliocene marine diatoms, which have been used to date the Sirius Group, are in situ or reworked (e.g. Webb et al., 1984; Wilson, 1995; Stroeven et al., 1996; Sugden, 1996; Harwood and Webb, 1998; McKay et al., 2008). The interpretation that the Pliocene diatoms have been reworked into the sediment, probably through atmospheric transport, is supported by surface exposure dating of boulders from a moraine overlying the Sirius Group till indicating a most likely age of >5 Ma (Ackert and Kurz, 2004). Other palaeobotanical findings from the Transantarctic Mountains (McMurdo Dry Valleys, Lewis et al., 2008; Prince Charles Mountains, Wei et al., 2014; Ross Embayment, ANDRILL, Warny et al., 2009) also failed to provide evidence for a Pliocene climate warm enough to support a substantial vegetation cover in East Antarctic. Instead, all findings rather suggest a rapid cooling during the Miocene, after 14 Ma. Ice-sheet modelling results indicate that temperatures of 17–20 °C above present levels are required to melt the EAIS (Huybrechts, 1993).

Whereas recent results indicate a large-scale deglaciation of East Antarctica during the Pliocene, this appears to be rather unlikely. Many studies demonstrate that the EAIS margins significantly retreated during warm Pliocene 'interglacials'. These events must therefore have had a major influence on global sea level. Near Prydz Bay, considerable changes in the volume of the Lambert Glacier system during the Oligocene/Miocene to Plio-/Pleistocene are indicated by fjord sediments preserved as far as several hundred kilometres inland from the open coast (Hambrey and McKelvey, 2000). At Vestvold Hills, fossil-rich glaciomarine sediments suggest that in the early Pliocene (between *c.* 4.5 and 3.5 Ma) ice margins may have been ~50 km farther inland and climate was considerably warmer between 4.5 and 4.1 Ma (e.g. Pickard et al., 1988; Quilty, 1993; Whitehead et al., 2001). Major fluctuations of the Lambert Glacier System at Prydz Bay are also recorded in marine cores from ODP Leg 119 (Hambrey et al., 1991). The early Pliocene extent of the EAIS probably decreased between 4.6–4.0 Ma, when only low frequencies of IRD from icebergs released by the Amery Ice Shelf were recorded at ODP Site 1165 (Leg 188, Prydz Bay, East Antarctica) (Passchier, 2011). During at least three time intervals (*c*. 4.8–4.55, 4.4–4.3 and 3.7 Ma) the Southern Ocean was warmer by *c*. 4–5.5 °C than present (Whitehead and Bohaty, 2003; Passchier, 2011). IRD-rich layers at Site 1165 at 7, 4.8, and 3.5 Ma have been interpreted by T. Williams et al. (2010) as indicators of short-lived, massive discharges, probably reflecting destabilisation, surge, and break-up of ice streams on the Wilkes Land and Adélie Land margins of the EAIS under warmer climate conditions. By comparing multi-proxy sedimentological evidence from ODP Leg 188 (Prydz Bay) with that of Leg 178 (Antarctic Peninsula), Escutia et al. (2009) identified glacial–interglacial cyclicity at both sites between 4 and 3.5 Ma with periods of prolonged or extreme warmth correlated with MIS Gi5 (*c*. 3.71–3.68 Ma), Gi1 (*c*. 3.60–3.62 Ma), MG11 (3.56–3.58 Ma) and MG7 (3.47–3.51 Ma). Sedimentological analyses from ODP Site 745, representing the East Kerguelen Ridge sediment drift, identified three periods of enhanced accumulation of Antarctic-derived sediment, at 6.4–5.9 Ma and 4.9–4.4 Ma, potentially indicative of warmer, less stable ice sheets at these times (Joseph et al., 2002). Sea-surface warming events in the Southern Ocean were also identified at *c*. 4.5, *c*. 4.3 and *c*. 3.6 Ma by (Bohaty and Harwood, 1998) on the Southern Kerguelen Plateau (ODP Sites 748 and 751), indicating the variability of the Pliocene Antarctic climate.

A substantial increase in IRD at 3.3 Ma off Prydz Bay, coincident with the glacial MIS M2, marks the termination of the early Pliocene ice-sheet minimum and a glacial advance, possibly even to the shelf break (Passchier, 2011). The late Pliocene peaks in IRD following 3.3 Ma could be related to a major advance of the grounding line of the EAIS hundreds of kilometres beyond its present position onto the Prydz Bay continental shelf (Passchier, 2011). This author also proposed linkages between East Antarctic ice extent, global ice volume and deep-water temperatures in the late Pliocene based on the strong correlation between the IRD mass accumulation rate and high-amplitude fluctuations in the LR04 benthic stack.

Substantial fluctuation of the EAIS during the Pliocene has also been identified from sediment cores taken in the Ross Sea region by multinational drilling projects in the McMurdo Sound region (e.g. the drill cores of the Dry Valley Drilling Project (McGinnis, 1981), MSSTS-1 (Barrett, 1986), CIROS-2 (Barrett, 1989; Barrett et al., 1992), the Cape Roberts Project (e.g. Hambrey et al., 1998) and ANDRILL (e.g. Naish et al., 2007). For the early Pliocene, *c*. 4.5–3.4 Ma ago, ice sheet, sea-surface temperature, and sea ice reconstructions from the ANDRILL AND-1B sediment core provide evidence for warmer-than-present marine conditions accompanied by a diminished marine-based ice sheet in the Ross Embayment (McKay et al., 2012). During peak Pliocene warmth, sea-surface temperatures adjacent to the Antarctic coastline reached 4–5 °C (4.75–3.4 Ma) and sea-ice cover was absent or limited to winter (Levy et al., 2012). The MIS M2 glaciation (*c*. 3.3 Ma) terminated the early Pliocene warm conditions and ice-sheet minimum. At the Ross Sea Embayment, cooling after *c*. 3.3 Ma caused a rapid expansion of an ice sheet followed by a stepwise increase in sea ice between 3.3 and 2.5 Ma (McKay et al., 2012). The diatom record indicates that at 2.9 Ma sea-surface temperatures may still have reached *c*. 3 °C and this appears to represent the end of the mid-Pliocene warm period (Sjunneskog and Winter, 2012). Marine sediment cores, taken off Wilkes Land by the IODP Expedition Leg 318 (Site U1361), support previous interpretations of a dynamic East Antarctic Ice Sheet during the warm Pliocene and provide evidence for retreat of the ice sheet margin several hundred kilometres inland (Cook et al., 2013). IRD is recorded at Site U1361 during the entire Pliocene (Expedition 318 Scientists, 2011), and glaciers may have stabilised at the Wilkes Land Subglacial Basin during the later Pleistocene, <0.54 Ma ago (Orejola et al., 2013).

