Earth-Science Reviews 104 (2011) 199–212

Contents lists available at ScienceDirect

Earth-Science Reviews

journal homepage: www.elsevier.com/locate/earscirev

Post-glacial regional climate variability along the East Antarctic coastal margin—Evidence from shallow marine and coastal terrestrial records

Elie Verleyen [a,*,1], Dominic A. Hodgson [b,1], Koen Sabbe [a], Holger Cremer [c], Steven D. Emslie [d], John Gibson [e], Brenda Hall [f], Satoshi Imura [g], Sakae Kudoh [g], Gareth J. Marshall [b], Andrew McMinn [h], Martin Melles [i], Louise Newman [j], Donna Roberts [k], Steve J. Roberts [b], Shiv M. Singh [l], Mieke Sterken [a], Ines Tavernier [a], Sergey Verkulich [m], Evelien Van de Vyver [a], Wim Van Nieuwenhuyze [a], Bernd Wagner [i], Wim Vyverman [a]

[a] Protistology & Aquatic Ecology, Department of Biology, Ghent University, Krijgslaan 281, S8, B-9000 Ghent, Belgium
[b] British Antarctic Survey, Natural Environment Research Council, High Cross, Madingley Road, Cambridge CB3 0ET, UK
[c] TNO Built Environment and Geosciences, Geological Survey of the Netherlands, Princetonlaan 6, 3584 CB Utrecht, The Netherlands
[d] University of North Carolina, Department of Biology and Marine Biology, 601 S. College Rd., Wilmington, NC 28403, USA
[e] Landscape Logic, Marine Research Laboratories, Tasmanian Aquaculture and Fisheries Institute, Private Bag 49, Hobart TAS 7001, Australia
[f] Department of Earth Sciences and the Climate Change Institute, Bryand Global Sciences Center, University of Maine, Orono, ME 04469, USA
[g] National Institute of Polar Research, 10-3, Midori-machi, Tachikawa-shi, Tokyo, 190-8518, Japan
[h] Institute of Antarctic and Southern Ocean Studies, University of Tasmania, Tasmania, Australia
[i] Institute of Geology and Mineralogy, University of Cologne, Zuelpicher Str. 49a, D 50674 Cologne, Germany
[j] PAGES International Project Office, Zähringerstrasse 25, Bern, 3012, Switzerland
[k] Antarctic Climate and Ecosystems Cooperative Research Centre, Private Bag 80, Hobart, Tasmania, 7001, Australia
[l] National Centre for Antarctic & Ocean Research, Ministry of Earth Sciences, Headland Sada, Vasco-Da-Gama, Goa 403804, India
[m] Department of Polar Geography, Arctic and Antarctic Research Institute, Bering St., 38, St. Petersburg 199397, Russian Federation

ARTICLE INFO

Article history:
Received 5 March 2010
Accepted 27 October 2010
Available online 3 November 2010

Keywords:
Antarctica
Holocene
climate change
warm period
paleolimnology
marine geology

ABSTRACT

We review the post-glacial climate variability along the East Antarctic coastline using terrestrial and shallow marine geological records and compare these reconstructions with data from elsewhere. Nearly all East Antarctic records show a near-synchronous Early Holocene climate optimum (11.5–9 ka BP), coinciding with the deglaciation of currently ice-free regions and the optimum recorded in Antarctic ice and marine sediment cores. Shallow marine and coastal terrestrial climate anomalies appear to be out of phase after the Early Holocene warm period, and show complex regional patterns, but an overall trend of cooling in the terrestrial records. A Mid to Late Holocene warm period is present in many East Antarctic lake and shallow coastal marine records. Although there are some differences in the regional timing of this warm period, it typically occurs somewhere between 4.7 and 1 ka BP, which overlaps with a similar optimum found in Antarctic Peninsula terrestrial records. The differences in the timing of these sometimes abrupt warm events in different records and regions points to a number of mechanisms that we have yet to identify. Nearly all records show a neoglacial cooling from 2 ka BP onwards. There is no evidence along the East Antarctic coastline for an equivalent to the Northern Hemisphere Medieval Warm Period and there is only weak circumstantial evidence in a few places for a cool event crudely equivalent in time to the Northern Hemisphere's Little Ice Age. There is a need for well-dated, high resolution climate records in coastal East Antarctica and particularly in Terre Adélie, Dronning Maud Land and Enderby Land to fully understand the regional climate anomalies, the disparity between marine and terrestrial records, and to determine the significance of the heterogeneous temperature trends being measured in the Antarctic today.

Contents

* Corresponding author. Tel.: +32 92648504; fax: +32 92648599.
E-mail address: elie.verleyen@UGent.be (E. Verleyen).
[1] These authors contributed equally to this paper.

0012-8252/$ – see front matter © 2010 Elsevier B.V. All rights reserved.
doi:10.1016/j.earscirev.2010.10.006

1. Introduction

Recent assessments of temperature change have revealed that warming in West Antarctica (WA) has exceeded 0.1 °C per decade during the past 50 yrs (Steig et al., 2009) and that the Antarctic Peninsula (AP) is one of the fastest warming regions on Earth (Vaughan et al., 2003). East Antarctica (EA) shows regional differences but the continent-wide average near-surface temperature trend is positive (Steig et al., 2009). To date little is known about how this pattern of recent warming in Antarctica relates to past natural variability in these regions, despite this being critical for improved prediction of the impact of future climate anomalies on the cryosphere and ecosystems of the continent (Hodgson et al., 2009a). Past climate change reconstructions in Antarctica are largely based on ice cores (e.g. Masson et al., 2000; Mayewski et al., 2009). These records have provided information on, for example, past global atmospheric composition (Petit et al., 1999; Jouzel et al., 2007), and productivity and iron flux in the Southern Ocean over several glacial–interglacial cycles (Wolff et al., 2006). In addition, ice cores have revealed the existence of a bipolar seesaw (Blunier et al., 1998; Broecker, 1998) with warm events being out of phase between the Northern and Southern Hemispheres during glacial periods (e.g. Stocker and Johnsen, 2003; Schneider and Steig, 2008; Barker et al., 2009) and likely also during interglacials (Masson-Delmotte et al., 2010a,b). Over the Pleistocene glacial–interglacial cycles, the climate of Antarctica has traditionally been considered to be largely controlled by changes in the Northern Hemisphere and particularly in the North Atlantic region including changes in the strength of deep water formation; yet recent studies have revealed that the Antarctic warming events preceded those in the North (Ahn and Brook, 2008; Mayewski et al., 2009). The mechanisms behind this forcing are still unclear, but a reduction in the stratification of the Southern Ocean and a subsequent release of CO_2 (Anderson et al., 2009; Hodgson and Sime, 2010) and a decrease in sea ice (Stevens and Keeling, 2000) are proposed as playing a critical role.

During the Holocene, the connection between climate changes in both hemispheres is however less clear both because the amplitude of the changes has been smaller (but see Masson-Delmotte et al., 2010a,b), and because the relative impact of regional driving or amplifying mechanisms has been greater. For example, Holocene glacier dynamics in New Zealand have been reported as neither in phase nor strictly anti-phased with glacier changes in both hemispheres (Schaefer et al., 2009). To better understand the past and present regional differences and the links between the climates of the Northern and Southern hemispheres during the Holocene, there is a clear need for well-dated paleoclimate records, particularly from the high latitudes in the Southern Hemisphere. Ice cores have some disadvantages in this respect because the more subtle climate anomalies such as those occurring during the Holocene (e.g. Mayewski et al., 2004; Wanner et al., 2008) are generally less well resolved in the records from the high central plateau(Masson et al., 2000, but see Schneider and Steig, 2008). This pattern is due to the small variations in isotopic composition in these inland locations compared with those generally observed at coastal sites (Bromwich et al., 1998; Masson et al., 2000). Near the coast, the recovery of reliable ice cores enabling high resolution reconstructions is in turn difficult, because the ice sheet is often too dynamic. In coastal regions, valuable and often overlooked alternatives are lake sediment cores and shallow marine sediment records (see Hodgson et al., 2004; Hodgson and Smol, 2008 for a review).

Here we aim to synthesise lacustrine and coastal marine records of regional climate and environmental changes along the margin of the East Antarctic Ice Sheet (EAIS), including the Ross Sea region (RSR), and compare these observations with existing reviews of changes in the AP region (e.g., Ingólfsson et al., 1998; Hjort et al., 2003; Hodgson et al., 2004; Bentley et al., 2009) and with a variety of Antarctic ice core records (e.g., Masson et al., 2000; Steig et al., 2000; Wolff et al., 2006).

2. Study area and materials and methods

2.1. Physical settings of the study area

The most prominent feature of Antarctica is its ice sheet, which covers over 99% of the continent and is made up of three distinct morphological zones, consisting of the EAIS, the West Antarctic Ice Sheet (WAIS) and the glaciers of the AP. The Transantarctic Mountains, which run between Victoria Land and the Ronne and Filchner Ice Shelves (Fig. 1), separate the EAIS from the WAIS. The AP extends north from a line between the southern part of the Weddell Sea and a point on the mainland south of the George VI Ice Shelf. The EAIS comprises by far the largest part of the ice sheet and drains directly into the Southern Ocean between c. 15°W and 150°E. It also has several outlet glaciers that flow through the Trans-Antarctic Mountains, with some of them draining directly into the Ross Sea and Weddell Sea embayments via ice shelves, and a minority that terminate on land (e.g. Taylor Glacier in the McMurdo Dry Valleys).

The ice-free terrain in EA includes coastal ice-free areas, together with inland nunataks for example in the Transantarctic Mountains, and the Dronning Maud Land regions. It is mainly in the coastal ice-free areas that (paleo)lake sediments and biological, geological and geomorphological evidence is present providing opportunities to study past climate and environmental changes along the EA coastline. The near shore environment of most of these coastal oases is characterised by bays and fjords, many of which have also been sampled for their marine geological records. The main ice-free regions studied to date are Schirmacher Oasis, the islands and peninsulas in the Lützow Holm Bay region, Amery Oasis, Larsemann Hills, Rauer Islands, Vestfold Hills, Bunger Hills, Windmill Islands and the RSR (Fig. 1). In brief, the Schirmacher Oasis (70°46′ S–11°44′ E) is a c. 50 km² wide area, bounded by the EAIS to the South

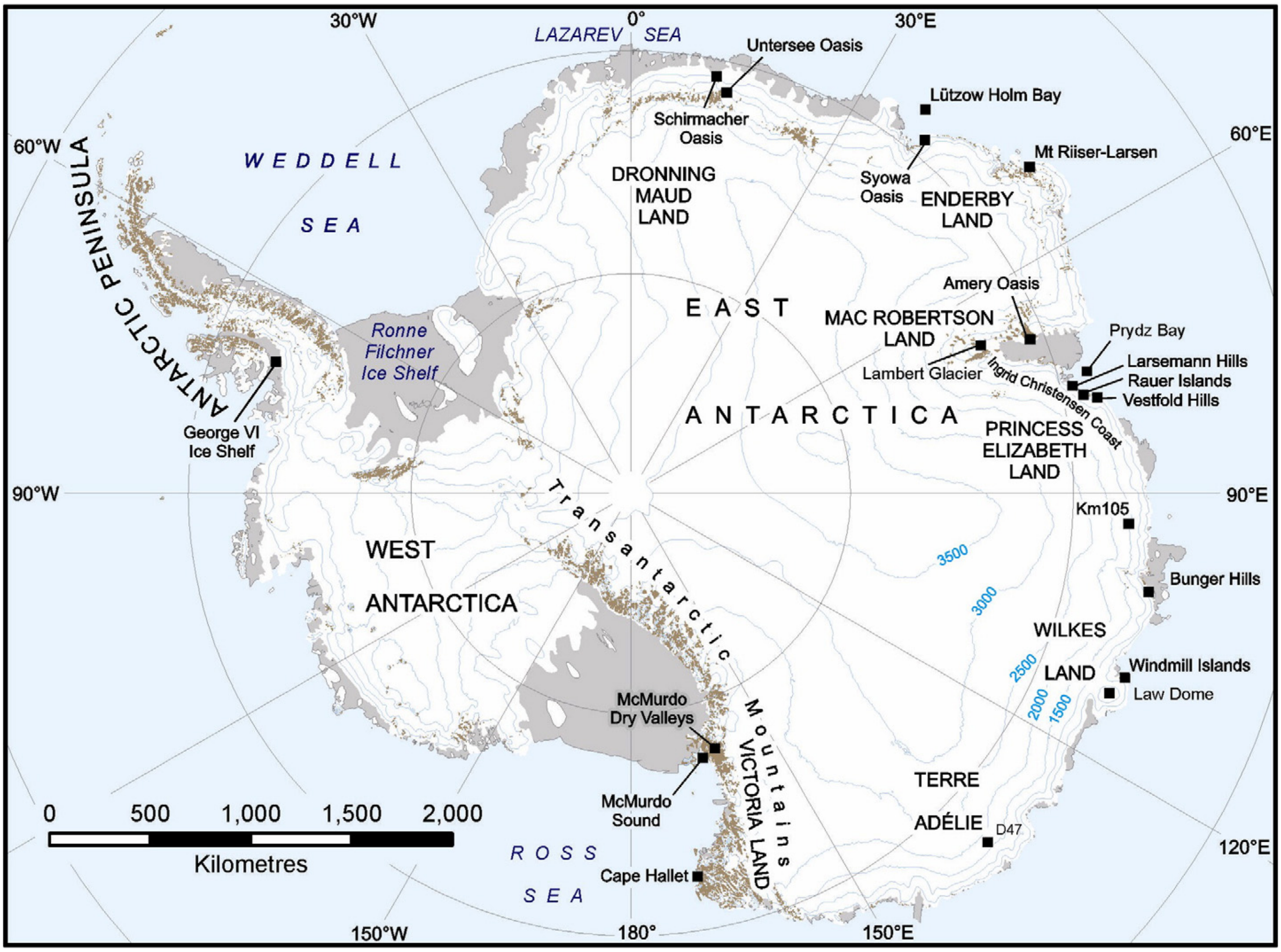

Fig. 1. Map of Antarctica showing the locations cited in the text.

and the Lazarev Ice Shelf to the North. Lützow Holm Bay includes several peninsulas and numerous ice-free islands that cover a total ice-free area of c. 200 km². Amery Oasis (70°40′ S–68°00′ E) is a 1800 km² area located some 200 km inland from the front of Amery Ice Shelf in Prydz Bay. The oasis is bordered by the Charybdis Glacier to the north, the EAIS to the west and southwest, and the drainage system of the Amery Ice Shelf/Lambert Glacier to the south and east. The Vestfold Hills (68°30′ S, 78°00′ E) are a 400 km^2 ice-free area on the Prydz Bay coast (Fig. 1), consisting of three main peninsulas (Mule, Broad and Long Peninsula) and a number of offshore islands. The Larsemann Hills (69°23′ S, 76°53′ E), is a 50 km^2 ice-free area on the Ingrid Christensen Coast located approximately midway between the eastern extremity of the Amery Ice Shelf and the southern boundary of the Vestfold Hills. The Rauer Islands (68°45′ S–68°55′ S and 77°30′ E–78°00′ E) are an ice-free coastal archipelago between the Larsemann Hills and the Vestfold Hills (Fig. 1). The Rauer Islands includes 10 major islands and promontories together with numerous minor islands covering a total area of some 300 km^2. The Bunger Hills (66°17′ S, 100°47′ E) is a 952 km^2 ice-free area of land and marine bays, bordered on the north by the Shackleton Ice Shelf and by the EAIS (with partly floating outlet glaciers) to the south, west and east. The Windmill Islands (66°20′ S–110° 30 E) are a group of ice-free islands and peninsulas on the EA coast north of Law Dome (Fig. 1) and cover a total area of 75–80 km². The RSR contains a number of ice-free coastlines, nunataks, and most notably the McMurdo Dry Valleys (77°28′ S–162°31′ E), which at 4800 km^2 are the largest relatively ice-free region in Antarctica.

