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Glacial and Climate History of the Antarctic Peninsula since the Last Glacial Maximum

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Abstract

During the Last Glacial Maximum (LGM), ice thickened considerably and expanded toward the outer continental shelf around the Antarctic Peninsula. Deglaciation occurred between >14 ka BP and ca. 6 ka BP, when interglacial climate was established in the region. Deglaciation of some local sites was as recent as 4–3 ka BP. After a climate optimum, peaking ca. 4–3 ka BP, a distinct climate cooling occurred. It is characterized at a number of sites by expanding glaciers and ice shelves. Rapid warming during the past 50 yr may be causing instability of some Antarctic Peninsula ice shelves. Detailed reconstructions of the glacial and climatic history of the Antarctic Peninsula since LGM are hampered by scarcity of available archives, low resolution of many datasets, and problems in dating samples. Consequently, the configuration of LGM ice sheets, pattern of subsequent deglaciation, and environmental changes are poorly constrained both temporally and spatially.

Introduction

The Antarctic Peninsula region (Fig. 1), defined as the area between 62° and 75°S and 55° and 80°W and including the South Shetland Islands and the islands in the northwestern Weddell Sea, can be roughly divided into two climatic zones: polar maritime on the western side and polar continental east of the peninsula (Martin and Peel, 1978). The western side is subject to frequent passages of relatively warm, saturated cyclones approaching from the west. Precipitation here may exceed 1000 to 1500 mm water equivalent yr^{-1}, while east of the topographic divide on the Antarctic Peninsula precipitation is in the range of 100 to 200 mm yr^{-1} (Robin and Adie, 1964; Aristarain et al., 1987). There is a high level of interannual variability in mean temperatures in the Antarctic Peninsula region, but generally the western-side mean annual air temperatures lie between –3 and –10°C on a transect from 63°S to 73°S, while they lie between –5 and –17°C on a similar transect along the east coast (Vaughan and Doake, 1996). These differences in both precipitation and temperatures are reflected in the position of the equilibrium line altitude (ELA): because of heavier precipitation on the western side of the Antarctic Peninsula ELA often lies at <100 m a.s.l., while due to a combination of low temperatures and low precipitation it can locally lie at >400 m a.s.l. on the Weddell Sea side. Antarctic Peninsula glaciers typically descend to the coast and terminate in ice cliffs along the coast, as calving tidewater glaciers or ice shelves.

The Antarctic Peninsula glacial system is sensitive to changes in climate and sea level. Smith et al. (1999) and Domack et al. (2001a) suggested that the Antarctic Peninsula region encompassed one of the most dynamic climate systems on Earth, where the ecological and cryospheric systems respond rapidly to climatic changes. Recent signs of accelerating retreat of fringing ice shelves, in combination with rapid warming (>2°C) in the central and southern part of the western Antarctic Peninsula over the past 50 yr (Stark, 1994) have raised concerns as to the future stability of the glacial system (Doake et al., 1998; Oppenheimer, 1998; Skvarca et al., 1999). Knowledge of past ice extent and the history of glacial and climatic fluctuations are essential for understanding what controls the dynamics and spatial patterns of glacial fluctuations. Outstanding research questions include the following: What was the Last Glacial Maximum (LGM) configuration of the Antarctic Peninsula ice sheet, and was it a concentric ice sheet or a number of confluent local ice caps and domes? What drove the initial ice retreat from LGM positions: global sea-level rise or regional warming? What was the timing and pattern of the deglaciation of presently ice-free coastal areas? What specifically characterizes the transition from glacial climate to the present interglacial climate situation? Can we define temporally a climate optimum and glacial minima in the Antarctic Peninsula region, and what is happening to Antarctic Peninsula glaciers today? Also, constraining the glacial and climatic development in the Antarctic Peninsula region since LGM is necessary for providing boundary conditions for glaciological modeling, and for understanding the contribution of Antarctic Peninsula ice to Holocene global sea-level rise (Bentley, 1999; Ingólfsson and Hjort, 1999).