5.2. West Antarctic Ice Sheet (WAIS)

The WAIS is thought to have been fully established during the early to middle Miocene (Hollister and Craddock, 1976; Abreu and Anderson, 1998; Barker and Camerlenghi, 2002). Magnesium/calcium data from planktonic foraminifera demonstrate that southwest Pacific sea-surface temperatures cooled 6–7 °C between 14.2 and 13.8 Ma (Shevenell et al., 2004). During the late Miocene, the WAIS reached its modern size and might have been at times even larger than today (De Santis et al., 1999; Anderson and Shipp, 2001; Bart, 2001). However, seismostratigraphic analyses of the Amundsen Sea Embayment shelf and slope indicate that the WAIS has responded probably more sensitive to Pliocene climate change than the EAIS (Weigelt et al., 2009; Gohl et al., 2013). Moreover, it might have even collapsed during Neogene interglacials as indicated by marine deposits in the AND-1 core (purple intervals on Fig. 2) (Naish et al., 2009; Pollard and Deconto, 2009).

Many Pliocene reconstructions of the WAIS depend in part on the East Antarctic sequence, i.e. whether or not they encompass a major deglaciation of interior East Antarctica (Denton et al., 1984, 1991; Ingólfsson, 2004; Naish et al., 2009). The most detailed results on Pliocene WAIS history stem from the ANDRILL cores taken in the western Ross Sea (East Antarctica). They indicate that orbitally induced oscillations of the WAIS resulted in transitions from grounded ice and/or ice shelves to open water conditions (Naish et al., 2009). Here the authors recognised 18 orbital cycles through the Pliocene from 4.86 to 2.60 Ma. The western Ross Sea was ice free around 4.5–4.4 and again between 3.6 and 3.4 Ma (both intervals are separated by an unconformity) and the M2 glacial terminates the early Pliocene warm conditions (Naish et al., 2009). In addition, they also identified a 60 m thick diamictite unit that they interpreted as representing a warmer-than-present 'interglacial' between *c*. 3.6 and 3.4 Ma.

5.3. Antarctic Peninsula Ice Sheet (APIS)

The Pliocene history of the APIS is comparatively well constrained by terrestrial glaciomarine and glaciovolcanic sediments (Hambrey and Smellie, 2006; Smellie et al., 2008; Johnson et al., 2009; Nelson et al., 2009; Smellie et al., 2009; Salzmann et al., 2011a) and marine geological coring projects on the Pacific and Weddell Sea margin of the Antarctic Peninsula, such as IODP Leg 178 (Barker and Camerlenghi, 2002) and SHALDRILL (Anderson et al., 2011). Independent evidence from both the terrestrial and marine realms suggests that mountain glaciation of the Antarctic Peninsula already began in the latest Eocene, approximately 37–34 Ma ago (Birkenmajer et al., 2005; Anderson et al., 2011). The transition from a temperate, alpine glaciation to a dynamic, polythermal ice sheet took place during the middle Miocene (Anderson et al., 2011), and may have been related to the development of a full deep Antarctic Circumpolar Current after the mid-Miocene climatic optimum (Dalziel et al., 2013). Throughout the Neogene progressive cooling, the APIS was highly dynamic and showed a strong glacial–interglacial cyclicity with repeated advance and retreat of grounded ice masses and supply of IRD eroded on the Antarctic Peninsula (e.g. Hillenbrand and Ehrmann, 2005; Hepp et al., 2006; Scheuer et al., 2006; Cowan et al., 2008; Escutia et al., 2009; Nývlt et al., 2011). At the Weddell Sea margin, at least 10 Pliocene–Pleistocene ice-grounding events of the APIS can be identified (Smith and Anderson, 2010). Diatoms, dinoflagellate cysts and increased opal deposition in marine sediment sequences point to a strong reduction of sea-ice cover and relatively warm climatic conditions on the Antarctic Peninsula during the early Pliocene (Wei et al., 2014). Growth increment analyses of bivalves coupled with stable isotopic data indicate a largely sea-ice free shallow marine environment at Cockburn Island, Antarctic Peninsula, during the early Pliocene (*c*. 4.7 Ma) (M. Williams et al., 2010). A reduction in opal deposition between 3.1 and 2.6 Ma likely reflects sea-ice expansion and cooling (Hillenbrand and Ehrmann, 2005). A comprehensive review of palaeontological and geochemical proxy data from bivalves, bryozoans, silicoflagellates, diatoms and cetaceans for sea-surface temperature suggest that the summers on the Antarctic shelf during the late Neogene experienced most of the warming, while winter sea-surface temperatures were little changed from present (Clark et al., 2013).

The APIS was probably the most dynamic component of the Antarctic cryosphere, but several lines of evidence from marine and terrestrial deposits demonstrate that the late Neogene APIS was surprisingly robust and might have been a persistent feature despite the substantially greater-than-present warmth during the Pliocene (Ehrmann et al., 1991; Pudsey, 2001; Hillenbrand and Ehrmann, 2005; Smellie et al., 2008; Johnson et al., 2009; Nelson et al., 2009). Terrestrial glaciogenic sediments and glaciovolcanic sequences that are exposed on the outlying islands of the Antarctic Peninsula indicate a progressively thickening wet-based, actively eroding ice sheet since the late Miocene (Smellie et al., 2009). This is supported by the deep-sea $\delta^{18}O$ record, which Denton et al. (1991) interpreted as indicating that the overall Antarctic ice volume was never significantly less than today throughout the Pliocene. Ice thickness of the APIS reached 500–850 m during late Neogene glacials and draped rather than drowned the topography (e.g. Smellie et al., 2008, 2009; Davies et al., 2012). On the southern Antarctic Peninsula, ice thickness ranged from <200 m in latest Miocene time to 400 to >700 m in Pliocene–Pleistocene time (Smellie et al., 2009). There is also evidence for several ice-poor periods in the James Ross Island group, particularly clustered in the early Pliocene (Nelson et al., 2009; Smellie et al., 2009). During the ice-poor 'interglacials', the APIS front could have retreated to the present coastline and open-water marine conditions may have prevailed at times, at least seasonally. However, Pliocene climate was apparently never sufficiently warm to sustain a substantial vegetation cover on the Antarctic Peninsula. Whereas localised pockets of limited tundra vegetation still existed at least until 12.8 Ma, palynological analyses indicate no unambiguous record of land vegetation for the late Miocene and Pliocene (Anderson et al., 2011; Salzmann et al., 2011a; Warny and Askin, 2011). The presence of IRD at ODP Site 1095 implies that the Antarctic Peninsula was never deglaciated for any longer period between 4.5–3.2 Ma (Pudsey, 2001).

5.4. Patagonia

According to Coronato et al. (2004, p. 64), "*Patagonia is perhaps the best region in the Southern Hemisphere where a late Cenozoic terrestrial glacial sequence has been established*". They point out that the available absolute dating allows the recognition of several glacial events that in turn can be related to the marine isotope stratigraphy. The dating indicates that the Patagonian Andes region was glaciated from the late Miocene and through the Pliocene and moreover, that the Antarctic Peninsula was already fully glaciated in the Miocene (Coronato and Rabassa, 2011). Substantial evidence for glaciation in the Piedmont areas of Argentina and Chile is recorded since the late Miocene between 7 and 5 Ma (Rabassa, 2008; Lagabrielle et al., 2010; Rabassa and Coronato, 2011). In the Lago Cardiel region, nine late Miocene glacier advances were identified, the two oldest of which suggesting an age as old as 10.5–9 Ma (Wenzens, 2006).