2.2. Present day climate

During the instrumental period (past ~50 yrs) the mean annual temperature in the EA coastal region has been about −10 °C (Steig et al., 2009), with positive temperatures generally confined to the summer months, when lakes might become ice-free. During summer, the Antarctic receives more solar radiation than the tropics, but the snow and ice surfaces have a high albedo and reflect much of this radiation back into space (Turner et al., 2009). Where snow melts, exposing large patches of bare (dark) ground, or where sea ice melts exposing ice-free (dark) ocean, solar radiation is absorbed rather than reflected, and the environment will locally warm. During winter, the Antarctic stratosphere is extremely cold due to the lack of incoming radiation (Turner et al., 2009), which results in surface cooling, more extensive snow cover and complete lake ice cover in coastal regions. A strong temperature gradient then develops between Antarctica and the Southern Hemisphere mid-latitudes, thereby isolating a pool of very cold air above Antarctica. Very strong winds develop along this thermal gradient. During much of the year a strong temperature inversion exists above the surface, which is maintained by radiative cooling from the highly reflective snow and ice surface. However, when strong katabatic winds, transporting cold, relatively dense air from the interior of Antarctica, reach the coast, they can destroy the inversion through turbulent mixing, leading to enhanced sensible heat flux towards the surface and an increase in temperature. Thus, near-surface temperatures often co-vary strongly with wind speed. For Antarctica as a whole, studies using reanalysis data and climate models suggest that the season of greatest precipitation is autumn (Marshall, 2009) although there is significant regional variation across the continent (e.g. van de Berg et al., 2005). In the coastal regions, the majority of precipitation falls in relation to the passage of synoptic-scale weather systems and associated cloud bands within the circumpolar trough of low pressure that rings the continent (Marshall, 2009).

The primary mode of atmospheric circulation variability of the Southern high latitudes is the Southern Annular Mode (SAM), which has a significant influence on EA coastal temperatures (e.g. Thompson and Solomon, 2002; Marshall, 2007). In recent decades the relationship is predominantly negative ($r \sim -0.6$) so that a more positive SAM leads to colder regional temperatures. Using model data, Van den Broeke and van Lipzig (2003) demonstrated that the negative relationship between the SAM and regional temperatures is a combination of free atmosphere and near-surface effects. A more positive SAM, which is associated with stronger circumpolar westerlies around Antarctica, isolates the continent more effectively from the advection of warm extra-tropical air masses into the coastal region. In addition, positive SAM reduces the speed of katabatic flow over much of EA, hence leading to less disruption of the surface temperature inversion, a sensible heat flux away from the surface and colder temperatures (Van den Broeke and van Lipzig, 2003). There have been statistically significant positive increases in the SAM over the past few decades in austral summer (DJF) and autumn (MAM) (e.g. Marshall, 2003), which will have acted to cool EA temperatures. Nonetheless, over the past 50 yrs EA coastal temperatures have changed very little so that the changes in the SAM—believed to be principally due to anthropogenic forcing, through ozone depletion and increasing greenhouse gases—can be considered to be buffering the regional impact of global warming.

However, recent work by Marshall et al. (in press) has revealed that the sign of the relationship between the SAM and temperature in EA is not necessarily temporally stable and can indeed reverse on decadal timescales. The changes in sign are related to variability in the positions of the long-waves over the Southern Ocean, the non-annular component of the SAM. Changes in the relative strength and longitude of the two climatological lows located off the EA coast (at ~20°E and 110°E) will impact significantly on the mean advection of temperature and moisture into the region. Using ice core data encompassing the 20th century, Marshall et al. (in press) showed that such changes are part of natural climate variability. Thus, care needs to be taken in relating proxy indices of the SAM to regional temperature and vice versa.

2.3. Materials and methods

The majority of the past climate and environmental change reconstructions in EA reviewed here are based upon lake or shallow near shore marine sediment cores. We are aware that some proxies are more reliable than others in specific settings. Similarly some proxies, such as the occupation of land by biota, might also show a lagged response to climate changes, whereas others respond more quickly (e.g. the response of lacustrine diatom assemblages to salinity changes). Moreover, in contrast to marine sediment and ice cores in which alkenones and stable isotopes of H and O can be used as paleothermometers, most proxies used in terrestrial environments provide indirect measurements of past climate changes, for example by reconstructing changes in the moisture balance which can be a function of both temperature and wind speed. In some studies, dating uncertainty can also be high, particularly where only a small number of ^{14}C dates have been used to constrain the timing of past climate and environmental changes. Combined, these shortcomings must be taken into account when formulating a consensus on past regional climate changes.

For an overview of the different proxies used here, the reader is referred to Hodgson et al. (2004) and Hodgson and Smol (2008). In short, lacustrine diatoms are generally used in combination with transfer functions to enable the quantitative reconstruction of past changes in the moisture balance (e.g., Roberts and McMinn, 1998; Verleyen et al., 2003) and/or nutrient levels (Roberts et al., 2004). Marine diatoms are divided into sea ice related and open water taxa (e.g., Verleyen et al., 2004a) based upon their presence in modern analogue assemblages of the Southern Ocean (e.g., Armand et al., 2005; Crosta et al., 2005). Faunal microfossils are used to study past changes in lake ecology (Cromer et al., 2005, 2006). Fossil pigments enable the reconstruction of changes in the marine and lacustrine autotrophic communities, primary productivity, and changes in ice cover dynamics (e.g., Hodgson et al., 2003; Verleyen et al., 2004a, 2005). Sedimentological and geochemical proxies are used to reconstruct transitions between different sedimentary environments (e.g., glacial vs. lacustrine), primary productivity [e.g., loss-on-ignition, total organic carbon content (TOC)], whether a reduction or oxidation occurs in the water column through reduced or increased ice cover [total sulphur content (TS) e.g., Wagner et al., 2006], or the presence of vegetation in the catchment area (the total carbon vs. total nitrogen content (TC/TN), Wagner et al., 2006). Oxygen and carbon isotopes can be used to infer changes in the moisture balance in closed lakes (e.g., Hodgson et al., 2005).

Other evidence for past climate and environmental changes in EA include the presence of relict deltas on land formed at times of higher lake levels indicating increased meltwater supply (Hall and Denton, 2000a,b) and the presence/absence of seal hairs on land and penguin remains (e.g., Hall et al., 2006; Emslie et al., 2007). For example, the presence of hairs from Southern Elephant Seals has been used to infer changes in sea ice extent in the RSR (Hall et al., 2006), and the remains of past Adélie Penguin colonies and fossilized otoliths of their prey, Antarctic silverfish (Lorenzini et al., 2009), indicate that pack ice was present providing foraging opportunities for penguins during spring; conversely, their absence indicates that cold episodes caused unfavourable marine conditions with permanent sea ice blocking access to nesting sites (Emslie et al., 2007).

Where radiocarbon dating is used to provide a chronology, calibrated ^{14}C dates are cited as reported in the original publications. Similarly, corrections for the radiocarbon reservoir effect are applied as in the original publications. To allow comparison between the studies where ^{14}C dates were used but not calibrated, we list the original ^{14}C dates (^{14}C ka BP), and then carried out our own calibration using CALIB 5.0.2 (http://calib.qub.ac.uk/calib/) citing the upper and lower limits (at 2-std deviations) of the calendar ages (ka BP). In cases where dates were derived from marine organisms, the radiocarbon dates were corrected for the marine reservoir effect by subtracting 1300 yrs following the Antarctic standard (Berkman et al., 1998) prior to calibration (i.e., the offset from the global marine reservoir (ΔR) was set at 900 yrs when using the marine calibration curve; Hughen et al., 2004). For lacustrine ^{14}C ages younger than 11 ka BP the Southern Hemisphere atmospheric calibration curve was used (McCormac et al., 2004); in all other cases the Northern Hemisphere atmospheric calibration curve (Reimer et al., 2004) was applied.

The time of deglaciation of the current ice-free regions was derived from published cosmogenic isotope dating of exposed rocks (Gosse and Phillips, 2001) and ^{14}C dating of fossils incorporated into raised beaches, organic material and fossils in lake sediments, and (abandoned) bird colonies (e.g. through the radiocarbon dating of stomach oil deposits from snow petrels (*Pagodroma nivea*)). The ^{14}C dates are interpreted as minimum ages since there is an unknown lag time between deglaciation and colonization of the land by biota (e.g., Gore, 1997; Ingólfsson et al., 2003).

3. Overview of past climate changes

3.1. The Pleistocene–Holocene transition

The widespread Antarctic Early Holocene climate optimum between 11.5 and 9 ka BP observed in all ice cores from coastal and continental sites (Masson et al., 2000; Steig et al., 2000; Masson-Delmotte et al., 2004, 2010a; Mayewski et al., 2009), coincided with the onset of biogenic sedimentation in lakes and the occupation of ice-free land by biota between c. 13.5 and 10 ka BP in Princess Elizabeth Land and Mac Robertson Land (Figs. 1 and 2). In the Larsemann Hills, Vestfold Hills and Amery Oasis some areas escaped full glaciation during the Last Glacial Maximum (LGM) (Hodgson et al., 2001; Fink et al., 2006; Gibson et al., 2009; Kiernan et al., 2009; Colhoun et al., 2010; Newman et al. unpubl. res.), whereas other areas were likely completely ice-covered and gradually deglaciated between c. 13.5 and 4 ka BP (Hodgson et al., 2001; White et al., 2009). Diatoms and pigment data point to the establishment of seasonally melting lake ice and snow cover, and the development of microbial mats at c. 10.8 ka BP (Hodgson et al., 2005), with evidence for relatively wet conditions between c. 11.5 and 9.5 ka BP in a lake on one of the northern islands in the Larsemann Hills (Verleyen et al., 2004b). This is consistent with paleolimnological evidence from the nearby Vestfold Hills, where some lakes became ice-free and diatoms and rotifers inhabited water bodies from c. 11.4 ^{14}C ka BP onwards (c. 13.2–13.4 ka BP; Roberts and McMinn, 1999; Cromer et al., 2005). In the Rauer Islands ice-free conditions were inferred prior to c. 11.6–11.2 ka BP and a marine climate optimum is present between 9.2 and 8.2 ka BP (Berg et al., 2009, 2010). At least parts of Amery Oasis were covered by locally expanded glaciers during the Late Pleistocene (Hambrey et al., 2007), but deglaciation of some areas had started prior to c. 11 ka BP (Wagner et al., 2004; Fink et al., 2006) and biogenic sediments started to accumulate in some lakes from c. 13.4–12.5 ka BP (Wagner et al., 2004; Wagner et al., 2007; Newman et al. unpubl. res.), coincident with deglaciation of coastal regions in Mac Roberston Land from 13 ka BP onwards (Mackintosh et al., 2007). Microfossils indicate a well developed faunal community in one of the lakes from c. 13 ka BP (Newman et al. unpubl. res.), which was followed by the establishment of a diatom community at c. 10.2 ka BP, likely related to increased nutrient inputs and a reduction in ice and snow cover, marking the start of relatively warm conditions (Wagner et al., 2004).

In Wilkes Land, parts of the Bunger Hills remained ice-free during the LGM (Gore et al., 2001), whereas the Windmill Islands (66°20'S–110°30'E) were glaciated (Goodwin, 1993). The minimum ages for deglaciation of the Windmill Islands are also slightly younger than those from the oases near the Lambert Glacier; post-glacial lake sediments accumulated at c. 10.2 ka BP (Roberts et al., 2004), biogenic sedimentation in the marine bays started around c. 10.5 ka BP (Cremer et al., 2003; Hodgson et al., 2003), and penguins occupied the region from at least c. 9 ka BP (Emslie and Woehler, 2005). Despite the temperature optimum recorded in the ice cores, relatively cool summer conditions persisted near the Windmill Islands during the Early Holocene, probably due to the influence of local glaciers as reflected by the well developed sea ice diatom assemblages seen in coastal marine sediment cores (Cremer et al., 2003). The Bunger Hills were occupied by nesting snow petrels from at least 10 ka (Verkulich