Timing of the last maximum ice extent in the Antarctic Peninsula region is poorly known, and there is a lack of ^{14}C dates and well-constrained chronologies for the LGM positions of ice margins and the subsequent deglaciation of the Antarctic Peninsula continental shelves. This is partly due to lack of datable material in transitional glacial marine sediments deposited below floating ice shelves in proximity to the grounding lines of the retreating glaciers. Another problem is that obtained ^{14}C ages can be both contaminated by old detrital carbon or affected by abnormally high local seawater reservoir effects, making reliable chronologies difficult to establish (Domack, 1992; Harden et al., 1992; Domack et al., 2000, 2001a; Pudsey and Evans, 2001; and general discussions on the dating of calcareous marine shells in Björck et al., 1991a; Gordon and Harkness, 1992; Berkman and Forman, 1996; Berkman et al., 1998; Ingólfsson et al., 1998). While most authors assume that the last Antarctic peak glaciation coincided with the timing of lowest global sea levels during marine oxygen isotope stage (MIS) 2, ca. 20–18 ka BP (^{14}C kiloyears before present), circum-Antarctic field data suggest considerable regional differences in glacial maxima between 20 and 10 ka BP (e.g., Stuiver et al., 1981; Anderson et al., 1992; Domack et al., 1995, 1999; Licht et al., 1996; Kellogg et al., 1996; Ingólfsson et al., 1998; Hall and Denton, 1999, 2000; Denton and Marchant, 2000). It has been suggested that the last maximum ice extent in the Antarctic Peninsula region occurred later than 30 ka BP (Sugden

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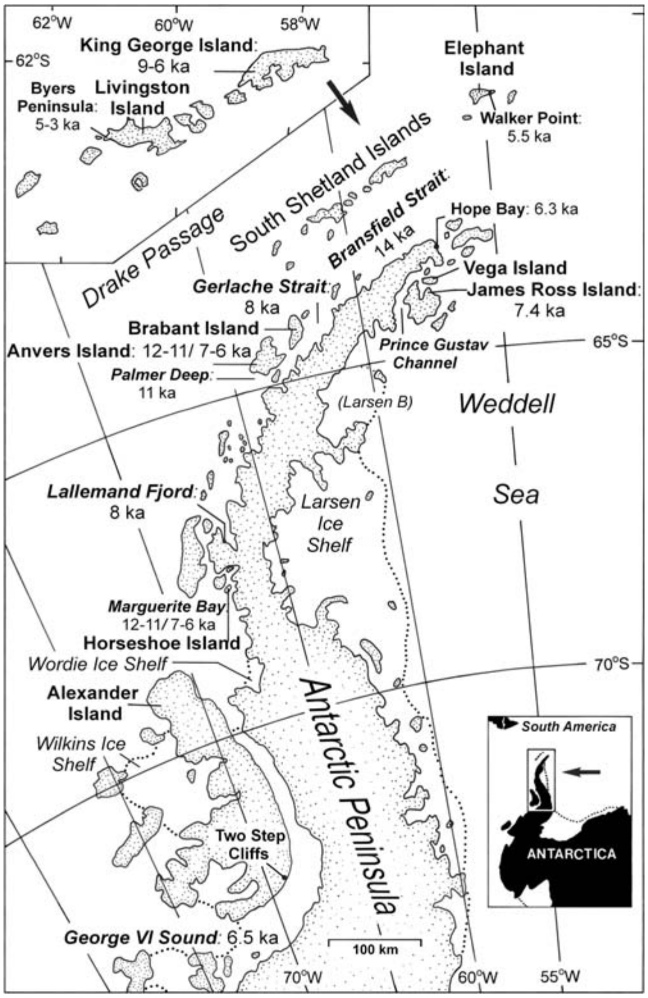