A key area for the recognition and differentiation of late Cenozoic glaciation in the region is Lago Buenos Aires. Here, a series of diamicton units (tills) are interbedded with lava flows that offer the opportunity to provide a geochronological framework for the glacial events. Evidence for the earliest glacial advances was originally established by Mercer (1976) and Mercer and Sutter (1982) who undertook K/Ar dating. Near the Lago Buenos Aires, central Patagonia, one >30 m thick till was dated by K/Ar to the latest Miocene (7.4–4.4 Ma) (Rabassa et al., 2005; Rabassa, 2008). Late Miocene lava flows overlie (*c.* 5.0 Ma) and are in contact (*c.* 6.85 Ma) with till and fluvioglacial conglomerates (Lagabrielle et al., 2010). One lava flow from a series of flows interbedded with tills was dated at *c.* 4.8 Ma (Lagabrielle et al., 2007, 2010). Three till deposits underlying basalt flows were identified between 4.9 and 4.3 Ma (Rabassa et al., 2005; Rabassa, 2008). All this evidence suggests that late Miocene to early Pliocene isolated icecaps existed in the Patagonian Andes. Still at the Lago Buenos Aires region, a lava flow dated at *c.* 3 Ma overlies a different till deposit, suggesting glacial advance also in the late Pliocene (Lagabrielle et al., 2010). Malagnino (1995) presented geomorphological evidence for an ancient piedmont glaciation ('Chipanque glaciation'), which was the most extensive piedmont glaciation in the region (Coronato et al., 2004). It includes an assemblage of melt water deposits and ice-marginal morainic ridges, for which Malagnino (1995) postulated an age of 7–4.6 Ma (latest Miocene to early Pliocene) or 3.5– > 2.3 Ma (late Pliocene to early Pleistocene age) (Coronato et al., 2004; Rabassa, 2008).

In the Lago Viedma and Lago Argentino region of southern Patagonia again diamicton units are interbedded with basalt flows. Detailed morphostratigraphy has been undertaken, supported by numerical dating. Two separate till deposits were dated to *c.* 3.68–3.55 Ma and *c.* 3.55–3.48 Ma respectively, based on K/Ar dating of the over- and underlying basaltic flows (Rabassa et al., 2005). Linking the evidence of Pliocene Patagonian glaciation to global climate variability indicates that cold events and accompanied glacier advance took place during MIS MG6 (*c.* 3.5 Ma), during one of the cool events of MIS KM4, KM6, M2 or MG2 (3.45–3.20 Ma) and MIS 100–88 (2.54–2.27 Ma) (Rabassa et al., 2005; Rabassa, 2008).

The late Pliocene glacier extent was similar to the Pleistocene extent in southern Patagonia, but there is no conclusive evidence for such ice cap extent in Northern Patagonia, which probably was covered only with local ice (Rabassa, 2008). It thus remains unclear at present whether these glacial events represent local montane glaciation or the development of a substantial temporary substantial ice cap over the Andean Cordillera (Lagabrielle et al., 2010). Although, the latter is supported by the widespread distribution of diamictons (tills) across the region.

5.5. Bolivian and Columbian Andes

There is little documented evidence of tropical Andean glaciers from the Pliocene. It is estimated that the tropical Andes have attained most of their present elevation only during the past 6–5 Ma, however, the earliest glaciation recorded in the Bolivian Andes near La Paz dates from at least 3.25 Ma (Clapperton, 1983; La Frenierre et al., 2011).

Further north, detailed investigation of the sequence filling the Bogotá Basin has shown that the first glaciation of the Columbian Andes was initiated at 2.6 Ma, based on magnetic polarity and fission-track dating (Andriessen et al., 1993). Here a sudden influx of glaciofluvial sediments is recorded at the base of the reverse-magnetised Subachoque Formation, evidence that contrasts markedly with that from the pre-existing deposits of the Upper Tilatá Formation. The latter contain palynological evidence indicating warm climates during the Gauss Chron (Helmens, 2004, 2011). This indication of an earliest Pleistocene age for initial glaciation of the northern South America, is also found further south in the Bolivian Andes (Thouveny and Servant, 1989).

5.6. Australasia

The earliest records of glaciation in Australasia are found in New Zealand. They date from the Pliocene–Pleistocene boundary interval (*c.* 2.6 Ma, MIS 104–98; Barrell, 2011).

6. Synthesis

6.1. Limitations

Next to the incompleteness of the terrestrial evidence for glaciation (Fig. 2 and Section 3 above), the dating of the glacial deposits also limits the interpretations of glacial extent and chronology. Absolute dating of terrestrial glacial deposits is considerably more difficult than that of marine deposits, which often have detailed magnetostratigraphy and orbitally-tuned isotope stratigraphy available. The ages of glacial diamictons and moraines reviewed here are mainly derived from magnetostratigraphic and absolute dating of overlying and/or underlying lava flows (e.g. Iceland, Patagonia). When tills are interbedded between lava flows a relatively accurate age constraint can be derived, but this is more difficult when only an overlying or underlying till is available. We have not attempted to improve the age estimates of glaciation events for each region, but have updated it to the most recent time scale (Geological Time Scale 2012, Hilgen et al., 2012) where necessary. By assuming the timing of the events to be relatively accurate, it was nevertheless possible to use the evidence for glaciation from both the terrestrial and marine realms to identify the evolution of the ice sheets in both the Northern and Southern Hemisphere. Examining the plots of the distribution of glaciation globally through the late Neogene (Fig. 2) suggests that there are three broadly distinct periods: (i) the Zanclean (5.33–3.60 Ma), when low frequency IRD is recorded at high latitude sites in the Northern Hemisphere, and two glaciation events extending beyond Greenland and Antarctica at *c.* 4.9 and 4.0 Ma were identified; (ii) gradual expansion of the Northern Hemisphere glaciers through the Piacenzian from *c.* 3.6 Ma, undergoing a marked step around 3.3 Ma, a set-back during the mid-Piacenzian Warm Period (*c.* 3.3–3.0 Ma), and (iii) culmination of the glacial/interglacial cyclicity around 2.7 Ma, just prior to the Pliocene–Pleistocene boundary, with major glacial expansion in the Northern Hemisphere that marks the transition to the Quaternary.

The glaciation events are compared to Pliocene climatic and oceanographic events in the synthesis below.

6.2. Two glaciation events in the warm early Pliocene (5.3–3.6 Ma)

In the early Pliocene (5.3–3.6 Ma), there is very little terrestrial evidence of glaciation, but occasional records are present in South America and the Northern Hemisphere. The limited evidence of ice sheets may be an expression of a lack of records, but are just as likely attributable to the generally warmer climate and pCO_2 concentrations well-above 400 ppm prior to *c.* 4.5 Ma (Pagani et al., 2009; Seki et al., 2010). Sea-surface temperatures in the Southern Ocean were 4–5.5 °C higher than present during three intervals (Whitehead and Bohaty, 2003; Passchier, 2011), more than 2–3 °C higher in the subtropical North Atlantic (Naafs et al., 2010) and over *c.* 6 °C higher in the eastern North Atlantic (Lawrence et al., 2009). The benthic isotope sequence (Lisiecki and Raymo, 2005) shows only small amplitude glacial/interglacial variations in the early Pliocene and most glacials probably remained warmer than today. In the southern Hemisphere, land sections and Southern Ocean IRD records and sediments in drill cores reflect a large EAIS that underwent occasional retreat, a persistent APIS, a WAIS that undergoes six expansion/reduction phases and a frequently ice-free Ross Sea between 5.0 and 4.3 Ma (Naish et al., 2009). There are no reports of early Pliocene IRD in the North Atlantic, but IRD is recorded in the Arctic Ocean, North Pacific/Bering Sea, Labrador Sea, Baffin Bay and the Nordic Seas. It is not always certain whether this IRD derives from sea ice or iceberg calving, although dropstones from the Labrador Sea/Baffin Bay and Greenland Sea indicate an origin from Greenland and the Canadian Arctic archipelago (Korstgård and Nielsen, 1989).