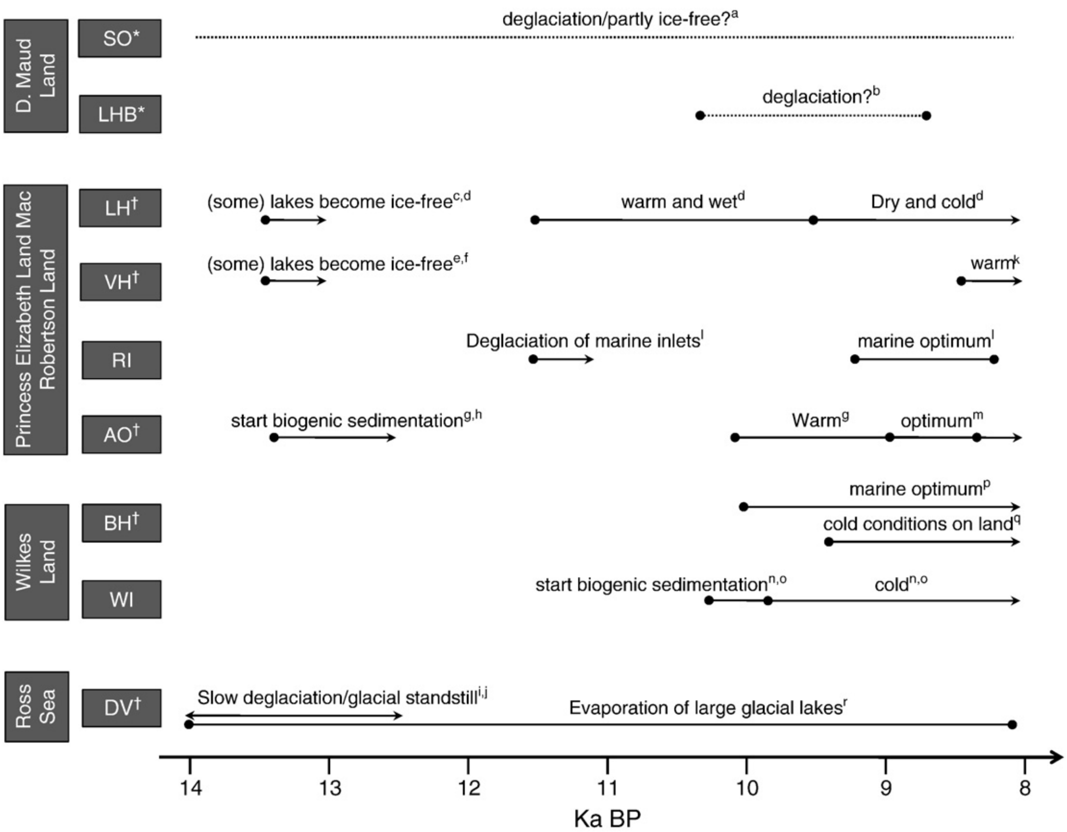

Fig. 2. Deglaciation and early Holocene warmth in ice-free regions along the East Antarctic coastline. Dashed lines represent poorly dated records. For regions indicated with an asterisk, detailed records are lacking. Regions indicated with a cross were partly ice-free during the LGM. SO: Schirmacher Oasis, LHB: Lützow Holm Bay region, LH: Larsemann Hills, VH: Vestfold Hills, RI: Rauer Islands, AO: Amery Oasis, BH: Bunger Hills, WI: Windmill Islands, and DV: McMurdo Dry Valleys. Superscript numbers refer to the original publications: [a]Gingele et al. (1997), [b]Miura et al. (1998), [c]Hodgson et al. (2005), [d]Verleyen et al. (2004b), [e]Roberts and McMinn (1999), [f]Cromer et al. (2005), [g]Wagner et al. (2004), [h]Fink et al. (2006), [i]Emslie et al. (2007), [j]Hall (2009), [k]Gibson (unpublished results), [l]Berg et al. (2010), [m]Cremer et al. (2007), [n]Cremer et al. (2003), [o]Hodgson et al. (2003), [p]Kulbe et al. (2001), [q]Verkulich et al., 2002, [r]Hendy (2000).

and Hiller, 1994), and organic sediments started to accumulate in the lakes at the Pleistocene–Holocene boundary in association with extensive and relatively rapid ice melting, which also supports the occurrence of an Early Holocene warm period at c. 9 ± 0.5 ka BP (Verkulich et al., 2002). Further radiocarbon evidence suggests that large parts of the southern Bunger Hills were rapidly deglaciated prior to 8 ka BP (Melles et al., 1997).

Although terrestrial climate archives are present in the ice-free regions in Dronning Maud, Enderby, and Mac Robertson Land (Sinha et al., 2000; Bera, 2004; Singh and Tiwari, 2004; Matsumoto et al., 2006), surprisingly little information is available about the deglaciation and post-glacial climate evolution there. At least some areas such as the Untersee Oasis (71°S–13°E) were probably ice-free during the LGM as shown by ^{14}C dating of stomach oil deposits from snow petrels, which indicates occupation during at least the past 34 ka (Hiller et al., 1988). Similarly, coastlines of some islands and peninsulas in Lützow-Holm Bay near Syowa Station (69°00′S–39°35′E) may have been ice-free for at least 40 ka and probably longer, as evidenced by AMS ^{14}C dates of individual *in situ* marine fossils incorporated into raised beaches (Miura et al., 1998). Other areas such as the Schirmacher Oasis likely deglaciated at the Pleistocene–Holocene boundary as indicated by a grounding line retreat in the Lazarev Sea (Gingele et al., 1997). Similarly, parts of Mt. Riiser-Larsen in Enderby Land likely deglaciated at the beginning of the Holocene as evidenced by the onset of organic sedimentation in Lake Richardson (Zwartz et al., 1998).

In the RSR, a grounded ice sheet fed from the glaciers of the Ross Embayment began filling the McMurdo Sound at ~27 ^{14}C ka BP (Emslie et al., 2007) until reaching its northern extent at the LGM, and remained at its LGM position until c. 12.7 ^{14}C ka BP (c. 14.7–15.2 ka BP). Following the LGM, the ice receded at a very slow rate, or stood still, until c. 10.8 ^{14}C ka BP (c. 12.7–12.9 ka BP) (Denton and Marchant, 2000; Hall and Denton, 2000a; see Hall, 2009 for a review). For example, the Wilson Piedmont Glacier probably maintained its LGM position until c. 10.7 ^{14}C ka BP (c. 12.6–12.8 ka BP) (Hall and Denton, 2000b; Hall et al., 2001) and persisted there into the Early Holocene (Hall et al., 2000; Hendy, 2000; Hall et al., 2001, 2002). This grounded ice sheet blocked several valleys in the McMurdo Dry Valleys and fed large proglacial lakes such as glacial Lakes Washburn and Wright, which had water over 500 m deep in some valleys (Hall and Denton, 1995) and existed until the Early Holocene (Hall et al., 2000; Hendy, 2000; Hall et al., 2002). Between c. 14 ka and 8 ka BP these large proglacial lakes started to evaporate (Hendy, 2000); the evaporation of Lake Washburn starting during the LGM. In the Taylor Valley, evaporation of Lake Fryxell was discontinuous, with a series of high and low stands (Wagner et al., 2006). Significant changes in the rate of carbonate precipitation (measured as Total Inorganic Carbon) in a sediment core from Lake Fryxell between c. 13 and 11 ka BP suggest a significant desiccation event at the Pleistocene–Holocene transition (Wagner et al., 2006, but see Whittaker et al., 2008). Offshore, an Early Holocene climate optimum with open marine conditions is inferred from the presence of varved diatomaceous ooze in a marine sediment core from Cape Hallett at >9.5–9.4 ka BP (Finocchiaro et al., 2005).

3.2. After the Early Holocene Warm period

The period following deglaciation shows complex and less consistent patterns than those observed at the Pleistocene–Holocene boundary (Fig. 3). In the Larsemann Hills, relatively dry conditions occurred on land between c. 9.5 and 7.4 ka BP, when lake levels dropped below their present position (Verleyen et al., 2004b). The composition of marine sediments deposited in isolation basins between c. 7.4 and 5.2 ka BP is consistent with a marine climate optimum, which is not clearly evident in the terrestrial sediments (Verleyen et al., 2004b; Hodgson et al., 2005). In the Amery Oasis, the slight warming inferred from a minor increase in lake organic matter content during the Early Holocene at 10.2 ka persisted until c. 6.7 ka BP, when a clear temperature optimum is inferred from geochemical proxies between c. 8.6 and 8.4 ka BP (Cremer et al., 2007). This warming was proposed as being in opposite phase (within dating uncertainty) to the 8.2 ka cold event reported from the North Atlantic region, yet recently obtained ^{14}C dates place this optimum earlier, between 8.9 and 8.7 ka BP (Newman et al. unpubl. res.), implying that the anomalies in both hemispheres are possibly not related. The optimum was followed by cold conditions from c. 6.7 to 3.7 ka BP (Wagner et al., 2004). In the Vestfold Hills, isostatic rebound and the emergence of isolation lakes from the sea resulted in major ecosystem changes, which hamper detailed paleoclimatological inferences being made for the period after the Early Holocene optimum, particularly in lower altitude lakes (Fulford-Smith and Sikes, 1996; Roberts and McMinn, 1999). However, in one of the lakes there is evidence for a warm period between c. 8.5 and 5.5 ka BP and a major cooling event between 5.5 and 5 ka BP (Gibson unpubl. res.). In the Rauer Islands, the cold conditions that started around 8.2 ka BP persisted until 5.7 ka BP were followed by a marine climate optimum which lasted until 3.5 ka BP (Berg et al., 2010). In the Windmill Islands, relatively cool summer conditions were observed in a marine bay with a combination of winter sea ice and seasonal open water conditions inferred from diatom assemblages between c. 10.5 and 4 ka BP (Cremer et al., 2003; Hodgson et al., 2003). These open water conditions are partly corroborated in the marine sediments of a nearby isolation basin but have less robust chronological constraints (Roberts et al., 2004). The peak of this (marine) cooling period was pinpointed at about c. 7 ka BP, when penguin colonies were abandoned on one of the peninsulas (Emslie and Woehler, 2005). In the Bunger Hills cold and dry conditions with limited meltwater and extended periods of lake ice cover prevailed on land between c. 9 and 5.5 ka BP (Verkulich et al., 2002), coincident with an expansion in snow petrel colonisation between c. 8 and 6 ka BP (Verkulich and Hiller, 1994). Ainley et al. (2006) link slightly elevated $\delta^{13}C$ values in stomach oil deposits from snow petrels to a shift in feeding to ^{13}C enriched prey which was made available by a reduction in sea ice during a marine climate optimum sometime between 7.5 and 5.5 ka BP, yet the temporal resolution of this record is particularly low during this period. Perhaps a better constraint on the timing of this marine optimum can be identified in coastal sediments where high Total Organic Carbon, high C/N and low concentrations of ice related diatoms were present from at least. 9.4 to 7.6 ka BP (Kulbe et al., 2001), followed by cold marine conditions between c. 7.6 and 4.5 ka BP.

In the RSR, the last remnants of grounded ice in Taylor Valley post-date c. 8.4 ^{14}C ka BP (c. 9–9.4 ka BP), and penguins did not recolonize the area until c. 8 ka BP, after an absence of c. 19,000 yrs (Baroni and Orombelli, 1994; Hall et al., 2006; Emslie et al., 2007). Relict lake deltas show an increase in the meltwater supply and increases in lake levels at c. 6 ^{14}C ka BP (c. 6.8 ka BP; Hall and Denton, 2000a,b) coinciding with a major retreat of the Ross Ice Sheet and a regional deglaciation inferred from radiocarbon dating of remains in raised beaches, till deltas, and marine sediments (c. 6.6 ^{14}C ka BP: c. 7.5 ka BP; Hall et al., 2004; Hall and Denton, 1999, 2000a). The recession of parts of the Wilson Piedmont Glacier was delayed until an ice shelf retreat in the area at c. 5.7 ^{14}C ka BP (c. 6.6–6.3 ka BP; Hall and Denton, 1999) and this retreat lasted until at least c. 4.4–3.1 ^{14}C ka BP (c. 5.2–4.8–3.5–3.3 ka BP; Hall and Denton, 2000b). This post-dates a secondary climate optimum detected in the ice cores of the RSR, between 8 and 6 ka BP (Masson et al., 2000, p 355). There is also sedimentological and geochemical evidence for increases in inferred salinity between c. 9 and 4 ka BP in a sediment core from Lake Fryxell (Wagner et al., 2006); although subsequent analyses based on stratigraphy, mineralogy and diatom assemblages found generally stable Holocene lake levels after the initial drop in the Early Holocene, with only a minor low-stand at ~6.4 ka BP (Whittaker et al., 2008). In a marine sediment core from northern Victoria Land, there is evidence

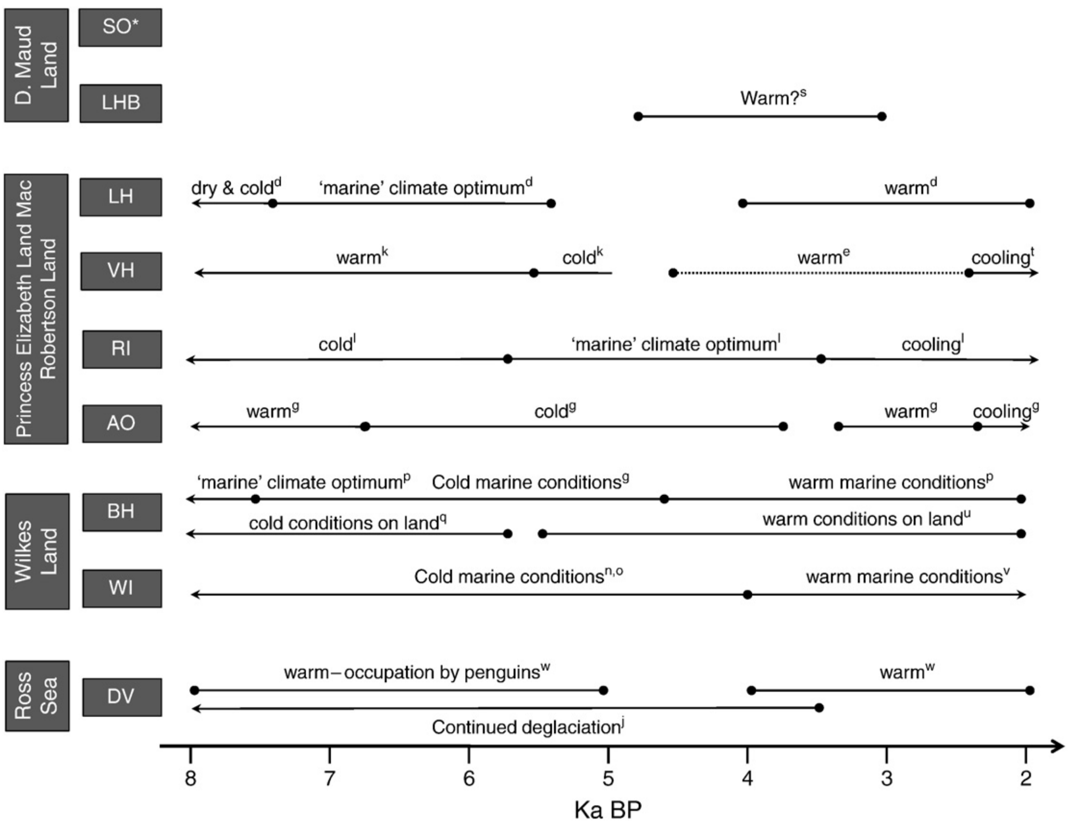

Fig. 3. Mid Holocene climate variability along the East Antarctic coastline. Dashed lines represent poorly dated records. For regions indicated with an asterisk, detailed records are lacking. Abbreviations as in Fig. 2. Superscript numbers refer to the original publications: for a–r see caption of Fig. 2; [s]Okuno et al. (2007), [t]Fulford-Smith and Sikes (1996), [u]Melles et al. (1997), [v]Kirkup et al. (2002).

for cooling from 9.4 ka and an abrupt decrease in palaeoproduction markers from 8.0 to 7.8 ka BP when sandy mud sediment suggests a rapid landward recession of the local/regional glaciers, together with the onset of seasonal sea ice formation (Finocchiaro et al., 2005).