FIGURE 1. The Antarctic Peninsula region, with localities mentioned in the text. Numbers refer to constraining minimum dates for the deglaciation, in ^{14}C kiloyears (ka) before present. Where reported as, e.g., Anvers Island 12–11/7–6 ka, the first set of numbers refers to deglaciation of outer-middle continental shelf areas; the second set refers to deglaciation of inner shelf, fjords, and embayments.

and Clapperton, 1980) and prior to 14–12 ka BP (Anderson et al., 2002; Banfield and Anderson, 1995; Domack et al., 2001a). Recent interpretations, however, suggest that the ultimate Wisconsinan ice sheet maximum may predate the 20–18 ka BP LGM: Bentley and Anderson (1998) suggested that ice volumes were considerably greater before 35 ka BP compared to after 35 ka BP, and Hjort et al. (1997, in press) likewise argue that the maximum ice extent since the last interglacial (MIS 5e) probably occurred before MIS 3 in the western Weddell Sea. Berkman et al. (1998) argued, on the basis of circum-Antarctic finds of fossil marine shells of ages at or beyond the limit of radiocarbon dating, that MIS 3 was an important interstadial period around Antarctica, but good stratigraphic and chronological control for a middle Wisconsin interstadial event is still lacking.

This paper reviews the most important types of proxy data used for putting constraints on the glacial and climatic history of the Antarctic Peninsula region. We aim to synthesize the available data into a tentative hypothesis concerning the environmental history of this area since the most recent pre-Holocene global glacial maxima, the 20–18 ka BP LGM. All ^{14}C age determinations of marine samples referred to below are reservoir corrected by 1,300 yr (Berkman and Forman, 1996) but not calibrated, and where appropriate, the material dated is also given.

Onshore Evidence for the Extent of Last Glacial Maximum Ice in the Antarctic Peninsula

ICE EXTENT, ICE THICKNESS, AND ICE FLOW DIRECTIONS

Evidence of more extensive ice cover than today is present on every ice-free lowland area along the Antarctic Peninsula and its outlying islands, in the form of glacial drift, erratics, and striations (e.g., Curl, 1980; Sugden and Clapperton, 1980; Clapperton, 1990). Directional evidence suggests ice flow offshore toward the continental shelf areas around the Antarctic Peninsula. The distribution of erratics and the direction of glacial striae on the lowlands mostly agree with ice flows constrained by local topography, where outlet glaciers have occupied valleys, fjords, and straits and flowed toward the coastal areas and shelves from ice divides in the Antarctic Peninsula mountain chain. The available chronological controls, in the form of radiocarbon dates from organic material in glacial drift and overlying sediments, provide minimum ages for this glacial event as being Late Wisconsinan–early Holocene (Sugden and John, 1973; Hjort et al., 1997; Ingólfsson et al., 1998).

Evidence of LGM ice thickness and ice movements includes ice-abraded ridge crests, striations, erratics, and glacial drift on nunataks and coastal mountains. Ancient glacial trimlines are rarely recognized. Systematic attempts to map altitudinal evidence of LGM ice thickness in the Antarctic Peninsula region have been made on its southern and central parts (Carrara, 1979, 1981; Clapperton and Sugden, 1982; Waitt, 1983; Bentley and Anderson, 1998; Bentley et al., 2000; Hjort et al., in press) and in the South Shetland Islands (John and Sugden, 1971; Sugden and John, 1973; Curl, 1980). Carrara (1979, 1981) and Waitt (1983) concluded that striations and erratics on nunataks on the southern Antarctic Peninsula showed that sometime in the past these were overridden by a considerably thickened ice sheet, at least 500 m thicker than today (Waitt, 1983; Bentley and Anderson, 1998). Bentley et al. (2000) suggested that in the southern and central parts of the Antarctic Peninsula, the Wisconsinan ice sheet, sometime prior to 35 ka BP, had expanded to form two ice domes and probably merged with an expanded West Antarctic Ice Sheet in the Weddell Sea. Based on cosmogenic exposure ages from erratics, they suggested ice-sheet lowering in the southern part of the Antarctic Peninsula, signifying deglaciation, already at ca. 16 ka BP. Bentley and Anderson (1998), Bentley (1999), and Anderson et al. (2002) reconstructed the LGM configuration of an Antarctic Peninsula ice sheet, with ice flowing out toward the shelf edges from an ice divide stretching north-south along the axis of the peninsula (Fig. 2).