6.2.1. Northern and Southern Hemisphere glaciation at c. 4.9–4.8 Ma

A first globally recognisable glacial event, recorded in both the marine and terrestrial realms, can be identified at *c.* 4.9–4.8 Ma (① in Figs. 2, 3, 4). The first Pliocene recorded glacial deposits in the Ross Embayment (*c.* 4.9 Ma) largely coincide with IRD in Prydz Bay from the EAIS (*c.* 4.8 Ma), glacial deposits in the McMurdo Sound region drill cores (5.0–4.8 Ma) and a longer phase of increased Antarctic-derived sediment on the Kerguelen Plateau (4.9–4.4 Ma). The increased IRD and sediment derived from Antarctica in the early Pliocene has been interpreted as reflecting less stable ice sheets as a consequence of warmer conditions (Joseph et al., 2002; T. Williams et al., 2010), which seemingly contrasts with the mountain glaciation in Patagonia (*c.* 4.8 Ma). Although IRD can be produced during melting or instabilities of the ice sheet as well as during glacial advance and glacial maxima (Joseph et al., 2002; Naish and Wilson, 2009; T. Williams et al., 2010), the IRD evidence nevertheless suggests that an early

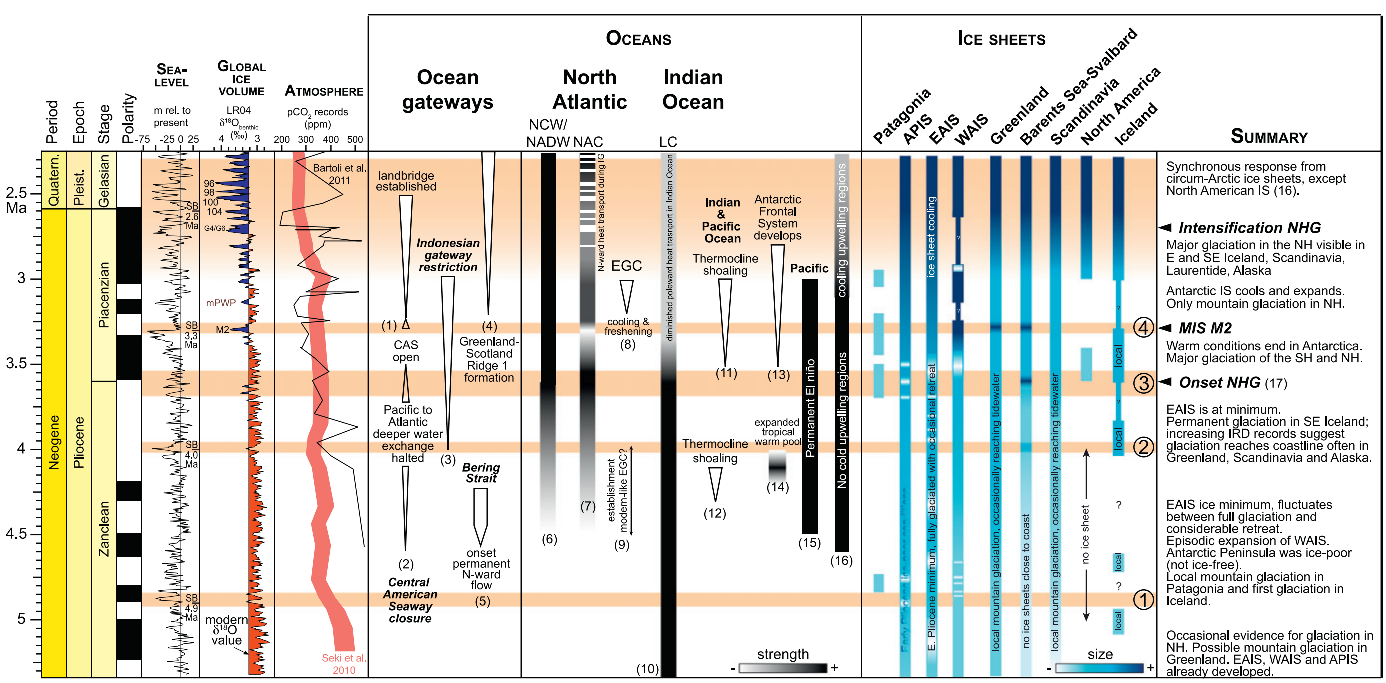


Fig. 3. Comparison of the Pliocene palaeoclimatic (sea level, global ice volume, atmospheric carbon dioxide) and palaeoceanographic (gateways, currents) conditions to the evolution of the Northern and Southern Hemisphere ice sheets. The benthic isotope record LR04 is from Lisiecki and Raymo (2005), sea level and sequence boundaries are from Miller et al. (2005). Abbreviations: NCW = Northern Component Water, NAC = North Atlantic Current, EGC = East Greenland Current, LC = Leeuwin Current, THC = Thermohaline circulation, mPWP = mid-Piacenzian Warm Period aka. mid-Pliocene Warm Period. References: (1) De Schepper et al., 2013; (2) Haug and Tiedemann, 1998, Haug et al., 2001, Steph et al., 2006, Groeneveld et al., 2006, Schmidt, 2007; (3) Cane and Molnar, 2001, Karas et al., 2009, Karas et al., 2011a,b; (4) Poore et al., 2006; (5) Marincovich and Gladenkov, 1999, Marincovich, 2000, Verhoeven et al., 2011; (6) Haug and Tiedemann, 1998, Ravelo and Andreasen, 2000, Poore et al., 2006, Schreck et al., 2013; (7) Lawrence et al., 2009, 2010, Naafs et al., 2010, De Schepper et al., 2009, 2013; (8) Sarnthein et al., 2009; (9) Bohrmann et al., 1990, Schreck et al., 2013; (10) Karas et al., 2011b; (11) Karas et al., 2009, Fedorov et al., 2006; (12) Larivière et al., 2012; (13) Karas et al., 2011a; (14) Brierley et al., 2009; (15) Wara et al., 2005; (16) Marlow et al., 2000, Dekens et al., 2007; (17) Kleiven et al., 2002, Bailey et al., 2013; (18) Mudelsee and Raymo, 2005.