3.3. The Mid to Late Holocene warm period

A Mid to Late Holocene warm period is inferred from various ice, lake and marine core records from Antarctica (see Hodgson et al. (2004) for a review). In the Larsemann Hills this warm period is dated between c. 4 and 2 ka BP (Fig. 3). There, relatively wet conditions resulted in increased lake water levels, which predate then overlap with the coastal marine optimum inferred from the presence of open water marine diatoms in isolation basins at 2.7 to 2.2 ka BP (Verleyen et al., 2004a,b). A short return to dry conditions and low water levels is present in one of the lake records at c. 3.2 ka BP (Verleyen et al., 2004b). The relatively wet period is coincident with ice thinning over Progress Lake and the resumption of biogenic sedimentation at 3.5 ka BP, after at least 40 ka of permanent ice cover (Hodgson et al., 2006a), and the formation of proglacial lakes on Stornes, the eastern of the two main peninsulas in the Larsemann Hills between c. 3.8 and 1.4 ^{14}C ka BP (c. 4.4–4.1 and 1.3 ka BP; Hodgson et al., 2001). In the Amery Oasis, relatively warm conditions between c. 3.2 and 2.3 ka BP are inferred from abundant organic matter deposition in Lake Terrasovoje (Wagner et al., 2004), yet newly obtained ^{14}C dates place this optimum slightly earlier at c. 3.5 and 3.1 ka BP (Newman et al. unpubl. res.). In the Vestfold Hills, a decline in lake salinity could be inferred between c. 4.2 and 2.2 ka BP, but the dates are at low resolution (Roberts and McMinn, 1996, 1999). This period of low salinity is however broadly supported by the warm and humid conditions between c. 4.7 and 3 ka BP proposed by Björck et al. (1996) after reinterpretation of previously published results (Pickard et al., 1986) and with reduced sea ice cover in one of the fjords between >3.5 and 2.5 ka BP (McMinn et al., 2001). Following this, a large increase in sea ice in the fjords suggests cooling occurred between 2.5 and 2 ka BP (McMinn et al., 2001).

In the Windmill Islands, enhanced biological production, probably reflecting more open water conditions and a climate optimum, occurred in the near shore environment between c. 4 and 1 ka BP (Kirkup et al., 2002), with stratified conditions caused by enhanced meltwater input (Cremer et al., 2003). In this area the Mid to Late Holocene warm period coincided with the readvance of the Law Dome ice margin after c. 4 ka BP, in response to an increase in precipitation (Goodwin, 1996). In the Bunger Hills, a stepwise increase in primary production was reported in the lakes between c. 4.7 and 2 ka BP (Melles et al., 1997) and was accompanied by the draining of some ice-dammed lakes between c. 5.5 and 2 ka BP (Verkulich et al., 2002). Similarly a gradual near shore marine warming was inferred for this area from c. 4.5 ka BP, followed by an optimum between c. 3.5 and 2.5 ka BP (Kulbe et al., 2001).

In the Lützow-Holm Bay region, a rapid isostatic rebound (6 m in c. 1000 yrs) occurred between c. 4.7 and 3 ka BP, which was linked to the rapid removal of part of the regional ice mass, most likely as a result of ice melting caused by warming (Okuno et al., 2007). No detailed information is available for other areas in the Dronning Maud Land region.

In the RSR, the presence of hairs from Southern Elephant Seals along with the remains of Adélie Penguins between c. 6 and 4 ^{14}C ka BP (c. 5.6–2.8 ka BP), indicates less sea ice than today, but sufficient pack ice for penguins to forage during spring according to Hall et al. (2006), but it is not until 4–2 ka BP that there was extensive occupation by Adélie Penguins along the central to southern coastlines (Baroni and Orombelli, 1994; Emslie et al., 2007) after a period of abandonment between 5 and 4 ka BP (Emslie et al., 2007). Moreover, the temporal distribution of otoliths from Antarctic silverfish, which is the preferred prey of Adélie penguins, shows a peak between 4 and 2 ka BP (Lorenzini et al., 2009). In Lake Fryxell, well developed microbial mats occurred from c. 4 ka BP onwards, which indicates similar environmental conditions and water depths to those found today (Wagner et al., 2006) although minor low stands have been inferred at 4.7, 3.8, and 1.6 ka BP (Whittaker et al., 2008).

3.4. Past 2000 yrs—neoglacial cooling, the Medieval Warm Period, the Little Ice Age and recent climate change

Much attention has been paid to the fluctuations in climate which in the Northern Hemisphere gave rise to the well-documented Medieval Warm Period (900–1300 AD or 1100–700 BP), and the Little Ice Age (LIA) (1450–1850 AD or 500 and 100 BP). The search in Antarctica for these climate signals is an important element in understanding how the Earth's climate system works, and in particular to determine if Late Holocene climate events are in phase or exhibit an anti-phase or seesaw pattern between the hemispheres.

In the Larsemann Hills, there is some evidence of neoglacial cooling (Fig. 4) leading to dry conditions around 2 ka BP (Hodgson et al., 2005), 700 yr BP and between c. 300 and 150 yr BP in some of the lakes (Verleyen et al., 2004b). These dry conditions parallel declines in sea bird populations during the past 2000 yrs which have been related to a 'climate deterioration' (Liu et al., 2007). In the Amery Oasis, neoglacial cooling occurred from c. 2.3 ka BP onwards, with a short return to a relatively warmer climate between c. 1.5 and 1 ka BP (Wagner et al., 2004). In the Vestfold Hills, meltwater input into the lakes gradually decreased from 3 ka BP onwards (Fulford-Smith and Sikes, 1996) and an increase in fast ice extent is observed from c. 1.7 ka BP (McMinn, 2000), broadly coincident with the Chelnock Glaciation on land (Adamson and Pickard, 1986). Cold conditions are similarly inferred between c. 1.3 ka BP and 250 yr BP (Bronge, 1992) and low precipitation from c. 1.5 ka ^{14}C BP (c. 1.3–1.5 ka BP, but dating is at low resolution; Roberts and McMinn, 1999). A palaeohydrological model reconstructing changes in lake salinity and water level from a lake sediment core shows no significant change in evaporation for the last c. 700 yrs, other than a slightly lower evaporation period at c. 150–200 yr BP, which is the only hint of a mild LIA-like event in the Vestfold Hills (Roberts et al., 2001).

In the Windmill Islands neoglacial cooling and persistent sea ice cover were inferred from a decrease in the biogeochemical proxies of production in the marine bays near the Windmill Islands (Kirkup et al., 2002; Cremer et al., 2003). A slow decrease in lake water salinity was observed on nearby islands during the Late Holocene, and there is no evidence for a LIA-like event there (Hodgson et al., 2006b). Instead, a very rapid salinity rise during the past few decades was present that has likely been brought about by increased wind speed and evaporation (Hodgson et al., 2006b). In the Bunger Hills, extended settlement by snow petrels after c. 2 ka BP was reported by Verkulich and Hiller (1994) from the presence of stomach oil deposits, and coincides with a lake sediment inferred climate cooling (Melles et al., 1997; Verkulich et al., 2002) and geomorphological evidence of a glacier readvance during recent centuries (Adamson and Colhoun, 1992). Cool conditions persisted from c. 2 ka BP with some minor fluctuations (Verkulich et al., 2002).

In the Lützow-Holm Bay region, lake sediment cores have yet to provide insights in past climate variability during the Late Holocene but preliminary investigations show excellent preservation of biogeochemical markers (Matsumoto et al., 2006). Detailed paleoclimatic records are still lacking for the region and for areas in Dronning Maud Land and other regions in Enderby Land.

In the RSR, the warmest period of the past 6000 yrs was inferred between c. 2.3 and 1.1 ka ^{14}C BP (c. 2.6–2.3 and 1.2–0.9 ka BP), based

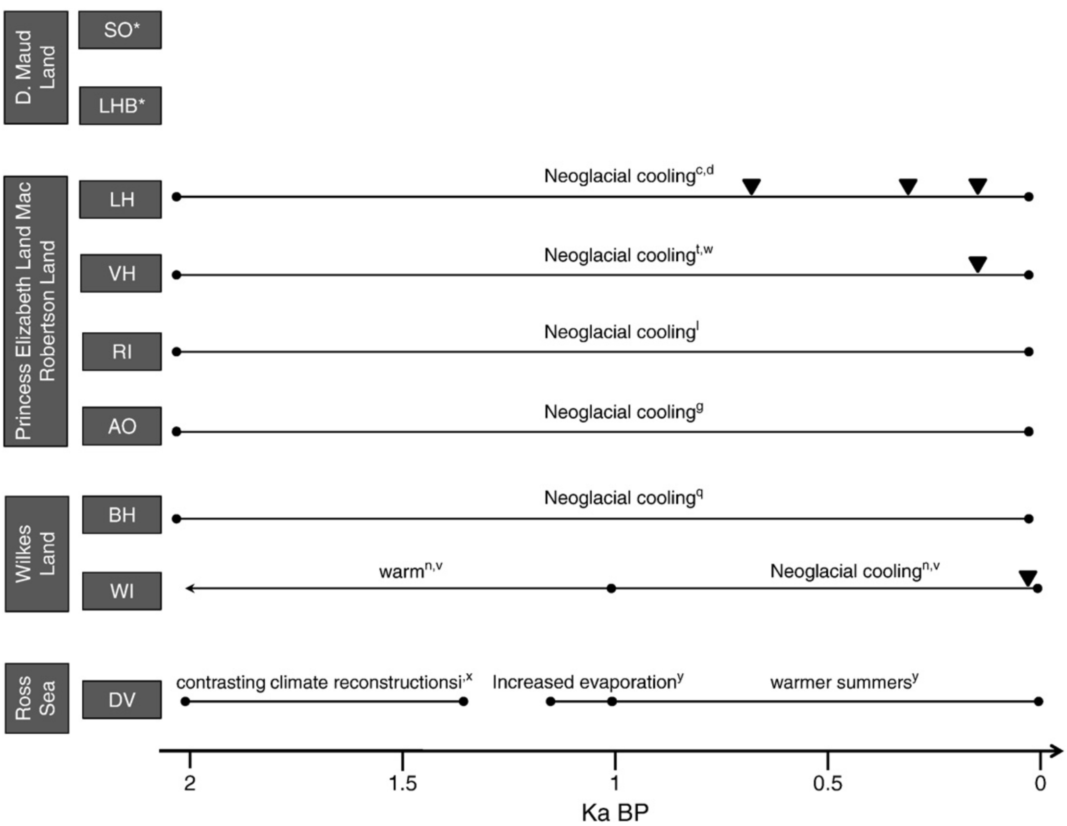

Fig. 4. Climate variability along the East Antarctic coastline during the past 2000 yrs. For regions indicated with an asterisk, detailed records are lacking. Short dry or cold periods are indicated with a triangle. Abbreviations as in Fig. 2. Superscript numbers refer to the original publication: for a–w see captions of Figs. 2 and 3; [w]McMinn (2000), [x]Hall et al. (2006), [y]Whittaker et al. (2008).

upon an expansion in elephant seal colonies which was linked to a decline in sea ice extent (Hall et al., 2006). However Emslie et al. (2007) conclude that Adélie penguins were absent from the Scott Coast and Ross Island between 2 and 1.1 ka BP, likely correlated with cooling episodes and extensive sea ice that caused unfavorable marine conditions for breeding in the southern Ross Sea, whilst Hall et al. (2006) do not infer there widespread development of land fast ice until after 1.2–0.9 ka BP which led to the abandonment of the elephant seal colonies. These differences in the climate inferences derived from elephant seal versus penguin data have yet to be resolved. Inland enhanced evaporation resulted in Lakes Fryxell, Vanda and Bonney becoming ice-free hypersaline ponds by c. 1.2–1 ka BP (Lyons et al., 1998; Wagner et al., 2006), and at the same time Lake Wilson, a perennially ice-capped, deep (>100 m) lake further South (80°S) in Southern Victoria Land, evaporated to a brine lake (Webster et al., 1996). After 1 ka BP increasing water levels and primary production in the lakes (Lyons et al., 1998; Wagner et al., 2006; Whittaker et al., 2008) are attributed to higher summer temperatures or to an increase in the number of clear, calm and snowless midsummer days (Hendy, 2000). During the last few centuries (< c. 0.2 ^{14}C ka BP, c. 0.4–0.1 ka BP), the Wilson Piedmont Glacier has been more extensive than today as documented by aerial photographs taken in 1956 (Hall and Denton, 2002) but, like other areas of EA there is no clear sign in the McMurdo Dry Valleys of the magnitude and style of glacier advances typical of the Northern Hemisphere Little Ice Age (Hall and Denton, 2002).

Studies in the framework of the US Long Term Ecological Research program have revealed a rapid ecosystem response to local climate cooling in the McMurdo Dry Valleys during recent decades, as evidenced by a decline in lake primary production and declining numbers of soil invertebrates (Doran et al., 2002). However, inter-annual fluctuations are large, as evidenced by the occurrence of an unusually warm summer in 2001–2002, which created sufficient meltwater to replace the prior 14 yrs of lake level lowering in a period of just three months (Doran et al., 2008).

4. Synthesis and discussion

Many of the larger currently ice-free regions studied, escaped full glaciation during the LGM, such as the Larsemann Hills (Hodgson et al., 2001, 2009b), the Vestfold Hills (Gibson et al., 2009; Colhoun et al., 2010), Amery Oasis (Fink et al., 2006) and the Bunger Hills (Gore et al., 2001) and possibly also some island and peninsula coastlines in Lützow Holm Bay (Miura et al., 1998), whereas other regions were probably completely glaciated and became gradually ice-free after the Pleistocene–Holocene boundary, with differences in local ice mass and extent causing regional differences in the timing of deglaciation and subsequent colonisation by biota. The onset of deglaciation of the EA oases generally preceded the deglaciation of the AP (Bentley, 1999; Anderson et al., 2002; see Hall, 2009 for a review). However, some areas such as Stornes in the Larsemann Hills and parts of the Lützow Holm Bay region became ice-free as recently as the Mid to Late Holocene, probably as a result of local ice configurations, further climate warming and ongoing relative sea level changes. In the RSR, deglaciation started during the transition from the Pleistocene and continued into the Early Holocene forming a series of large proglacial lakes. These evaporated to approximately their present size as late as the Mid Holocene (Hall and Denton, 2000a,b; Wagner et al., 2006). This multi stage deglaciation pattern across EA is similar to that observed in the Northern Hemisphere, where the ice sheets started to melt immediately after the LGM and continued to disintegrate in various stages until the Late Holocene (Siegert, 2001). For example, the western part of the North American continent became ice-free during the Pleistocene to Holocene transition (Carlson et al., 2008), whereas the Laurentide Ice Sheet persisted over the land masses bordering the North Atlantic region during the Early Holocene. The mechanisms behind these differences in deglaciation patterns along the EA coastline are still unclear. Dating uncertainties are sometimes high which may (partly) underlie the differences in deglaciation history between the different regions. However, it is reasonable to conclude that the main phases of deglaciation were closely linked to the major climate shifts during Termination 1, and the widespread thermal maxima of the Early Holocene and Mid to Late Holocene with regional variations brought about by differences in the response of the ice sheets to internal glaciological dynamics, the volume of overriding ice at the LGM, global sea level change, and local climate forcing (cf. Hall, 2009).