A number of additional investigations infer former ice thickness on the northern part of the Antarctic Peninsula region and the islands in the western Weddell Sea (e.g., Elliot, 1981; Rabassa, 1983; Ingólfsson et al., 1992; Hjort et al., 1997). Hjort et al. (1997, in press) presented evidence suggesting that presently ice-free areas on James Ross, Vega, and Seymour Islands in the northwest Weddell Sea had, at least once, been overridden by glaciers advancing from the Antarctic Peninsula, depositing tills and crystalline erratics up to 300–400 m a.s.l. Hjort et al. (in press) suggest that this period of extensive glaciation was contemporaneous with Bentley and Anderson's (1998) LGM grounding line position some 200 km farther east, but was probably older than 30 ka BP. Hjort et al. (in press) suggest that the ice flowing east off the peninsula during LGM was thinner, channeled through the straits and troughs, and did not attain much greater thickness there than ca. 150 m a.s.l.

RAISED BEACHES

Raised beaches on the Antarctic Peninsula can generally be regarded as isostatic fingerprinting of earlier expanded ice volumes compared to present. Raised marine beach deposits, which postdate the last major glaciation in the Antarctic Peninsula region, have been described from most ice-free areas on the peninsula and nearby islands (e.g., John and Sugden, 1971; Curl, 1980; Birkenmajer, 1981; Ingólfsson et al., 1992; López-Martinez et al., 1996; Hjort et al., 1997). These suggest expanded ice volumes prior to the early–mid-Holocene, compared to the present. Since there are as yet no well-constrained relative sea-level curves for any part of the region, the history of relative sea-level changes cannot give more than rough constraints on possible ice volumes and deglaciation history. In the northern part of the region the postglacial marine limit generally lies at altitudes between 15 and 20 m a.s.l., with the highest marine limit on James Ross Island at 30 m a.s.l. (Hjort et al., 1997). There, a ^{14}C analysis on *in situ* mollusk shells from sublittoral sediments related to the marine limit date it to ca. 7.5 ka BP. The marine limit rises toward the south, and is highest on Horseshoe Island in Marguerite Bay, at ca. 55 m a.s.l. (Hjort and Ingólfsson, 1990). The Horseshoe Island marine limit has not been absolutely dated, but is constrained by minimum ages for the deglaciation of Marguerite Bay as being older than 7–6 ka BP (Kennedy and Anderson, 1989; Harden et al., 1992; Pope and Anderson, 1992). The north-south gradient in altitude of the marine limit agrees with the conclusion of Bentley and Anderson (1998) and Anderson et al. (2002) that there were greater LGM ice volumes in the central and southern parts of the region than in the northern part.

In summary, there is ample evidence on land of more extensive ice cover during LGM than today. In the central and southern part, the ice may have been >500 m thicker than today (Waitt, 1983; Bentley and Anderson, 1998), while in the northern part estimates vary between 150 and 400 m (Hjort et al. 1997, in press).

Offshore Evidence for the Extent of Last Glacial Maximum Ice

There is strong evidence of grounded ice sheets and outlet glaciers extending onto the Antarctic Peninsula continental shelf areas during the LGM. (e.g., Kennedy and Anderson, 1989; Herron and Anderson, 1990; Larter and Bartek, 1991; Anderson et al., 1991a, 1992, 2002; Pope and Anderson, 1992; Banfield and Anderson, 1995; Bart and Anderson, 1996; Larter and Vanneste, 1995; Sloan et al., 1995; Canals et al., 2000; Domack et al., 2001a). The marine geological evidence includes the following:

1. Shelf bathymetry: glacial troughs, moraine ridges, drumlins, flutes, and other glacial lineations. These delineate the drainage of glaciers and show that the shelf areas have been shaped by erosion and deposition below and in front of moving outlet glaciers and ice streams.
2. High-resolution seismic records: glacial unconformities, sediment cover, grounding-zone wedges. These records signify the extent of glacial erosion and subsequent deposition on the shelf.
3. Sediment cores: sedimentological and petrographic analyses for identifying tills and glaciomarine sediments. The tills are first-order data on former ice extent, and ^{14}C dates from glaciomarine sediments provide constraining minimum dates for deglaciation of the shelf areas.