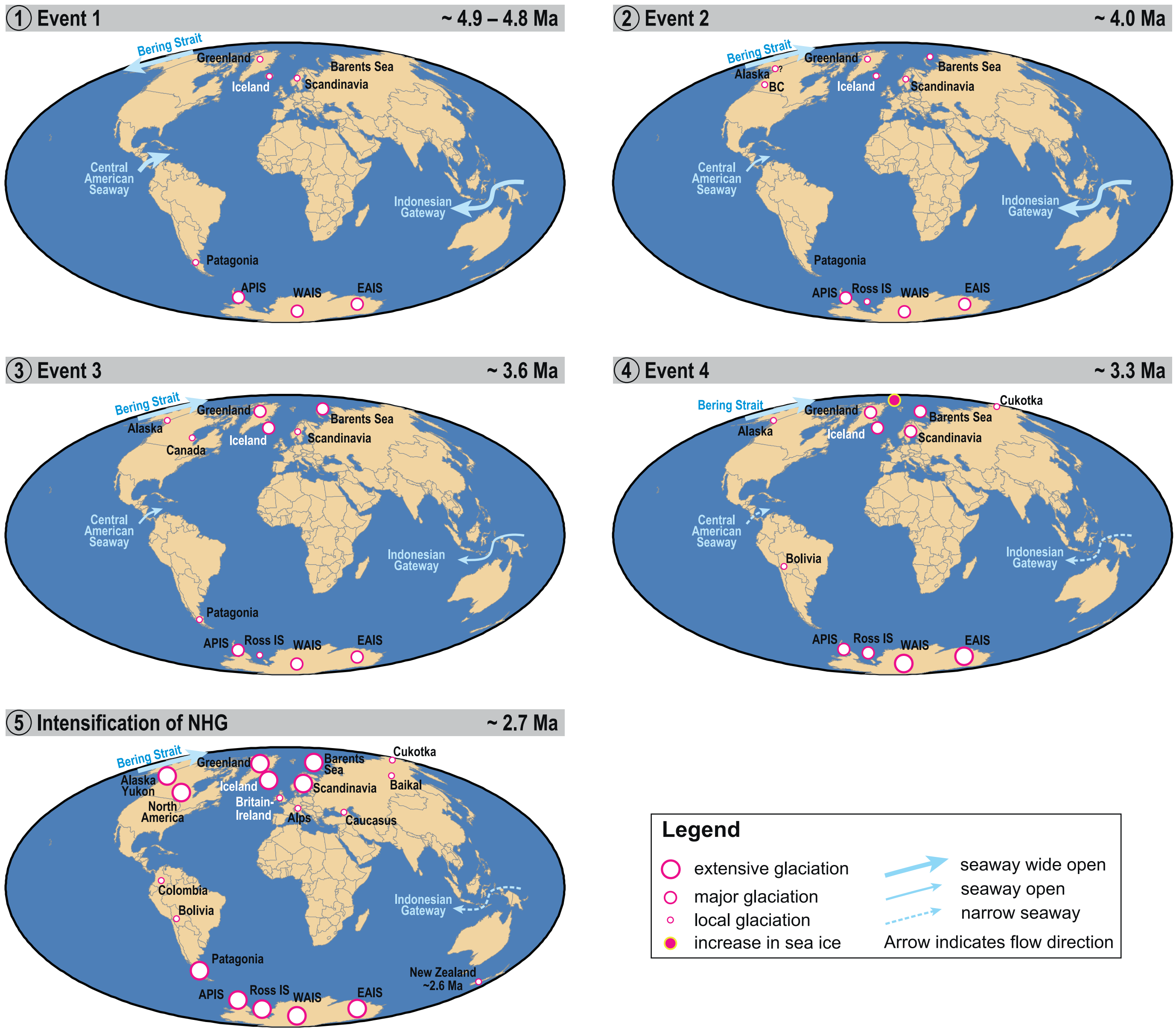


Fig. 4. Schematic representation of the distribution of ice sheets during the Pliocene glaciation events at around 4.9–4.8, 4.0, 3.6 and 3.3 Ma, and the intensification of Northern Hemisphere Glaciation around 2.7 Ma.

Pliocene Antarctic Ice Sheet was large enough by *c.* 4.9–4.8 Ma at least to reach the coastline. In addition to the evidence from the Southern Hemisphere, also evidence for glaciation in the Northern Hemisphere is recorded on Iceland (*c.* 5.0 and 4.7 Ma). Furthermore, IRD pulses from Greenland are detected in the Irminger Basin and Iceland Sea at *c.* 4.8 Ma, and from Scandinavia in the Norwegian Sea at *c.* 5.0–4.9 Ma.

Assuming the timing of these events is robust, they can be attributed to the glacial MIS Si6 and/or Si4 (Lisiecki and Raymo, 2005) between 4.9 and 4.8 Ma, providing evidence for the first early Pliocene bipolar expansion of the continental Northern and Southern Hemisphere ice sheets to the coastline (① in Figs. 2, 3, 4). The glacial expansion also lowered level to just below present-day levels (Naish and Wilson, 2009) or as much as about −45 m, and is expressed in a global sequence boundary at 4.9 Ma (Miller et al., 2005). This implies that continental ice volume may have been comparable to present, or maximally marginally larger. The cause of this glaciation is unclear, but because the glaciation event predates the early Pliocene oceanographic reorganisations arising from gateway changes (Fig. 3), these did not seem to have an effect on the expansion of continental ice sheets at this time.

6.2.2. Early Pliocene changes at ocean gateways did not immediately impact glaciation

The Central American Seaway shoaling/closure has been put forward as a possible cause for Northern Hemisphere glaciation (e.g. Keigwin, 1982). Between 4.6 and 4.2 Ma, the Central American Seaway shoals considerably and restricts deeper water exchange from the Pacific to the Caribbean Sea and Atlantic (Haug and Tiedemann, 1998; Groeneveld et al., 2006; Steph et al., 2006). An immediate and causal effect of the Central American Seaway closure on the Northern Hemisphere ice sheets is not confirmed in the terrestrial and marine records, nor in the small amplitude variations of the benthic isotope record (Figs. 2, 3). However, recent modelling studies (Lunt et al., 2008a,b) corroborate the conclusion that the closure of the Central American Seaway did not lead to expansion of the Greenland Ice Sheet.

Around the same time as the Central American Seaway shoaling, Pacific water flow through the Bering Strait reversed and began to flow northwards into the Arctic and eventually North Atlantic (*c.* 4.5 Ma; Marincovich, 2000; Verhoeven et al., 2011). Although for a different time period (3.2–3.0 Ma), Sarnthein et al. (2009) argue that strengthening the East Greenland Current, arising from increased northward Pacific water flow via the Bering Strait, is essential for Greenland Ice Sheet expansion because it isolates the continental ice mass from warm Atlantic waters entering the Nordic Seas from the south. Although an East Greenland Current, similar to that of today, may have developed around *c.* 4.5–4.0 Ma (Bohrmann et al., 1990; Schreck et al., 2013), this also did not seem to have had an immediate effect on the Northern Hemisphere Ice Sheet (Figs. 2, 3).

The Central American Seaway nevertheless appears to play an important role in the early Pliocene thermocline depth. A shoaling/closure of the Central American Seaway caused a shoaling of the thermocline, which lead to coupling of the thermocline with sea-surface temperatures and increased the climate sensitivity in the Pliocene (Larivière et al., 2012). These authors suggest an increased sensitivity of the early Pliocene climate system, with stronger coupling of pCO_2, sea-surface temperatures and climate, and illustrating that Pliocene climate became more susceptible to perturbations, including changes in atmospheric carbon dioxide.