The Early Holocene in EA is characterised by the first thermal optimum, which is well resolved in the majority of terrestrial and coastal marine records between c. 11.5 and 9.5 ka BP, centred on c. 10 ka BP. This climate optimum is seen in Antarctic ice cores between 11.5 and 9 ka BP (Masson et al., 2000; Stenni et al., 2010), and detected across the continent, for example in a marked phase of deglaciation of the AP Ice Sheet (e.g. Bentley, 1999; Hall, 2009), and thinning of the WA Ice Sheet in Marie Byrd Land (Johnson et al., 2008). Moreover, this Early Holocene optimum on land is also evident in most marine sediment cores (e.g., Finocchiaro et al., 2005; Crosta et al., 2007 and references therein) and coincides with the transition from grounded ice to open marine conditions in some regions of the EA continental shelf, such as in Prydz Bay (Domack et al., 1991; Taylor and McMinn, 2002) and Terre Adélie (Denis et al., 2009). The Early Holocene thermal optimum appears to be a nearly bipolar phenomenon as it is also detected in records from Beringia, Alaska and Northwest America (e.g. Kaufman et al., 2004). However, the North Atlantic region remained cold and didn't experience warmer conditions until c. 2.5 ka later at 7–6 ka BP, as inferred from pollen-based temperature reconstructions for northern Europe (e.g., Davis et al., 2003). The climate at high latitudes in parts of the Southern and Northern Hemisphere (except the North Atlantic region) thus likely responded in phase with the summer insolation maximum at the high latitudes in the Northern Hemisphere during the Early Holocene (Berger and Loutre, 1991). Interestingly, summer insolation values at the high latitudes in the Southern Hemisphere did not peak during this Early Holocene period (Berger and Loutre, 1991), but later during the Mid to Late Holocene; hence the Early Holocene climate optimum in Antarctica is likely related to other forcings such as the culmination of the deglacial rise in CO_2 changes in the position and strength of the Southern Westerlies leading to more negative SAM (e.g. Bentley et al., 2009) and/or changes in the thermohaline circulation (THC; Blunier et al., 1998). For the latter scenario geological evidence is still lacking, but it has been suggested that Antarctic interglacial climate optima are caused by transient heat transport redistribution comparable with glacial north–south seesaw abrupt climatic changes (Masson-Delmotte et al., 2010a,b). A possible scenario would then be that decreased North Atlantic Deep Water formation as a result of the final deglaciation in the Northern Hemisphere during the Early Holocene weakened the Atlantic Meridional Overturning Circulation (AMOC) and led to warmer conditions over Antarctica. This scenario would then be in agreement with the relatively cold conditions observed in the North Atlantic region (Kaufman et al., 2004).

The period following the Early Holocene optimum shows complex regional patterns with cold and dry conditions in some regions, coinciding with relatively warmer conditions elsewhere (Figs. 2 and 3), suggesting that regional rather than global forcing mechanisms dominated the climate. This is similar to the marine sediment cores from the continental shelf in which there is also no consensus yet regarding the early Holocene climate evolution, but most records there point to the presence of a climate optimum (see Crosta et al., 2008 and reference therein). This thermal optimum either started during the Early Holocene and lasted until the Mid or Late Holocene, or was only shortly interrupted by a cooling trend between 9 and 8 ka BP, after which warmer conditions were re-established until c. 5 ka BP in EA or c. 3 ka BP in West Antarctica (Leventer et al., 1996). For example, less sea ice cover was inferred in the Prydz Bay region between 11,650 and 2600 ^{14}C yr BP (Taylor and McMinn, 2002), which was possibly interrupted by a readvance of floating glacial ice between 7400 and 3300 ^{14}C yr BP as observed in the ODP 740A site (Domack et al., 1991). In Adélie Land, a cooler Early Holocene was followed by a Mid Holocene warm period (from 7.7 ka BP onwards) with a transition to colder, Neoglacial conditions at c. 4 ka BP (Crosta et al., 2007, 2008). In Victoria Land, the climate cooled from 9.4 ka BP onwards until 8 ka BP, after which seasonal open water conditions and circulation patterns as those of today were established (Finocchiaro et al., 2005). In summary, a marine inferred climate optimum in some areas is apparently out of phase with terrestrially inferred temperatures, or even coincident with cool and dry conditions on land. These anti-phased pattern in climate variability and the disparity between records might reflect (i) differences in heat capacity and hence thermal inertia between the different systems, which is typically higher for the ocean compared with the terrestrial realm (Renssen et al., 2005) and the fact that (ii) the organic fraction in marine sediments reflects spring, summer and autumn conditions (including sea ice blooms), whereas lacustrine biotic assemblages largely reflect summer conditions when the lakes are ice-free and primary production peaks (Hodgson and Smol, 2008), and (iii) solar insolation maxima during spring, summer and autumn are out of phase at the high latitudes in the Southern Hemisphere during the Holocene (Berger and Loutre, 1991; Bentley et al., 2009). For example, solar insolation was at a maximum during spring at the beginning of the Holocene, whereas the summer insolation was relatively high during the Mid to Late Holocene (Renssen et al., 2005). In addition dating uncertainties associated with variable marine reservoir effects in near shore coastal records may also hamper comparison between the different records and anomalies, together with lags in response time of the different archives and proxies used.

In the Amery Oasis there is evidence of a climate optimum between c. 8.6 and 8.4 ka BP (Cremer et al., 2007) or between 8.9 and 8.7 ka BP (Newman et al. unpubl. res.), which slightly predates the colonisation of the RSR by penguins at c. 8 ka BP after an absence of c. 19,000 yrs (Hall et al., 2006; Emslie et al., 2007). Interestingly, this climate optimum coincides with a cooling trend observed in ice cores (Stenni et al., 2010) and an open marine record from the South Atlantic Polar Frontal Zone between 9 and 8 ka BP (Nielsen et al., 2004), and, although dating uncertainties are large, it also predates or overlaps (within the dating error) with the well-known temperature decline around 8.2 ka BP in the ice core record from central Greenland (Alley et al., 1997) and between 8.5 and 8 ka BP in records outside the North Atlantic region (Rohling and Pälike, 2005). However, it remains to be seen whether there is a causal link (e.g., seesaw) between the expression of these climate anomalies at the high latitudes in the Northern and Southern hemispheres or if, as some models suggest, these anomalies (after 9 ka) can be explained by the combined effects of local orbital forcing and the long memory of the system (Renssen et al., 2005). Moreover, local processes in the terrestrial realm, such as isostatic uplift and deglaciation, might have had an overriding effect on the regional and global changes such as solar insolation and differences in the thermohaline circulation. Isostatic uplift is known to be high after deglaciation, which, together with glacier and ice sheet retreat, leads to an increase in the amount of ice-free land which was previously below sea level. This relative increase in ice-free ground likely had a negative effect on the local albedo and hence could have led to warmer conditions on land.

The Mid to Late Holocene in EA is characterised by the second thermal optimum, which is well resolved along most parts of the coastline but not well resolved in regional ice cores (although secondary warm periods are observed ca. 3 ka BP at Dome C, Dominion Range, and Byrd (Masson et al., 2000, p. 355)), and in marine sediment cores from the continental shelf (e.g. Crosta et al., 2007) and near the Antarctic Polar Front (between 50 and 53.2°S; Divine et al., 2010) in which a Neoglacial cooling is observed from 4 ka BP onwards. This Mid to Late Holocene warm period in EA can be roughly placed between 4.7 and 1 ka BP and is evidenced by open water coastal marine conditions, increased rates of deglaciation, and wet/warm conditions on land. It is not known if it occurs in coastal Dronning Maud Land, Enderby Land and Terre Adélie due to a lack of long-term high resolution records in these regions. Similarly, it is not well resolved in the RSR where ice cores point to an earlier secondary warm period at ca. 8–6 ka BP (Masson et al., 2000, p. 355) coincident with an optimum based on the occupation of the region by seals and penguins (Hall et al., 2006). The seal and penguin occupation data then either suggest a later optimum at c. 2.3 and 1.1 ka ^{14}C BP (c. 2.6–2.3 and 1.2–0.9 ka BP) (Hall et al., 2006), or cooling and persistent sea ice (Emslie et al., 2007); a difference in interpretation that requires further research. This is followed by an intensification of Southern circumpolar westerlies at c. 1.2–1 ka BP, accompanied by relatively cooler conditions in other regions along the EA coastline and in the interior of EA (Masson et al., 2000) and WA (Siple Dome; Mayewski et al., 2009), which is consistent with conditions experienced when the SAM is in its positive mode. In the AP (excluding the northernmost islands), this warm period occurs somewhere between c. 4 and 2 ka BP, whereas in northern sites of the AP it spanned 3.8 – 1.4 ka BP (see Hodgson et al., 2004 for a review). This Late Holocene climate optimum in terrestrial environments in Antarctica may be a result of low altitude coastal sites being closer to the 0 °C threshold which would have the effect of amplifying various climate proxies (e.g. Quayle et al., 2002), together with the occurrence of maxima in spring and summer solar radiation during this period (Bentley et al., 2009). Late spring and summer are of high importance in driving primary productivity and inter-annual environmental variability in polar terrestrial and lake ecosystems (Hodgson and Smol, 2008), hence solar insolation maxima during this part of the year are expected to have a profound effect on the records preserved on land. In the AP, an alternative explanation for this warm period is that a poleward displacement of the Southern Westerlies brought warm, moist air to the west side of the AP leading to higher temperatures and precipitation (Bentley et al., 2009). Finally, local processes might also underlie this terrestrial optimum. For example, in the Larsemann Hills, some lakes and likely also their catchments deglaciated during this period which likely led to changes in albedo and a regional warming. It is however unclear whether this deglaciation is the result of the climate optimum, or rather the trigger. This Antarctic Late Holocene climate optimum is apparently out of phase with anomalies observed in the Northern hemisphere (Mayewski et al., 2004) and in records from the continental shelf (e.g. Crosta et al., 2008). It is clear that there is an urgent need for well-dated records around the EA coastline in order to better compare them with the AP region and with records from the continental shelf and sub-Antarctic regions (e.g. Divine et al., 2010). In addition, these records are needed to study the influence of past climate variability on ecosystem functioning (Hodgson et al., 2004), because the Late Holocene warming event acts as one of the natural analogues for the climate change that parts of Antarctica are experiencing today.

After 2 ka BP, most areas of EA experienced neoglacial cooling, with markedly cooler or drier events in some regions. With the exception of some ice cores where the most recent cold anomaly is called the 'Antarctic Little Ice Age' (Masson et al., 2000, p354), there is little convincing evidence for a coeval Little Ice Age event (Roberts et al., 2001), nor for anything corresponding convincingly and region-wide to the Medieval Warm Period of the Northern Hemisphere (e.g., Mann et al., 2008).

In the last few decades, a rapid salinity increase has been recorded in lakes in the Windmill Islands region, and a number of ancient moss banks have become desiccated (Wasley et al., 2006) possibly in response to the increased wind speed observed in some EA coastal regions (Hodgson et al., 2006b). High resolution records from other areas in EA are however lacking. In the RSR, a re-assessment of temperature measurements has revealed that the continent-wide average near-surface temperature trend is positive (Steig et al., 2009). Also in WA, recent measurements and ice core data have revealed that surface temperatures are significantly rising (Schneider and Steig, 2008; Steig et al., 2009). The most vulnerable region in terms of global climate change is without doubt the AP, where temperatures are rising by ~0.55 °C per decade, which is six times the global mean (Vaughan et al., 2003). At present, the 'ozone hole' is actually buffering global warming in EA and when it closes, warming is predicted to accelerate there as well (Turner et al., 2009).

5. Conclusions

Well-dated terrestrial and shallow marine records provide valuable contributions to resolving the magnitude and geographical extent climate changes in EA. Several past climate anomalies appear to be in phase with changes at the higher latitudes in the Northern Hemisphere, whilst others are out of phase (or in anti-phase). There is clear evidence for a nearly Antarctic wide Early Holocene (11.5–9 ka BP) optimum in phase with changes at Northern Hemisphere high latitudes, but out of phase with changes in the North Atlantic region; and a Mid to Late Holocene (4–1 ka BP) optimum near the coast, subject to regional variations in its precise timing, but occurring well after the orbitally forced (Hewitt and Mitchell, 1998) summer Mid Holocene warm period experienced between 7 and 5 ka BP in the Northern Hemisphere high latitudes. There is little or no evidence for a coeval Little Ice Age and Medieval Warm Period in Antarctica. In the most recent period, there is evidence that global warming is starting to significantly impact some ecosystems in various regions in EA, for example through increased primary productivity in lakes, lake level lowering, increases in lake salinity and desiccation of moss banks. At the moment, there is a pressing need for well-dated records to study this present climate anomaly in relation to past natural variability, to reveal the likely ecological consequences and identify both similarities and differences in the respective forcing mechanisms involved. The paucity of data is particularly acute for coastal regions in the Dronning Maud and Enderby Land regions and from Terre Adélie. In the latter region, shallow marine sediments may be the only alternative due to the general lack of ice-free land containing lakes and geological evidence of past environmental changes.

Acknowledgements

This work was initiated by the author's contribution to the special SCAR Antarctic Climate Change and the Environment report. EV is a post-doctoral research fellow with the Fund for Scientific research, Flanders (Belgium). DH is funded by the Natural Environment Research Council (UK). IT is funded by the Institute for the Promotion of Science and Technology in Flanders (Belgium). DR and AM acknowledge the financial support of the Australian Research Council, MM and BW acknowledge the German Research Foundation and SMS acknowledges the National Indian Centre for Antarctic and Ocean Research and the Indian Ministry of Earth Sciences. This work benefited from the HOLANT project funded by the Belgian Federal Science Policy Office and was only possible with the logistic support of the Australian Antarctic Division, the British Antarctic Survey, the Japanese National Institute for Polar Research, Maaike Vancauwenberghe (BelSPO), the Office of Polar Programs of the U.S. National Science Foundation and the Programma Nazionale di Richerche in Antartide (Italy). Fig. 1 was prepared by Peter Fretwell, British Antarctic Survey. Two anonymous reviewers are thanked for their constructive comments on an earlier version of the manuscript.