ICE EXTENT ON THE WESTERN CONTINENTAL SHELF

Kennedy and Anderson (1989) and Anderson et al. (1991b) suggested that grounded ice had extended across the shelf areas off Marguerite Bay (Fig. 1) during the LGM, based on a prominent glacial erosion surface identified. Seismic studies by Bart and Anderson (1996) showed that this unconformity extends to the shelf break. Hiemstra (2001) recognized a grounded-ice-to-recessional-ice scenario for Marguerite Bay, based on microscopic analyses of glacigenic core sediments. Anderson et al. (2002) presented swath bathymetric records that are interpreted to show drumlins, flutes, and striations in the landward portions of the Marguerite Bay trough. They also described megascale glacial lineations in the seaward portion of the trough, extending into a prominent grounding zone wedge on the midshelf.

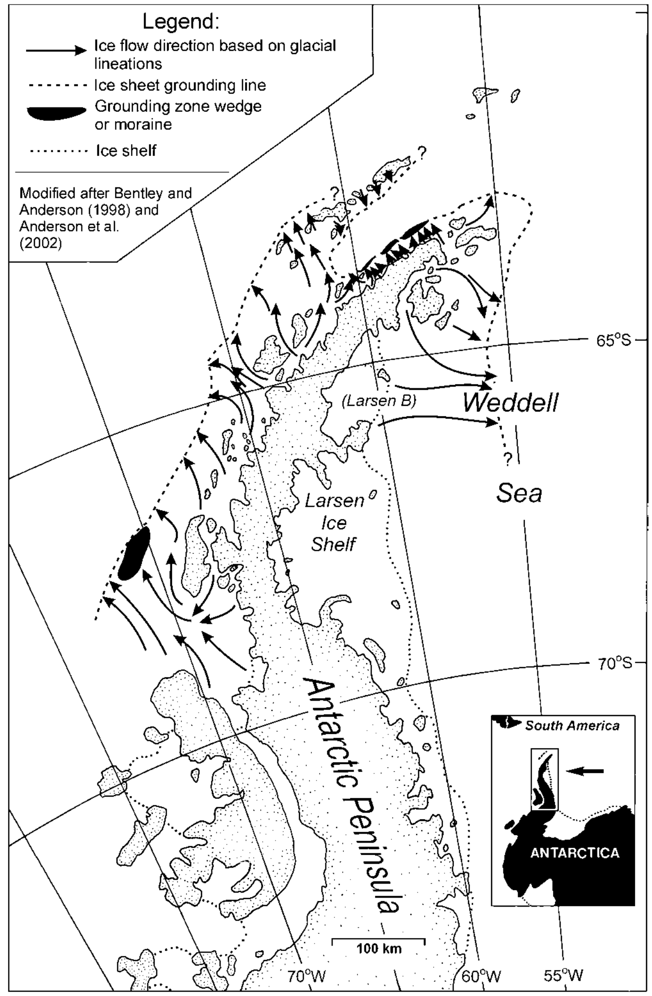

FIGURE 2. Grounding line positions and paleodrainage of the LGM Antarctic Peninsula ice. Modified after Bentley and Anderson (1998) and Anderson et al. (2002). Present-day ice shelves are also shown.