6.2.3. Northern Hemisphere glaciation at c. 4.0 Ma

Around *c.* 4.0 Ma, MIS Gi22/Gi20 corresponds to a major sequence boundary (Miller et al., 2005) and marks a second pronounced early Pliocene expansion of global ice volume (Lisiecki and Raymo, 2005) under atmospheric pCO_2 concentrations of 300–350 ppm (Seki et al., 2010). The glacial expansion of MIS Gi22/Gi20 seems to be mainly represented in the Northern Hemisphere (② in Figs. 2, 3, 4). The glacial expansion in Greenland, Iceland and British Columbia (Canada) around *c.* 4.0 Ma occurred at a time when uplift on Greenland (Japsen et al., 2006) and along the norwestern European continental margin (Knies et al., 2014) may have reached a critical point allowing glaciation. At the same time, increased recovery of IRD is observed in the Iceland Sea and off Norway, which suggests extension of the Scandinavian glaciers to the coast. Sea-ice and icebergs are recorded in the Arctic Ocean (Moran et al., 2006) and IRD from Alaska and British Columbia becomes more frequent in the North Pacific from *c.* 4.2 Ma onwards, evidencing that glaciers are present around the circum-Arctic oceans. Evidence for the glaciation in the mountains of British Columbia and possibly also SE Alaska is found around 4.0 Ma. Also in the Bering Sea, IRD is recorded consistently starting before 3.8 Ma.

The early Pliocene EAIS is generally characterised by an ice minimum and fluctuations in size between a full cover to a substantial retreat. Southern Hemisphere glacial deposits in the McMurdo Sound drill cores (CIROS-2, DVDP-10 DVDP-11), and considerable IRD offshore East Antarctica (Site U1361) and on the Kerguelen Plateau could correspond to the glacial expansion at *c.* 4.0 Ma. IRD was recorded in low frequency in Prydz Bay between 4.6 and 4.0 Ma, but this increases considerably from 4.0 Ma onwards suggesting a shift from a retreated Antarctic ice margin to a larger Antarctic Ice Sheet (Naish et al., 2009; Passchier, 2011). Because a glacial erosional surface was encountered in the ANDRILL core, probably corresponding to the 4.3 to 3.6 Ma interval (Naish et al., 2009), it is not possible to assess the state of the WAIS at that time.

Around *c.* 4 Ma, Brierley et al. (2009) reports a reduced meridional sea-surface temperature gradient between the equator and the subtropics, and thus evidence for an expanded tropical warm pool. Based on model simulations, these authors propose important poleward heat transport and weakening of the Hadley circulation in both hemispheres at *c.* 4 Ma (Fig. 3). In such globally warm world, it is unlikely that ice sheets could have expanded. As a result of the frequent fluctuations of the southern hemisphere ice sheets at this time, occurrence of warming events (Fig. 2), and limited age control on the glacial deposits, it is difficult to be certain of an expansion of the Antarctic Ice Sheets at *c.* 4.0 Ma. In the Northern Hemisphere, the several lines of evidence indicate glaciation in Greenland, Iceland and British Columbia around *c.* 4.0 Ma, possibly as a result of circum-Arctic regional uplift and mountain building (Japsen et al., 2006; Solgaard et al., 2013; Knies et al., 2014). The mismatch between the warm global climate and the evidence for glaciation could be the result of the poor age control on the terrestrial records of glaciation, although the IRD record in the Norwegian Sea, the major sequence boundary, and the high benthic isotope values during MIS Gi20 are well constrained at *c.* 4.0 Ma. If not an issue of age models and/or difference in resolution between Brierley et al. (2009) and this study, possibly increased seasonality in the Northern Hemisphere (Knowles et al., 2009) might explain the contradiction between an expanded tropical warm pool and glaciation in the Northern Hemisphere at this time.

6.2.4. Other early Pliocene glaciations?

Glacial expansion during MIS Gi14 and Gi12 (3.9–3.8 Ma)—as interpreted from the global benthic isotope stack—to a continental ice-volume comparable or just larger than today apparently did not produce significant IRD pulses or terrestrial evidence for glaciation (Fig. 2). Possibly glaciers and ice sheets did not reach the coast, evidence is not yet discovered and/or has been removed by subsequent glaciations. The absence of a major sequence boundary around that time would rather suggest that ice volume expansion, and associated sea-level fall, was not that large. It is true that changes in intermediate water temperature may also affect the global benthic isotope record, which is not only an expression of ice volume (e.g. Zachos et al., 2001).

6.3. Early/Late Pliocene transition (c. 3.6 Ma) marked by expansion of the Northern Hemisphere ice sheets

The general opinion is that Northern Hemisphere glaciation developed as a gradual transition, with the onset around 3.6 Ma and leading to the intensification around 2.7 Ma (e.g. Ravelo et al., 2004; Mudelsee and Raymo, 2005). The gradual increase in ice volume through the Piacenzian resulted in a glacio-eustatic sea-level lowering of c. 43 m superimposed on short intervals of glacial expansions at *c.* 3.3 Ma and 2.7 Ma, and the mid-Pliocene climate optimum (3.3–3.0 Ma) (Mudelsee and Raymo, 2005). Several high- and low latitude sea-surface temperature records demonstrate a gradual yet important cooling starting around *c.* 3.6 Ma (e.g. Dekens et al., 2008; Lawrence et al., 2009; Martinez-Garcia et al., 2010; Naafs et al., 2010).

Around *c.* 3.6 Ma, evidence for glaciation is found mainly in the Northern Hemisphere terrestrial and marine data. Expansions of the Svalbard–Barents Sea Ice Sheet, and the occurrence of ice caps in Alaska, northern and eastern Canada and Iceland are documented (③ in Figs. 2, 3, 4). The suggestion of a considerable North American Ice Sheet at *c.* 3.5 Ma (e.g. James Bay Lowland, Canada in Gao et al., 2012) contrasts according to Brigham-Grette et al. (2013) with the relatively high Arctic temperatures recorded in Eastern Siberia. Also the possible presence of boreal forest on Ellesmere Island (Csank et al., 2011a,b; Rybczynski et al., 2013) during that interval seems conflicting, however the dating of the Ellesmere Island deposits suggests an age of 3.8–3.4 Ma and, taking the maximum error estimates into account, these deposits may be early as well as late Pliocene (Fig. 2).

Records from the Southern Hemisphere are also somewhat contradictory. A local ice cap is reported in Patagonia, but the abundant IRD in the Ross Embayment, Prydz Bay and on the Kerguelen Plateau might reflect the destabilisation of the EAIS margin during short-lived warming events of the surface waters in the Southern Ocean rather than an expansion of the ice sheet. Nevertheless, the Antarctic Ice Sheet must have reached the coast to release the large amount of IRD recorded in the Southern Ocean at that time.

The evidence of glaciation cannot be tied to one or more marine isotope stages in the benthic isotope stack (Lisiecki and Raymo, 2005) during the Gauss Chron, but they are close to the well-expressed Gilbert Chron MIS Gi2 and/or Gi4 (*c.* 3.7–3.6 Ma, latest early Pliocene).