References

Adamson, D.A., Colhoun, E.A., 1992. Late Quaternary glaciation and deglaciation of the Bunger Hills, Antarctica. Antarct. Sci. 4, 435–446.

Adamson, D.A., Pickard, J., 1986. Cainozoic history of the Vestfold Hills. In: Pickard, J. (Ed.), Antarctic Oasis. Academic Press, Sydney, pp. 63–97.

Ahn, J., Brook, E.J., 2008. Atmospheric CO_2 and climate on millennial time scales during the Last Glacial Period. Science 322, 83–85.

Ainley, D.G., Hobson, K.A., Crosta, X., Rau, G.H., Wassenaar, L.I., Augustinus, P.C., 2006. Holocene variation in the Antarctic coastal food web: linking δD and δ^{13}Cin snow petrel diet and marine sediments. Mar. Ecol. Prog. Ser. 306, 31–40.

Alley, R.B., Mayewski, P.A., Sowers, T., Stuiver, M., Taylor, K.C., Clark, P.U., 1997. Holocene climatic instability: a prominent, widespread event 8200 yr ago. Geology 25, 483–486.

Anderson, J.B., Shipp, S.S., Lowe, A.L., Wellner, J.S., Mosola, A.B., 2002. The Antarctic Ice Sheet during the Last Glacial Maximum and its subsequent retreat history: a review. Quatern. Sci. Rev. 21, 49–70.

Anderson, F.F., Ali, S., Bradtmiller, L.I., Nielsen, S.H.H., Fleisher, M.Q., Anderson, B.E., Burckle, L.H., 2009. Wind-driven upwelling in the Southern Ocean and the deglacial rise in atmospheric CO_2. Science 323, 1443–1448.

Armand, L.K., Crosta, X., Romero, O., Pichon, J.J., 2005. The biogeography of major diatom taxa in Southern Ocean sediments: 1. sea ice related species. Palaeogeogr. Palaeoclimatol. Palaeoecol. 223, 93–126.

Barker, S., Diz, P., Vautravers, M.J., Pike, J., Knorr, G., Hall, I.R., Broecker, W.S., 2009. Interhemispheric Atlantic seesaw response during the last deglaciation. Nature 457, 1097–1102.

Baroni, C., Orombelli, G., 1994. Abandoned penguin rookeries as Holocene paleoclimatic indicators in Antarctica. Geology 22, 23–26.

Bentley, M.J., 1999. Volume of Antarctic ice at the Last Glacial Maximum, and its impact on global sea level change. Quatern. Sci. Rev. 18, 1569–1595.

Bentley, M.J., Hodgson, D.A., Smith, J.A., Cofaigh, C.O., Domack, E.W., Larter, R.D., Roberts, S.J., Brachfeld, S., Leventer, A., Hjort, C., Hillenbrand, C.D., Evans, J., 2009. Mechanisms of Holocene palaeoenvironmental change in the Antarctic Peninsula region. Holocene 19, 51–69.

Bera, S.K., 2004. Late Holocene palaeo-winds and climatic changes in Eastern Antarctica as indicated by long-distance transported pollen–spores and local microbiota in polar lake core sediments. Curr. Sci. 86, 1485–1488.

Berg, S., Wagner, B., White, D., Cremer, H., Bennike, O., Melles, M., 2009. Late Pleistocene glaciation history of Rauer Group, East Antarctica. Antarct. Sci. 21, 299–300.

Berg, S., Wagner, B., Cremer, H., Leng, M.J., Melles, M., 2010. Late Quaternary environmental and climate history of Rauer Group, East Antarctica. Palaeogeogr. Palaeoclimatol. Palaeoecol. 297, 201–213.

Berger, A., Loutre, M.F., 1991. Insolation values for the climate of the last 10000000 years. Quatern. Sci. Rev. 10, 297–317.

Berkman, P.A., Andrews, J.T., Björck, S., Colhoun, E.A., Emslie, S.D., Goodwin, I.D., Hall, B.L., Hart, C.P., Hirakawa, K., Igarashi, A., Ingólfsson, O., Lopez-Martinez, J., Lyons, W.B., Mabin, M.C.G., Quilty, P.G., Taviani, M., Yoshida, Y., 1998. Circum-Antarctic coastal environmental shifts during the Late Quaternary reflected by emerged marine deposits. Antarct. Sci. 10, 345–362.

Björck, S., Hjort, C., Ingólfsson, Ó., Zale, R., Ising, J., 1996. Holocene deglaciation chronology from lake sediments. In: López-Martínez, J., Thomson, M.R.A., Thomson, J.W. (Eds.), Geomorphological map of Byers Peninsula, Livingston Island. : BAS GEOMAP Series Sheet, 5-A. British Antarctic Survey, Cambridge, pp. 49–51.

Blunier, T., Chappellaz, J., Schwander, J., Dallenbach, A., Stauffer, B., Stocker, T.F., Raynaud, D., Jouzel, J., Clausen, H.B., Hammer, C.U., Johnsen, S.J., 1998. Asynchrony of Antarctic and Greenland climate change during the Last Glacial period. Nature 394, 739–743.

Broecker, W.S., 1998. Paleocean circulation during the Last Deglaciation: a bipolar seesaw? Paleoceanography 13, 119–121.

Bromwich, D.H., Chen, B., Hines, K.M., 1998. Global atmospheric impacts induced by year-round open water adjacent to Antarctica. J. Geophys. Res. Atmos. 103, 11173–11189.

Bronge, C., 1992. Holocene climatic record from lacustrine sediments in a fresh-water lake in the Vestfold Hills, Antarctica. Geogr. Ann. Ser. A Phys. Geogr. 74, 47–58.

Carlson, A.E., Legrande, A.N., Oppo, D.W., Came, R.E., Schmidt, G.A., Anslow, F.S., Licciardi, J.M., Obbink, E.A., 2008. Rapid early Holocene deglaciation of the Laurentide Ice Sheet. Nat. Geosci. 1, 620–624.

Colhoun, E.A., Kiernan, K.W., McConnell, A., Quilty, P.G., Fink, D., Murray-Wallace, C.V., Whitehead, J., 2010. Late Pliocene age of glacial deposits at Heidemann Valley, East Antarctica: evidence for the last major glaciation in the Vestfold Hills. Antarct. Sci. 22, 53–64.

Cremer, H., Gore, D., Melles, M., Roberts, D., 2003. Palaeoclimatic significance of Late Quaternary diatom assemblages from Southern Windmill Islands, East Antarctica. Palaeogeogr. Palaeoclimatol. Palaeoecol. 195, 261–280.

Cremer, H., Heiri, O., Wagner, B., Wagner, B., 2007. Abrupt climate warming in East Antarctica during the early Holocene. Quatern. Sci. Rev. 26, 2012–2018.

Cromer, L., Gibson, J.A.E., Swadling, K.M., Ritz, D.A., 2005. Faunal microfossils: indicators of Holocene ecological change in a saline Antarctic lake. Palaeogeogr. Palaeoclimatol. Palaeoecol. 221, 83–97.

Cromer, L., Gibson, J.A.E., Swadling, K.M., Hodgson, D.A., 2006. Evidence for a lacustrine faunal refuge in the Larsemann Hills, East Antarctica, during the Last Glacial Maximum. J. Biogeogr. 33, 1314–1324.

Crosta, X., Romero, O., Armand, L.K., Pichon, J.J., 2005. The biogeography of major diatom taxa in Southern Ocean sediments: 2. open ocean related species. Palaeogeogr. Palaeoclimatol. Palaeoecol. 223, 66–92.