Pudsey et al. (1994) published side-scan sonar records offshore Anvers Island that show glacial flutes on the inner shelf there. Larter and Vanneste (1995) described relict subglacial till deltas on the Antarctic Peninsula outer shelf, off Anvers Island, which they interpreted as products of progradation at former ice-stream grounding lines. They took this discovery to imply that grounded ice on the outer shelf had been low profile and fast flowing. Constraining minimum age for the till deltas of 11.3 ka BP was provided by ^{14}C dates from stratigraphically overlying glacial marine sediments (Pudsey et al., 1994). A study by Domack et al. (2001a) of glaciomarine sediments on top of a diamicton in the Palmer Deep also suggests that the shelf close to Anvers Island was covered by ice prior to ca. 11 ka BP.

Banfield and Anderson (1995) recognized a widespread glacial unconformity extending to the continental shelf break off the Bransfield Strait, with prominent ridges occurring above the unconformity, within troughs on the middle shelf. Bentley and Anderson (1998) suggest that these ridges are moraines, marking the LGM grounding line (Fig. 2). A recent swath bathymetric documentation by Canals et al. (2000) showed that during a past glacial maximum a major ice stream had flowed off the northwestern side of the Antarctic Peninsula, depositing a >100-km-long, elongated, ice-molded sediment body. They argued that it had been deposited below >1000-m-thick ice, under conditions of very fast ice flow. The age of these deposits is not yet constrained by ^{14}C dates, and a LGM age is still circumstantial. Canals et al. (2000) proposed that during LGM a separate ice dome had existed over Brabant and Anvers Islands, with its eastward draining entering the Gerlache Strait and flowing northward into the Bransfield basin.

Bentley and Anderson (1998) and Anderson et al. (2002) reviewed available data on LGM grounding line positions. They concluded that LGM glaciers flowing west from the Antarctic Peninsula had generally grounded on the middle-outer shelf, at water depths of ca. 400 m (Fig. 2), and that Antarctic Peninsula ice had met with the local ice cap over the South Shetland Islands.

ICE EXTENT IN THE NORTHWESTERN WEDDELL SEA

Due to inaccessibility, the geologic data from the western Weddell Sea is scarcer than data from west of the peninsula. Sea ice has long prevented the acquisition of marine geologic data south of 65°S. Shelf bathymetric data indicate submarine troughs, which have drained large outlet glaciers from the Antarctic Peninsula into the Weddell Sea (Sloan et al., 1995). Anderson et al. (1992) and Bentley and Anderson (1998) suggested, on the basis of high-resolution seismic records and sediment cores, that Prince Gustav Channel and adjoining bays contain little sediment. They interpreted this to indicate recent glacial erosion. Multibeam records from the inner shelf show flutes and other glacial lineations (Domack et al., 2001b). Seismic profiles and sediment cores suggest a widespread erosional unconformity extending all the way to the shelf break (Anderson et al., 1992), and Anderson et al. (2002) described a prominent grounding zone sediment wedge above the unconformity on the midshelf. They suggested that the unconformity marked the extension of an Antarctic Peninsula ice sheet, and that it had grounded on the outer continental shelf, some 200 km east of the peninsula, during a recent glaciation (Fig. 2). The age of the unconformity is uncertain since it is not constrained by any available dates (Anderson et al., 2002), but Bentley and Anderson (1998) and Anderson et al. (2002) correlated it to the LGM.