6.4. Major global glaciation at c. 3.3 Ma (MIS M2)

Late Pliocene North Atlantic sea-surface temperature records (Lawrence et al., 2009, 2010; Naafs et al., 2010) demonstrate a 2–3 °C decline after 3.5 Ma reaching temperatures comparable to the present-day at around 3.3 Ma (De Schepper et al., 2013). Siberian Arctic climate became relatively dry and cool at this time, but conditions remained Holocene-like but not 'glacial' in the Quaternary sense (Brigham-Grette et al., 2013). This led these authors to conclude that a significant ice-sheet expansion must have occurred on Antarctica to explain the large benthic isotope shift during MIS M2. Indeed, in the Southern Hemisphere, MIS M2 is expressed in the termination of warm conditions in the Ross Sea, seaward expansion of the ice shelf beyond the Ross Sea, termination of EAIS minimum followed by cooling of the Antarctic Ice Sheet (Naish et al., 2009; Passchier, 2011; McKay et al., 2012) (④ in Figs. 2, 3, 4). The Patagonian and Bolivian Andes were also covered by an ice cap at this time (Rabassa et al., 2005; La Frenierre et al., 2011). Nevertheless, our review of the terrestrial and marine evidence demonstrates a major glacial expansion in the Northern Hemisphere at *c.* 3.3 Ma, which probably can be seen best as a transition from occasional, local glaciers and ice caps to a more modern-like Northern Hemisphere ice configuration (④ in Figs. 2, 3, 4). It is the first time in the Pliocene that the Greenland Ice Sheet became the major source of IRD at *c.* 3.3 Ma (Jansen et al., 2000; Kleiven et al., 2002). The sediment records in the circum-Arctic oceans indicate that the Greenland and also the northern Svalbard–Barents Sea Ice Sheets extended beyond the coastline. The Icelandic Ice Sheet also increased its size in the south-east of the island after 3.6 Ma, reflecting a step-like expansion and evolution of mountain glaciers into one larger ice cap (Geirsdottír, 2011). In addition, terrestrial evidence suggests the presence of an ice cap on North America, although the timing and extent of these glaciations is uncertain. The cooling heralded permafrost conditions in the Siberian Arctic whilst the vegetation record reveals significant cooling to conditions probably comparable to the Holocene average (Brigham-Grette et al., 2013; Wennrich et al., 2013).

The marine and terrestrial evidence of glaciation around 3.3 Ma corresponds to MIS M2, a major feature in the benthic isotope stack that is also characterised by a major sequence boundary (Figs. 2, 3). Sea-level estimates for this glacial maximum during MIS M2 vary from >−10 m (Wanganui section; Naish and Wilson, 2009), −40 m ± 10 m (based on the benthic isotope stack; Miller et al., 2012) and even to −65 m ± 15–25 m (Mg/Ca of ostracods; Dwyer and Chandler, 2009) compared to present. Around this time, a tectonically reduced Indonesian Throughflow led to less pole-ward heat transport in the Indian Ocean, a weakened Leeuwin Current and development of a modern Antarctic Frontal system (Karas et al., 2011a,b) (Fig. 3). The establishment of such a system and consequent further thermal isolation of the Antarctic Ice Sheet, may have facilitated cooling and expansion of the Antarctic Ice Sheet. The mechanism for the extensive Northern Hemisphere glaciation may be related to reduction in Pacific-to-Atlantic flow through the Central American Seaway that lead to a southward shift of the North Atlantic Current and polar front in the North Atlantic (De Schepper et al., 2009, 2013). This cooled the higher latitudes surface waters to temperatures comparable to present and caused a major expansion of the Northern Hemisphere ice sheets, while also the Southern Hemisphere Ice Sheet expanded.

6.5. Mid-Piacenzian Warm Period (3.27–2.97 Ma)

Modelling studies suggest that the Antarctic and Greenland ice sheets retreated during the mid-Piacenzian Warm Period to become considerably smaller than today (Hill et al., 2007, 2010; Dolan et al., 2011). During this interval, the reconstructed global annual mean temperature was 2–3 °C higher than pre-industrial levels (Haywood et al., 2009; Dowsett et al., 2012; Lunt et al., 2012; Haywood et al., 2013), the Arctic region was markedly warm (Brigham-Grette et al., 2013, and references therein) and atmospheric pCO_2 concentrations were in the range of 330–450 ppm (Fig. 3). Global ice volume was reduced as reflected in the higher than present level estimated to be +10 to +40 m (Raymo et al., 2011), averaging around +20–25 m (Miller et al., 2012). Nevertheless, the reconstruction of ice sheets remains particularly difficult during this time.

IRD is recorded off Antarctica and glacial deposits were recovered in the McMurdo Sound region, but Patagonia probably remained ice-free (Fig. 2). In the Northern Hemisphere, the Siberian Arctic showed relatively warm conditions, and there is no indication of extensive glaciation in the terrestrial records (Fig. 2). IRD is regularly recorded in the circum-Arctic oceans at this time, but in comparable low quantities as in the warm early Pliocene. Modelling experiments of the Greenland Ice Sheet during the mid-Piacenzian Warm Period indicate a smaller Greenland Ice Sheet compared to present. They place most ice in the mountains of East Greenland, from where it was able to expand into more low-lying central Greenland areas (Hill et al., 2010; Dolan et al., 2011). Interestingly, a modelling study revealed variability for both the Greenland and Antarctic Ice Sheets due to orbital forcing. This variability could be in anti-phase, meaning that the largest ice sheet reduction in Antarctica may correspond to the smallest reduction of the Greenland Ice Sheet (Dolan et al., 2011).

6.6. Late Pliocene intensification of the Northern Hemisphere glaciation, c. 2.7 Ma

From 3.1 Ma onwards, sea-ice expansion arising from climatic deterioration is evident in the Southern Ocean (Hillenbrand and Ehrmann, 2005). Observations from the ANDRILL drill holes indicate that during the latest Pliocene the Antarctic Ice Sheet underwent progressive cooling, which culminated in a major cooling step and a major expansion around the Pliocene–Pleistocene boundary (Naish et al., 2009). At the same time, the Greenland, Iceland, Svalbard–Barents Sea and Scandinavian Ice Sheets gradually expanded to reach the extensive size typical of the late Pliocene/early Pleistocene glacial/interglacial oscillations. In Scandinavia, glaciation remained mainly restricted to mountain, valley or possibly fjord glaciers until *c.* 2.7 Ma (Jansen and Sjøholm, 1991; Jansen et al., 2000; Kleiven et al., 2002; Sejrup et al., 2005). This supports the interpretation of the Svalbard–Barents Sea Ice Sheet, where a synchronous moderate IRD flux and kaolinite/illite-rich sediment supply along its northern margin implies the occurrence of glaciation reaching the coastline only during short-term glacial events (Knies et al., 2014). Direct evidence of glaciation and distinct IRD pulses in the North Atlantic (e.g. Bailey et al., 2013), Fram Strait–Barents Sea region (Knies et al., 2009, 2014) and the mid-Norwegian margin at 2.74 Ma (Jansen et al., 2000; Kleiven et al., 2002) indicates ice-sheet expansion on Greenland and the Scandinavian Peninsula around *c.* 2.72 Ma (Figs. 2, 3, 4). This expansion was thought to reflect a synchronous response across the Northern Hemisphere to deteriorating climate, thus also including the North American Ice Sheet (Kleiven et al., 2002). However, recent IRD fingerprinting in the North Atlantic indicates that Greenland and Scandinavia were the major source of icebergs until MIS G2, whereas the contribution from a North American Ice Sheet was only found from 2.64 Ma onwards (Bailey et al., 2013). While ice sheets existed somewhere on North America (Naafs et al., 2012), the extension of a North American Ice Sheet to the coastline and a size sufficiently large to produce significant numbers of icebergs was delayed relative to the Greenland and Scandinavian Ice Sheets (Bailey et al., 2013). At *c.* 2.6 Ma, the British and Irish Ice Sheet also became a source for IRD in the North Atlantic (Thierens et al., 2012). The later expansion of the more southern North American Ice Sheet (and British/Irish Ice Sheets) is in agreement with a gradual forcing resulting from decreasing atmospheric carbon dioxide concentrations (Deconto et al., 2008; Seki et al., 2010; Bailey et al., 2013).