Crosta, X., Debret, M., Denis, D., Courty, M.A., Ther, O., 2007. Holocene long- and short-term climate changes off Adelie Land, East Antarctica. Geochemistry Geophysics Geosystems, 8. art. Nr. Q11009.
Crosta, X., Denis, D., Ther, O., 2008. Sea ice seasonality during the Holocene, Adelie Land, East Antarctica. Mar. Micropaleontol. 66, 222–232.
Davis, B.A.S., Brewer, S., Stevenson, A.C., Guiot, J., 2003. The temperature of Europe during the Holocene reconstructed from pollen data. Quatern. Sci. Rev. 22, 1701–1716.
Denis, D., Crosta, X., Schmidt, S., Carson, D.S., Ganeshram, R.S., Renssen, H., Crespin, J., Ther, O., Billy, I., Giraudeau, J., 2009. Holocene productivity changes off Adelie Land (East Antarctica). Paleoceanography, 24. Art. No. PA3207.
Denton, G.H., Marchant, D.R., 2000. The Geologic basis for a reconstruction of a grounded ice sheet in McMurdo Sound, Antarctica, at the Last Glacial Maximum. Geogr. Ann. Ser. Phys. Geogr. 82A, 167–211.
Divine, D.V., Koc, N., Isaksson, E., Nielsen, S., Crosta, X., Godtliebsen, F., 2010. Holocene Antarctic climate variability from ice and marine sediment cores: insights on ocean–atmosphere interaction. Quatern. Sci. Rev. 29, 303–312.
Domack, E.W., Jull, A.J.T., Nakao, S., 1991. Advance of fast Antarctic outlet glaciers during the Hypsithermal—implications for the volume state if the Antarctic Ice-Sheet under global warming. Geology 19, 1059–1062.
Doran, P.T., Priscu, J.C., Lyons, W.B., Walsh, J.E., Fountain, A.G., McKnight, D.M., Moorhead, D.L., Virginia, R.A., Wall, D.H., Clow, G.D., Fritsen, C.H., McKay, C.P., Parsons, A.N., 2002. Antarctic climate cooling and terrestrial ecosystem response. Nature 415, 517–520.
Doran, P.T., McKay, C.P., Fountain, A.G., Nylen, T., McKnight, D.M., Jaros, C., Barrett, J.E., 2008. Hydrologic response to extreme warm and cold summers in the McMurdo Dry Valleys, East Antarctica. Antarct. Sci. 20, 499–509.
Emslie, S.D., Woehler, E.J., 2005. A 9000-year record of Adélie Penguin occupation and diet in the Windmill Islands, East Antarctica. Antarct. Sci. 17, 57–66.
Emslie, S.D., Coats, L., Licht, K., 2007. A 45, 000 yr record of Adélie Penguins and climate change in the Ross Sea, Antarctica. Geology 35, 61–64.
Fink, D., Mckelvey, B., Hambrey, M.J., Fabel, D., Brown, R., 2006. Pleistocene deglaciation chronology of the Amery Oasis and Radok Lake, Northern Prince Charles Mountains, Antarctica. Earth Planet. Sci. Lett. 243, 229–243.
Finocchiaro, F., Langone, L., Colizza, E., Fontolan, G., Giglio, F., Tuzzi, E., 2005. Records of early Holocene warming in a laminated sediment core from Cape Hallett Bay (Northern Victoria Land, Antarctica). Global Planet. Change 45, 193–206.
Fulford-Smith, S.P., Sikes, E.L., 1996. The evolution of Ace Lake, Antarctica, determined from sedimentary diatom assemblages. Palaeogeogr. Palaeoclimatol. Palaeoecol. 124, 73–86.
Gibson, J.A.E., Paterson, K.S., White, C.A., Swadling, K.M., 2009. Evidence for the continued existence of Abraxas Lake, Vestfold Hills, East Antarctica during the Last Glacial Maximum. Antarct. Sci. 21, 269–278.
Gingele, F., Kuhn, G., Maus, B., Melles, M., Schone, T., 1997. Holocene retreat from the Lazarev Sea shelf, East Antarctica. Cont. Shelf Res. 17, 137–163.
Goodwin, I.D., 1993. Holocene deglaciation, sea-level change, and the emergence of the Windmill Islands, Budd Coast, Antarctica. Quat. Res. 40, 70–80.
Goodwin, I.D., 1996. A Mid to Late Holocene readvance of the Law Dome ice margin, Budd Coast, East Antarctica. Antarct. Sci. 8, 395–406.
Gore, D.B., 1997. Blanketing snow and ice; constraints on radiocarbon dating deglaciation in East Antarctic oases. Antarct. Sci. 9, 336–346.
Gore, D.B., Rhodes, E.J., Augustinus, P.C., Leishman, M.R., Colhoun, E.A., Rees-Jones, J., 2001. Bunger Hills, East Antarctica: ice free at the Last Glacial Maximum. Geology 29, 1103–1106.
Gosse, J.C., Phillips, F.M., 2001. Terrestrial *in situ* cosmogenic nuclides: theory and application. Quatern. Sci. Rev. 20, 1475–1560.
Hall, B.L., Denton, G.H., 1995. Late Quaternary lake levels in the Dry Valleys, Antarctica. Antarct. J. US 30, 52–53.
Hall, B., Denton, G., 1999. New relative sea-level curves for the southern Scott Coast, Antarctica: evidence for Holocene deglaciation of the western Ross Sea. J. Quat. Sci. 14, 641–650.
Hall, B.L., Denton, G.H., 2000a. Extent and chronology of the Ross Sea Ice Sheet and the Wilson Piedmont Glacier along the Scott Coast at and since the Last Glacial Maximum. Geogr. Ann. Ser. Phys. Geogr. 82A, 337–363.
Hall, B.L., Denton, G.H., 2000b. Radiocarbon chronology of Ross Sea Drift, eastern Taylor Valley, Antarctica: evidence for a grounded ice sheet in the Ross Sea at the Last Glacial Maximum. Geogr. Ann. Ser. Phys. Geogr. 82A, 305–336.
Hall, B.L., Denton, G.H., Hendy, C.H., 2000. Evidence from Taylor Valley for a grounded ice sheet in the Ross Sea, Antarctica. Geogr. Ann. Ser. Phys. Geogr. 82A, 275–303.
Hall, B., Denton, G., Overturf, B., 2001. Glacial Lake Wright, a high-level Antarctic lake during the LGM and early Holocene. Antarct. Sci. 13, 53–60.
Hall, B.L., Denton, G.H., 2002. Holocene history of the Wilson Piedmont Glacier along the Southern Scott Coast, Antarctica. Holocene 12, 619–627.
Hall, B.L., Denton, G.H., Overturf, B., Hendy, C.H., 2002. Glacial Lake Victoria, a high-level Antarctic lake inferred from lacustrine deposits in Victoria Valley. J. Quat. Sci. 17, 697–706.
Hall, B., Baroni, C., Denton, G., 2004. Holocene relative sea-level history of the southern Victoria Land Coast, Antarctica. Global Planet. Change 42, 241–263.
Hall, B.L., Hoelzel, A.R., Baroni, C., Denton, G.H., Le Boeuf, B.J., Overturf, B., Topf, A.L., 2006. Holocene elephant seal distribution implies warmer-than-present climate in the Ross Sea. Proc. Nat. Acad. Sci. U.S.A. 103, 10213–10217.
Hall, B.L., 2009. Holocene glacial history of Antarctica and the Sub-Antarctic islands. Quatern. Sci. Rev. 28, 2213–2230.
Hambrey, M.J., Glasser, N.F., Mckelvey, B.C., Sugden, D.E., Fink, D., 2007. Cenozoic landscape evolution of an East Antarctic oasis (Radok Lake area, Northern Prince Charles Mountains), and its implications for the glacial and climatic history of Antarctica. Quatern. Sci. Rev. 26, 598–626.
Hendy, C.H., 2000. Late Quaternary lakes in the McMurdo Sound region of Antarctica. Geogr. Ann. Ser. Phys. Geogr. 82A, 411–432.
Hewitt, C.D., Mitchell, J.F.B., 1998. A Fully Coupled GCM Simulation of the Climate of the Mid-Holocene. Geophys. Res. Lett. 25, 361–364.
Hiller, A., Wandu, A., Kämpf, H., Stackebrandt, W., 1988. Occupation of the Antarctic continent by petrels during the past 35000 years: inferences from a 14 C study of stomach oil deposits. Polar Biol. 9, 69–77.
Hjort, C., Ingólfsson, O., Bentley, M.J., Björck, S., 2003. The late Pleistocene and Holocene glacial and climate history of the Antarctic Peninsula region as documented by the land and lake sediment records—a review. In: Domack, E., Leventer, A., Burnett, A., Bindshadler, R., Convey, P., Kirby, M. (Eds.), Antarctic Peninsula climate variability: Historical and paleonvironmental perspectives, pp. 95–102.
Hodgson, D.A., Noon, P.E., Vyverman, W., Bryant, C.L., Gore, D.B., Appleby, P., Gilmour, M., Verleyen, E., Sabbe, K., Jones, V.J., Ellis-Evans, J.C., Wood, P.B., 2001. Were the Larsemann Hills ice-free through the Last Glacial Maximum? Antarct. Sci. 13, 440–454.
Hodgson, D.A., McMinn, A., Kirkup, H., Cremer, H., Gore, D., Melles, M., Roberts, D., Montiel, P., 2003. Colonization, succession, and extinction of marine floras during a glacial cycle: a case study from the Windmill Islands (East Antarctica) using biomarkers. Paleoceanography 18. doi:10.1029/2002PA000775.
Hodgson, D.A., Doran, P.T., Roberts, D., McMinn, A., 2004. Paleolimnological studies from the Antarctic and sub-Antarctic islands. In: Pienitz, R., Douglas, M.S.V., Smol, J.P. (Eds.), Developments in paleoenvironmental research volume 8. Long-term environmental change in Arctic and Antarctic Lakes. Springer, Dordracht, The Netherlands, pp. 419–474.
Hodgson, D.A., Verleyen, E., Sabbe, K., Squier, A.H., Keely, B.J., Leng, M.J., Saunders, K.M., Vyverman, W., 2005. Late Quaternary climate-driven environmental change in the Larsemann Hills, East Antarctica, multi-proxy evidence from a lake sediment core. Quat. Res. 64, 83–99.
Hodgson, D.A., Verleyen, E., Squier, A.H., Sabbe, K., Keely, B.J., Saunders, K.M., Vyverman, W., 2006a. Interglacial environments of coastal East Antarctica: Comparison of MIS 1 (Holocene) and MIS 5e (last interglacial) lake-sediment records. Quatern. Sci. Rev. 25, 179–197.
Hodgson, D.A., Roberts, D., McMinn, A., Verleyen, E., Terry, B., Corbett, C., Vyverman, W., 2006b. Recent rapid salinity rise in three east Antarctic lakes. J. Paleolimnol. 36, 385–406.
Hodgson, D.A., Smol, J.P., 2008. High latitude paleolimnology. In: Vincent, W.F., Laybourn-Parry, J. (Eds.), Polar Lakes and Rivers—Limnology of Arctic and Antarctic Aquatic Ecosystems. Oxford University Press, Oxford, UK, pp. 43–64.
Hodgson, D.A., Abram, N., Anderson, J., Bargelloni, L., Barrett, P., Bentley, M.J., Bertler, N.A.N., Chown, S., Clarke, A., Convey, P., Crame, A., Crosta, X., Curran, M., di Prisco, G., Francis, J.E., Goodwin, I., Gutt, J., Massé, G., Masson-Delmotte, V., Mayewski, P.A., Mulvaney, R., Peck, L., Pörtner, H.-O., Röthlisberger, R., Stevens, M.I., Summerhayes, C.P., van Ommen, T., Verde, C., Verleyen, E., Vyverman, W., Wiencke, C., Zane, L., 2009a. Antarctic climate and environment history in the pre-instrumental period. In: Turner, J., Convey, P., Di Prisco, G., Mayewski, P.A., Hodgson, D.A., Fahrbach, E., Bindschadler, R., Gutt, J. (Eds.), Antarctic Climate Change and the Environment. Scientific Committee for Antarctic Research, Cambridge, pp. 115–182.
Hodgson, D.A., Verleyen, E., Vyverman, W., Sabbe, K., Leng, M.J., Pickering, M., Keely, B.J., 2009b. A geological constraint on relative sea level in Marine Isotope Stage 3 in the Larsemann Hills, Lambert Glacier region, East Antarctica (31 366–33 228 cal yr BP). Quatern. Sci. Rev. 28, 2689–2696.
Hodgson, D.A., Sime, L.C., 2010. Southern westerlies and CO_2. Nature Geoscience 3, 666–667.
Hughen, K.A., Baillie, M.G.L., Bard, E., Beck, J.W., Bertrand, C.J.H., Blackwell, P.G., Buck, C.E., Burr, G.S., Cutler, K.B., Damon, P.E., Edwards, R.L., Fairbanks, R.G., Friedrich, M., Guilderson, T.P., Kromer, B., Mccormac, G., Manning, S., Ramsey, C.B., Reimer, P.J., Reimer, R.W., Remmele, S., Southon, J.R., Stuiver, M., Talamo, S., Taylor, F.W., Van Der Plicht, J., Weyhenmeyer, C.E., 2004. Marine04 marine radiocarbon age calibration, 0–26 cal kyr BP. Radiocarbon 46, 1059–1086.
Ingólfsson, O., Hjort, C., Berkman, P.A., Björck, S., Colhoun, E., Goodwin, I.D., Hall, B., Hirakawa, K., Melles, M., Moller, P., Prentice, M.L., 1998. Antarctic glacial history since the Last Glacial Maximum: an overview of the record on land. Antarct. Sci. 10, 326–344.
Ingólfsson, O., Hjort, C., Humlum, O., 2003. Glacial and climate history of the Antarctic Peninsula since the Last Glacial Maximum. Arct. Antarct. Alp. Res. 35, 175–186.
Johnson, J.S., Bently, M.J., Bently, O., 2008. First exposure ages from the Amundsen Sea embayment, West Antarctica: the late Quaternary context for recent thinning of Pine Island, Smith, and Pope Glaciers. Geology 36, 223–226.
Jouzel, J., Masson-Delmotte, V., Cattani, O., Dreyfus, G., Falourd, S., Hoffmann, G., Minster, B., Nouet, J., Barnola, J.M., Chappellaz, J., Fischer, H., Gallet, J.C., Johnsen, S., Leuenberger, M., Loulergue, L., Luethi, D., Oerter, H., Parrenin, F., Raisbeck, G., Raynaud, D., Schilt, A., Schwander, J., Selmo, E., Souchez, R., Spahni, R., Stauffer, B., Steffensen, J.P., Stenni, B., Stocker, T.F., Tison, J.L., Werner, M., Wolff, E.W., 2007. Orbital and millennial Antarctic climate variability over the past 800, 000 years. Science 317, 793–796.
Kaufman, D.S., Ager, T.A., Anderson, N.J., Anderson, P.M., Andrews, J.T., Bartlein, P.J., Brubaker, L.B., Coats, L.L., Cwynar, L.C., Duvall, M.L., Dyke, A.S., Edwards, M.E., Eisner, W.R., Gajewski, K., Geirsdottir, A., Hu, F.S., Jennings, A.E., Kaplan, M.R., Kerwin, M.N., Lozhkin, A.V., Macdonald, G.M., Miller, G.H., Mock, C.J., Oswald, W.W., Otto-Bliesner, B.L., Porinchu, D.F., Ruhland, K., Smol, J.P., Steig, E.J., Wolfe, B.B., 2004.

Holocene thermal maximum in the Western Arctic (0–180 degrees W). Quatern. Sci. Rev. 23, 529–560.

Kiernan, K., Gore, D.B., Fink, D., White, D.A., McConnell, A., Sigurdsson, I.A., 2009. Deglaciation and weathering of Larsemann Hills, East Antarctica. Antarct. Sci. 21, 373–382.

Kirkup, H., Melles, M., Gore, D.B., 2002. Late Quaternary environment of Southern Windmill Islands, East Antarctica. Antarct. Sci. 14, 385–394.

Kulbe, T., Melles, M., Verkulich, S.R., Pushina, Z.V., 2001. East Antarctic climate and environmental variability over the Last 9400 years inferred from marine sediments of the Bunger Oasis. Arct. Antarct. Alp. Res. 33, 223–230.

Leventer, A., Domack, E.W., Ishman, S.E., Brachfeld, S., McClennen, C.E., Manley, P., 1996. Productivity cycles of 200–300 years in the Antarctic Peninsula region: Understanding linkages among the sun, atmosphere, oceans, sea ice, and biota. Geol. Soc. Am. Bull. 108, 1626–1644.

Liu, X.D., Sun, L.G., Xie, Z.Q., Yin, X.B., Zhu, R.B., Wang, Y.H., 2007. A preliminary record of the historical seabird population in the Larsemann Hills, East Antarctica, from geochemical analyses of Mochou Lake sediments. Boreas 36, 182–197.

Lorenzini, S., Olmastroni, S., Pezzo, F., Salvatore, M.C., Baroni, C., 2009. Holocene Adelie penguin diet in Victoria Land, Antarctica. Pol. Biol. 32, 1077–1086.

Lyons, W.B., Tyler, S.W., Wharton, R.A., Mcknight, D.M., Vaughn, B.H., 1998. A Late Holocene desiccation of Lake Hoare and Lake Fryxell, McMurdo Dry Valleys, Antarctica. Antarct. Sci. 10, 247–256.

MacKintosh, A., White, D., Fink, D., Gore, D.B., Pickard, J., Fanning, P.C., 2007. Exposure ages from mountain dipsticks in Mac. Robertson Land, East Antarctica, indicate little change in ice-sheet thickness since the Last Glacial Maximum. Geology 35, 551–554.

Mann, M.E., Zhang, Z.H., Hughes, M.K., Bradley, R.S., Miller, S.K., Rutherford, S., Ni, F.B., 2008. Proxy-based reconstructions of hemispheric and global surface temperature variations over the past two millennia. Proc. Nat. Acad. Sci. U.S.A. 105, 13252–13257.

Marshall, G.J., 2003. Trends in the Southern Annular Mode from observations and reanalyses. J. Climatol. 16, 4134–4143.

Marshall, G.J., 2007. Half-century seasonal relationships between the Southern Annular Mode and Antarctic temperatures. Int. J. Climatol. 27, 373–383.

Marshall, G.J., 2009. On the annual and semi-annual cycles of precipitation across Antarctica. Int. J. Climatol. 29, 2298–2308.

Marshall, G.J., Di Battista, S., Naik, S.S., Thamban, M., in press. Analysis of a regional change in the sign of the SAM–temperature relationship in Antarctica. Climate Dynamics. doi:10.1007/s00382-009-0682-9.

Masson, V., Vimeux, F., Jouzel, J., Morgan, V., Delmotte, M., Ciais, P., Hammer, C., Johnsen, S., Lipenkov, V.Y., Mosley-Thompson, E., Petit, J.R., Steig, E.J., Stievenard, M., Vaikmae, R., 2000. Holocene climate variability in Antarctica based on 11 ice-core isotopic records. Quat. Res. 54, 348–358.

Masson-Delmotte, V., Stenni, B., Jouzel, J., 2004. Common millennial-scale variability of Antarctic and Southern Ocean temperatures during the past 5000 years reconstructed from the EPICA Dome C ice core. Holocene 14, 145–151.

Masson-Delmotte, V., Stenni, B., Pol, K., Braconnot, P., Cattani, O., Falourd, S., Kageyama, M., Jouzel, J., Landais, A., Minster, B., Barnola, J.M., Chappellaz, J., Krinner, G., Johnsen, S., Rothlisberger, R., Hansen, J., Mikolajewicz, U., Otto-Bliesner, B., 2010a. EPICA Dome C record of glacial and interglacial intensities. Quatern. Sci. Rev. 29, 113–128.

Masson-Delmotte, V., Stenni, B., Blunier, T., Cattani, O., Chappellaz, J., Cheng, H., Dreyfus, G., Edwards, R.L., Falourd, S., Govin, A., Kawamura, K., Johnsen, S.J., Jouzel, J., Landais, A., Lemieux-Dudon, B., Lourantou, A., Marshall, G., Minster, B., Mudelsee, M., Pol, K., Rothlisberger, R., Selmo, E., Waelbroeck, C., 2010b. Abrupt change of Antarctic moisture origin at the end of Termination II. Proc. Nat. Acad. Sci. U.S.A. 107, 12091–12094.

Matsumoto, G.I., Komori, K., Enomoto, A., Imura, S., Takemura, T., Ohyama, Y., Kanda, H., 2006. Environmental changes in Syowa Station area of Antarctica during the last 2300 years inferred from organic components in lake sediment cores. Pol. Biosci. 19, 52–61.

Mayewski, P.A., Rohling, E.E., Stager, J.C., Karlen, W., Maasch, K.A., Meeker, L.D., Meyerson, E.A., Gasse, F., Van Kreveld, S., Holmgren, K., Lee-Thorp, J., Rosqvist, G., Rack, F., Staubwasser, M., Schneider, R.R., Steig, E.J., 2004. Holocene climate variability. Quat. Res. 62, 243–255.

Mayewski, P.A., Meredith, M.P., Summerhayes, C.P., Turner, J., Worby, A., Barrett, P.J., Casassa, G., Bertler, N.A.N., Bracegirdle, T., Garabato, A.C.N., Bromwich, D., Campbell, H., Hamilton, G.S., Lyons, W.B., Maasch, K.A., Aoki, S., Xiao, C., Van Ommen, T., 2009. State of the Antarctic and Southern Ocean Climate System. Rev. Geophys. 47, 1–38.