Evidence for Timing of the Deglaciation of the Continental Shelves

The oldest ^{14}C dates constraining minimum ages for ice retreat from the outer and middle shelf areas west of the Antarctic Peninsula come from the Bransfield Strait (Banfield and Anderson, 1995) and the shelf areas west of Anvers Island (Pope and Anderson, 1992; Pudsey et al., 1994). These indicate ice retreat prior to 14–13 ka BP and 12–11 ka BP, respectively (^{14}C dates on organic carbon in diatomaceous mud samples). Most dates from the inner shelf areas and fjords and bays on the peninsula constrain deglaciation to as late as 8–6 ka BP (e.g., Harden et al., 1992; Pudsey et al., 1994; Shevenell et al., 1996; Hjort et al., 2001). According to Harden et al. (1992), glacial-marine sedimentation began in the central part of the Gerlache Strait after 8 ka BP (^{14}C analyses on diatomaceous sediment samples), providing a constraining minimum date for deglaciation and shelf ice retreat in the strait. Similarly, Shevenell et al. (1996) dated the onset of glaciomarine sedimentation in Lallemand Fjord to ca. 8 ka BP (^{14}C analyses on a scaphopod [*Dentalium*] shell). The inner shelf areas around Anvers Island (^{14}C analyses on organic carbon; Pudsey et al., 1994) and in Marguerite Bay (^{14}C analyses on diatomaceous sediment samples; Harden et al. 1992), as well as George VI Sound (^{14}C analyses on barnacle shells; Clapperton and Sugden, 1982; Hjort et al., 2001), were ice free prior to 7–6 ka BP. Domack et al. (2001a) recently published a Late Wisconsin–Holocene chronology for environmental changes in the Palmer Deep (Fig. 1), based on AMS ^{14}C analyses of acid insoluble organic matter and foraminiferal calcite, which showed that the inner shelf areas there might have been deglaciated as early as ca. 11 ka BP.

The low resolution in the deglaciation chronologies from most of the shelf areas makes it currently impossible to recognize more than the general pattern, that the outer and middle shelf areas deglaciated between 14 and 8 ka BP, while most inner shelf areas, fjords, and bays were deglaciated before 8–6 ka BP. Improved chronologies for marine sediment cores, which are currently being developed (Domack et al., 2001a), are the key to better resolution in reconstructions of spatial and temporal patterns of the deglaciation of the Antarctic Peninsula shelf areas. Domack et al. (2001a) recognized a rapid deglacial episode in the Palmer Deep record, characterized by high primary production and iceberg rafting, between ca. 11–10 ka BP. This deglacial event occurs concurrently with the Northern Hemisphere Younger Dryas stadial. In the Palmer Deep record, it was followed by a climate cooling between ca. 10–8 ka BP.

There is as yet no conclusive data to suggest that ice retreat from the LGM positions and subsequent deglaciation of the continental shelf areas were controlled by regional warming. Hence the ice retreat was probably controlled by the global sea-level rise due to meltwater input from the Northern Hemisphere glaciers, causing grounding lines retreat of Antarctic Peninsula glaciers (c.f. Hollin 1962; Stuiver et al. 1981).

Timing of Deglaciation in Presently Ice-Free Terrestrial Coastal Areas

The deglaciation of coastal areas that are presently ice free is constrained by minimum ages obtained from ^{14}C dating of fossil mollusk shells from raised marine deposits, peat deposits in mossbanks on the islands off the Antarctic Peninsula, organics (bulk organics, microbial mats, aquatic mosses, and algal flakes) from lake sediments, and fossil penguin remains (bones, eggshell fragments, squid peaks) from coastal rookeries. It is important to stress that age constraints referred to below are all minimum ages for the deglaciation, since there is an unknown lag time between deglaciation and colonization by plants or animals.

The oldest ^{14}C dates, on fossil mollusks from raised marine deposits and organics from lake sediments, give a minimum age for initial deglaciation on King George Island as 9–8 ka BP (Sugden and John, 1973; Mäusbacher, 1991). A well-dated lithostratigraphic record from northern James Ross Island, where glaciomarine and sublittoral deposits overlie till in coastal sections, constrains deglaciation there as prior to ca. 7.4 ka BP (^{14}C analyses on mollusk shells; Hjort et al., 1997). According to Zale (1994a), Hope Bay was deglaciated prior to *ca.* 6.3 ka BP (chronology based on basal ^{14}C dates from a lake sediment core). In the southern part of the peninsula, ^{14}C dated barnacle fragments from ice-shelf moraines at Two Step Cliffs on Alexander Island (Fig. 1) provide minimum ages for deglaciation and ice-shelf retreat in George VI Sound of 6.5–5.7 ka BP (Clapperton and Sugden, 1982; Hjort et al. 2001).