We do not intend to review mechanisms responsible for the initiation of the Pleistocene glaciation, but refer to the literature for detailed discussions (e.g. reviews in Raymo, 1994; Ravelo et al., 2007). The debated mechanisms for intensification of Northern Hemisphere glaciation include: the influence, delaying effect or a necessary precondition of the closure of the Central American Seaway (Berger and Wefer, 1996; Haug and Tiedemann, 1998; Bartoli et al., 2005; Klocker et al., 2005; Lunt et al., 2008a,b), changes in orbital parameters (Maslin et al., 1998); stratification, development of a permanent halocline and increased seasonality in the North Pacific (Haug et al., 1999), declining carbon dioxide levels (Lunt et al., 2008a; Pagani et al., 2009; Seki et al., 2010; Bartoli et al., 2011), tectonic uplift of the Greenland–Scotland Ridge (Poore et al., 2006), variation in sea-floor spreading rates (Raymo, 1994), narrowing of the Indonesian Seaway and associated changes in tropical Pacific oceanography (Cane and Molnar, 2001), cooling of upwelling regions (Marlow et al., 2000), termination of early Pliocene “permanent El Niño conditions” (Ravelo et al., 2004), changes in thermocline depth (Fedorov et al., 2006), and necessary tectonic uplift in the circum-Arctic region (Knies et al., 2014). In addition, changes in the Southern Ocean frontal system may have led to reduced Atlantic Meridional Overturning Circulation and heat transport to high latitudes (McKay et al., 2012).

7. Conclusions

We have compiled all available terrestrial and marine evidence for glaciation in the Pliocene, bridging a gap between the terrestrial and marine communities investigating Pliocene climate and cryosphere (Figs. 2, 3). The Pliocene glacial history is mainly a history of the progressive glaciation of the circum-Arctic region, and cooling and waxing of the EAIS and WAIS (Fig. 4). Already in the Miocene, East Antarctica became fully glaciated, limiting further geographical expansion of continental ice sheets in the Southern Hemisphere.

Our review of the terrestrial and marine records primarily emphasises the paucity of terrestrial records, which sometimes have poor age control. Nevertheless, it appears that major Pliocene glacial expansion occurred on at least four different occasions (*c*. 4.9–4.8, 4.0, 3.6 and 3.3 Ma) prior to the intensification of the Northern Hemisphere glaciation in the latest Pliocene. These events are exceptions in the otherwise globally warm Pliocene, when atmospheric carbon dioxide concentrations were still considerably higher than present, but they may be a consequence of the increased climate sensitivity in the Pliocene (Pagani et al., 2009; Larivière et al., 2012). Although it is not possible to identify the mechanism(s) causing each of the events, the early Pliocene events (*c*. 4.9–4.8 and 4.0 Ma) do seem to be unrelated to the oceanographic change due to reorganisation of the Pacific–Atlantic gateways (Central American Seaway, Bering Strait). In contrast, the late Pliocene glaciation in the Northern Hemisphere during MIS M2 (3.3 Ma) might in fact be caused by changes in Pacific-to-Atlantic flow via the Central American Seaway and consequent changes in North Atlantic surface. The direct and indirect evidence for an extensive Northern and Southern Hemisphere glaciation around *c*. 3.6 Ma is not reflected in global ice volume records. This could be an artefact of poor age control on these records, since a major increase in global ice volume characterises the earlier MIS Gi2 and Gi4 (3.7–3.6 Ma). It is clear that the cause(s) for each of these events are not fully understood at present, and requires further detailed investigation.

Anti-phasing of the Greenland and East Antarctic ice sheets has been demonstrated by modelling studies for the mid-Piacenzian Warm Period (Dolan et al., 2011). Because we reviewed data on the long-term Pliocene time scale and accurate ages of terrestrial evidence for glaciation are scarcely available, we could not identify any (anti-)phasing of the Greenland and Antarctic Ice Sheets. Future high-resolution studies of the marine and terrestrial realms for the four identified Pliocene glacial events (Fig. 2) may be able to determine the synchronicity of each ice-sheet expansion. For example, using geochemical fingerprinting of IRD, a delayed response of the North American Ice Sheet to the intensification of the Northern Hemisphere glaciation has been demonstrated (Bailey et al., 2013).

Examination of the frequency of glaciation through the Neogene indicates that Southern Hemisphere glaciation, having been established principally in Antarctica and southern South America first, occurred continuously from the early Neogene to the present day. By contrast, Northern Hemisphere glaciation, although initially somewhat restricted, increased markedly around the beginning of the Quaternary, increasing again in frequency in the latest early Pleistocene and reaching very high intensity in the middle to late Pleistocene. The declining carbon dioxide concentrations in the atmosphere (Seki et al., 2010), possibly in combination with necessary preconditions (e.g. circum-Arctic tectonic uplift, Central American Seaway closure, Bering Strait through flow), were probably ultimately responsible for the intensification in Northern Hemisphere glaciation at the Pliocene–Pleistocene transition. Until *c*. 2.8 Ma, the Pliocene is characterised with low-amplitude glacial–interglacial oscillations, marked with only few larger amplitude glacial events (e.g. MIS Si4/S6, G20, M2). Whilst these Pliocene glaciations were substantial, they remained less extensive than their Pleistocene counterparts, although this observation may be biased by the incompleteness of the geological record. Under the globally warm conditions and high atmospheric carbon dioxide concentrations of the Pliocene, it is improbable that large Pleistocene-like glaciations could occur in the Northern Hemisphere.

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Appendix A. Abbreviations used in text

IRD	ice-rafted debris or detritus
MIS	marine isotope stage
DSDP	Deep Sea Drilling Project
ODP	Ocean Drilling Program
IODP	Integrated Ocean Drilling Program
GIS	Greenland Ice Sheet
EAIS	East Antarctic Ice Sheet
WAIS	West Antarctic Ice Sheet
APIS	Antarctic Peninsula Ice Sheet

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