McCormac, F.G., Hogg, A.G., Blackwell, P.G., Buck, C.E., Higham, T.F.G., Reimer, P.J., 2004. SHcal04 Southern Hemisphere Calibration, 0–11.0 cal kyr BP. Radiocarbon 46, 1087–1092.

McMinn, A., 2000. Late Holocene increase in sea ice extent in fjords of the Vestfold Hills, Eastern Antarctica. Antarct. Sci. 12, 80–88.

McMinn, A., Heijnis, H., Harle, K., McOrist, G., 2001. Late-Holocene climatic change recorded in sediment cores from Ellis Fjord, eastern Antarctica. Holocene 11, 291–300.

Melles, M., Kulbe, T., Verkulich, S.R., Pushina, Z.V., Hubberten, H.-W., 1997. Late Pleistocene and Holocene environmental history of Bunger Hills, East Antarctica, as revealed by fresh-water and epishelf lake sediments. In: Ricci, C.A. (Ed.), The Antarctic Region: Geological Evolution and Processes, Sienna University, Sienna.

Miura, H., Moriwaki, K., Maemoku, H., Hirakawa, K., 1998. Fluctuations of the East Antarctic Ice Sheet margin since the last glaciation from the stratigraphy of raised beach deposits along the Soya Coast. Ann. Glaciol. 27, 297–301.

Nielsen, S.H.H., Koc, N., Crosta, X., 2004. Holocene climate in the Atlantic sector of the Southern Ocean: controlled by insolation or oceanic circulation? Geology 32, 317–320.

Okuno, J., Miura, H., Maemoku, H., 2007. The possibility of the rapid melting of ice sheet in Holocene around the Lützow-Holm Bay, Antarctica. Geophys. Res. Abstr. 7.

Petit, J.R., Jouzel, J., Raynaud, D., Barkov, N.I., Barnola, J.M., Basile, I., Bender, M., Chappellaz, J., Davis, M., Delaygue, G., Delmotte, M., Kotlyakov, V.M., Legrand, M., Lipenkov, V.Y., Lorius, C., Pepin, L., Ritz, C., Saltzman, E., Stievenard, M., 1999. Climate and atmospheric history of the past 420, 000 years from the Vostok ice core, Antarctica. Nature 399, 429–436.

Pickard, J., Adamson, D.A., Heath, C.W., 1986. The evolution of Watts Lake, Vestfold Hills, East Antarctica, from marine inlet to freshwater lake. Palaeogeogr. Palaeoclimatol. Palaeoecol. 53, 271–288.

Quayle, W.C., Peck, L.S., Peat, H., Ellis-Evans, J.C., Harrigan, P.R., 2002. Extreme responses to climate change in Antarctic lakes. Science 295 645–645.

Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Bertrand, C.J.H., Blackwell, P.G., Buck, C.E., Burr, G.S., Cutler, K.B., Damon, P.E., Edwards, R.L., Fairbanks, R.G., Friedrich, M., Guilderson, T.P., Hogg, A.G., Hughen, K.A., Kromer, B., MccCormac, G., Manning, S., Ramsey, C.B., Reimer, R.W., Remmele, S., Southon, J.R., Stuiver, M., Talamo, S., Taylor, F.W., Van Der Plicht, J., Weyhenmeyer, C.E., 2004. Intcal04 terrestrial radiocarbon age calibration, 0–26 cal kyr BP. Radiocarbon 46, 1029–1058.

Renssen, H., Goosse, H., Fichefet, T., Masson-Delmotte, V., Koc, N., 2005. Holocene climate evolution in the high-latitude Southern Hemisphere simulated by a coupled atmosphere–sea ice–ocean-vegetation model. Holocene 15 (7), 951–964.

Roberts, D., McMinn, A., 1996. Relationships between surface sediment diatom assemblages and water chemistry gradients in saline lakes of the Vestfold Hills, Antarctica. Antarct. Sci. 8, 331–341.

Roberts, D., McMinn, A., 1998. A weighted-averaging regression and calibration model for inferring lake water salinity from fossil diatom assemblages in saline lakes of the Vestfold Hills: a new tool for interpreting Holocene lake histories in Antarctica. J. Paleolimnol. 19, 99–113.

Roberts, D., McMinn, A., 1999. A diatom-based palaeosalinity history of Ace Lake, Vestfold Hills, Antarctica. Holocene 9, 401–408.

Roberts, D., Van Ommen, T.D., McMinn, A., Morgan, V., Roberts, J.L., 2001. Late-Holocene East Antarctic climate trends from ice-core and lake-sediment proxies. Holocene 11, 117–120.

Roberts, D., McMinn, A., Cremer, H., Gore, D.B., Melles, M., 2004. The Holocene evolution and palaeosalinity history of Beall Lake, Windmill Islands (East Antarctica) using an expanded diatom-based weighted averaging model. Palaeogeogr. Palaeoclimatol. Palaeoecol. 208, 121–140.

Rohling, E.J., Pälike, H., 2005. Centennial-scale climate cooling with a sudden cold event around 8, 200 years ago. Nature 434, 975–979.

Schaefer, J.M., Denton, G.H., Kaplan, M., Putnam, A., Finkel, R.C., Barrell, D.J.A., Andersen, B.G., Schwartz, R., Mackintosh, A., Chinn, T., Schluchter, C., 2009. High-Frequency Holocene Glacier Fluctuations in New Zealand Differ From the Northern Signature. Science 324, 622–625.

Schneider, D.P., Steig, E.J., 2008. Ice Cores Record Significant 1940 s Antarctic Warmth Related to Tropical Climate Variability. Proc. Nat. Acad. Sci. U.S.A. 105, 12154–12158.

Siegert, M.J., 2001. Ice sheets and late Quaternary environmental change. John Wiley & Sons. 231 pp.

Singh, S.M., Tiwari, A.K., 2004. Deep lake sampling in Antarctica using helicopters. Curr. Sci. 87, 420.

Sinha, R., Sharma, C., Chauhan, M.S., 2000. Sedimentological and pollen studies of Lake Priyadarshini, Eastern Antarctica. Palaeobotanist 49, 1–8.

Steig, E.J., Morse, D.L., Waddington, E.D., Stuiver, M., Grootes, P.M., Mayewski, P.A., Twickler, M.S., Whitlow, S.I., 2000. Wisconsinan and Holocene climate history from an ice core at Taylor Dome, Western Ross Embayment, Antarctica. Geogr. Ann. Ser. Phys. Geogr. 82A, 213–235.

Steig, E.J., Schneider, D.P., Rutherford, S.D., Mann, M.E., Comiso, J.C., Shindell, D.T., 2009. Warming of the Antarctic Ice-Sheet surface since the 1957 International Geophysical Year. Nature 457, 459.

Stenni, B., Masson-Delmotte, V., Selmo, E., Oerter, H., Meyer, H., Rothlisberger, R., Jouzel, J., Cattani, O., Falourd, S., Fischer, H., Hoffmann, G., Iacumin, P., Johnsen, S.J., Minster, B., Udisti, R., 2010. The deuterium excess records of EPICA Dome C and Dronning Maud Land ice cores (East Antarctica). Quatern. Sci. Rev. 29, 146–159.

Stevens, B.B., Keeling, R.F., 2000. The influence of Antarctic sea ice on glacial–interglacial CO_2 variations. Nature 404, 171–174.

Stocker, T.F., Johnsen, S.J., 2003. A minimum thermodynamic model for the bipolar seesaw. Paleoceanography 18 (4). doi:10.1029/2003pa000920.

Taylor, F., McMinn, A., 2002. Late quaternary diatom assemblages from Prydz Bay, Eastern Antarctica. Quat. Res. 57, 151–161.

Thompson, D.W.J., Solomon, S., 2002. Interpretation of recent Southern Hemisphere climate change. Science 296, 895–899.

Turner, J., Arthern, R., Bromwich, D., Marshall, G., Worby, T., Bockheim, J., di Prisco, G., Verde, C., Convey, P., Roscoe, H., Jones, A., Vaughan, D., Woodworth, P., Scambos, T., Cook, A., Lenton, A., Comiso, J., Gugliemin, M., Summerhayes, C., Meredith, M., Naveira-Garabato, A., Chown, S., Stevens, M., Adams, B., Worland, R., Hennion, F., Huiskes, A., Bergstrom, D., Hodgson, D.A., Bindschadler, R., Bargagli, R., Metzl, N., van der Veen, K., Monaghan, A., Speer, K., Rintoul, S., Hellmer, H., Jacobs, S., Heywood, K., Holland, D., Yamanouchi, T., Barbante, C., Bertler, N., Boutron, C., Hong, S., Mayewski, P., Fastook, J., Newsham, K., Robinson, S., Forcarda, J., Trathan, P., Smetacek, V., Gutt, J., Pörtner, H.-O., Peck, L., Gili, J.-M., Wiencke, C., Fahrbach, E., Atkinson, A., Webb, D., Isla, E., Orejas, C., Rossi, S., Shanklin, J., 2009. The Instrumental Period. In: Turner, J., Convey, P., Di Prisco, G., Mayewski, P.A., Hodgson, D.A., Fahrbach, E., Bindschadler, R., Gutt, J. (Eds.), Antarctic Climate Change and the Environment. Scientific Committee for Antarctic Research, Cambridge, pp. 183–298.

Van de Berg, W.J., Van den Broeke, M.R., Reijmer, C.H., Van Meijgaard, E., 2005. Characteristics of the Antarctic surface mass balance, 1958–2002, using a regional atmospheric climate model. Ann. Glaciol. 41, 97–104.

Van den Broeke, M.R., van Lipzig, N.P.M., 2003. Response of wintertime Antarctic temperatures to the Antarctic Oscillation: results of a regional climate model. In: Domack, E., Leventer, A., Burnett, A., Bindschadler, R., Convey, P., Kirby, M. (Eds.), Antarctic Peninsula climate variability: Historical and paleoenvironmental perspectives. : Antarctic Research Series, Vol. 79. American Geophysical Union, Washington, DC, pp. 43–58.

Vaughan, D.G., Marshall, G.J., Connolley, W.M., Parkinson, C., Mulvaney, R., Hodgson, D.A., King, J.C., Pudsey, C.J., Turner, J., 2003. Recent rapid regional climate warming on the Antarctic Peninsula. Clim. Change 60, 243–274.

Verkulich, S.R., Hiller, A., 1994. Holocene deglaciation of the Bunger Hills revealed by C-14 measurements on stomach oil deposits in snow petrel colonies. Antarct. Sci. 6, 395–399.

Verkulich, S.R., Melles, M., Hubberten, H.W., Pushina, Z.V., 2002. Holocene environmental changes and development of Figurnoye Lake in the Southern Bunger Hills, East Antarctica. J. Paleolimnol. 28, 253–267.

Verleyen, E., Hodgson, D.A., Vyverman, W., Roberts, D., McMinn, A., Vanhoutte, K., Sabbe, K., 2003. Modelling diatom responses to climate induced fluctuations in the moisture balance in continental Antarctic lakes. J. Paleolimnol. 30, 195–215.

Verleyen, E., Hodgson, D.A., Sabbe, K., Vanhoutte, K., Vyverman, W., 2004a. Coastal oceanographic conditions in the Prydz Bay Region (East Antarctica) during the Holocene recorded in an isolation basin. Holocene 14, 246–257.

Verleyen, E., Hodgson, D.A., Sabbe, K., Vyverman, W., 2004b. Late Quaternary deglaciation and climate history of the Larsemann Hills (East Antarctica). J. Quat. Sci. 19, 361–375.

Verleyen, E., Hodgson, D.A., Sabbe, K., Vyverman, W., 2005. Late Holocene changes in ultraviolet radiation penetration recorded in an East Antarctic lake. J. Paleolimnol. 34, 191–202.

Wagner, B., Cremer, H., Hultzsch, N., Gore, D.B., Melles, M., 2004. Late Pleistocene and Holocene history of Lake Terrasovoje, Amery Oasis, East Antarctica, and its climatic and environmental implications. J. Paleolimnol. 32, 321–339.

Wagner, B., Melles, M., Doran, P.T., Kenig, F., Forman, S.L., Pierau, R., Allen, P., 2006. Glacial and postglacial sedimentation in the Fryxell Basin, Taylor Valley, Southern Victoria Land, Antarctica. Palaeogeogr. Palaeoclimatol. Palaeoecol. 241, 320–337.

Wagner, B., Hultzsch, N., Melles, M., Gore, D.B., 2007. Indications of Holocene sea-level rise in Beaver Lake, East Antarctica. Antarct. Sci. 19, 125–128.

Wanner, H., Beer, J., Butikofer, J., Crowley, T.J., Cubasch, U., Fluckiger, J., Goosse, H., Grosjean, M., Joos, F., Kaplan, J.O., Kuttel, M., Muller, S.A., Prentice, I.C., Solomina, O., Stocker, T.F., Tarasov, P., Wagner, M., Widmann, M., 2008. Mid- to Late Holocene climate change: an overview. Quatern. Sci. Rev. 27, 1791–1828.

Wasley, J., Robinson, S.A., Lovelock, C.E., Popp, M., 2006. Some like it wet—an endemic Antarctic bryophyte likely to be threatened under climate change induced drying. Funct. Plant Biol. 33, 443.

Webster, J., Hawes, I., Downes, M., Timperley, M., Howardwilliams, C., 1996. Evidence for regional climate change in the recent evolution of a high latitude pro-glacial lake. Antarct. Sci. 8, 49–59.

White, D.A., Bennike, O., Berg, S., Harley, S.L., Fink, D., Kiernan, K., Mcconnell, A., Wagner, B., 2009. Geomorphology and glacial history of Rauer Group, East Antarctica. Quat. Res. 72, 80–90.

Whittaker, T., Hall, B., Hendy, C., Spaulding, S., 2008. Holocene depositional environments and surface-level changes at Lake Fryxell, Antarctica. Holocene 18, 775–786.

Wolff, E.W., Fischer, H., Fundel, F., Ruth, U., Twarloh, B., Littot, G.C., Mulvaney, R., Rothlisberger, R., De Angelis, M., Boutron, C.F., Hansson, M., Jonsell, U., Hutterli, M.A., Lambert, F., Kaufmann, P., Stauffer, B., Stocker, T.F., Steffensen, J.P., Bigler, M., Siggaard-Andersen, M.L., Udisti, R., Becagli, S., Castellano, E., Severi, M., Wagenbach, D., Barbante, C., Gabrielli, P., Gaspari, V., 2006. Southern Ocean sea-ice extent, productivity and iron flux over the past eight glacial cycles. Nature 440, 491–496.

Zwartz, D.P., Miura, H., Takada, M., Moriwaki, K., 1998. Holocene lake sediments and sea-level change at Mt. Riiser-Larsen. Pol. Geosci. 11, 249–259.