A number of studies, exploring primarily lake-sediment archives and mossbank peats, suggest that once the glaciers were inside the present coastline, glacial retreat and ice disintegration were slow (Barsch and Mäusbacher, 1986a; Mäusbacher et al., 1989; Zale and Karlén, 1989; Mäusbacher, 1991; Ingólfsson et al., 1992; Björck et al., 1993, 1996a, 1996b; Lopes-Martinez et al., 1996; Hjort et al., 1997): On King George Island, glaciers were at or within their present limits by ca. 6 ka BP (Martinez-Macchiavello et al., 1996), on northern James Ross Island prior to 4.7 ka BP (Hjort et al., 1997), but parts of Byers Peninsula on Livingston Island deglaciated as late as 5–3 ka BP (Björck et al., 1996a). Minimum date for deglaciation on Elephant Island is provided by the onset of moss-bank formation on the steep slopes at Walker Point by 5.5 ka BP (Björck et al., 1991b).

Penguin colonization and location of rookeries is determined by a number of climate-dependent factors, such as availability of ice-free coastal areas suitable for nesting, absence of persistent ice foot, access to open water during the nesting season, and the availability of food (Baroni and Orombelli, 1994). The timing of penguin occupation at coastal sites is therefore a good proxy for minimum age of deglaciation and climate/sea-ice situation similar to the present. Radiocarbon-dated organic remains from fossil penguin rookeries in Marguerite Bay suggest initial colonization in the period 6.5–5.5 ka BP (Emslie, 2001), which confirms minimum ages for deglaciation of 7–6 ka BP provided by marine-geological studies. Zale (1994a, 1994b) studied the history of Adélie penguin occupation at the large Hope Bay rookery. He dated first penguin colonization there at ca. 5.5 ka BP, about 0.8 ka subsequent to the deglaciation of Hope Bay. The oldest dated occupation of penguins at King George Island is 5.8–5.3 ka BP (Barsch and Mäusbacher, 1986b).

It can be concluded that the transition from glacial to interglacial conditions in the Antarctic Peninsula region was broadly completed by ca. 6 ka BP. Here, interglacial conditions are defined by glacier volumes and configurations similar to or less than at present, lake-sediment accumulation in ice-free basins, mossbank growth on the islands off the peninsula, and penguin occupation of coastal rookeries.

Mid-Holocene Glacial Readvances

There are indications from a number of sites for mid-Holocene glacial expansions: Mäusbacher (1991) found evidence of increased glacial activity between 5 and 4 ka BP on King George Island, supporting Sugden and John (1973), who had suggested glacial expansion there after 6 ka BP. Hansom and Flint (1989) documented glacial readvance on Brabant Island some time after 5.3 ka BP, and the post–5.7 ka BP expansion of the George VI Ice Shelf is well documented (Sugden and Clapperton, 1980, 1981; Clapperton and Sugden, 1982; Hjort et al., 2001). A mid-Holocene glacial readvance on northern James Ross Island, initially described by Rabassa (1983), was constrained by further stratigraphical and chronological data to have culminated in at least 7 km of ice advance around 4.6 ka BP (Hjort et al., 1997). Zale (1994b) described a set of distinct moraines in Hope Bay, at the northern tip of the Antarctic Peninsula, which he suggested marked a glacial oscillation at ca. 4.7 ka BP.

Yoon et al. (2000) identified cold waters with extensive sea-ice cover between ca. 6.2 and 4 ka BP, from a multiproxy study of gravity cores from fjord margin sediments on King George Island, which correlates well with Mäusbacher's (1991) suggestion of increased glacial activity at that time. The marine records from the Palmer Deep and Lallemand Fjord do not recognize a mid-Holocene glacial event (Shevenell et al., 1996; Taylor et al., 2000; Domack et al., 2001b; Brachfeld et al., 2002), although Shevanell et al. (1996) identified a mid-Holocene cool event in the Lallemand Fjord record. Hjort et al. (1997) suggested that the glacial advance might be a regional response to increased precipitation, due to warming and increased cyclonic activity in mid-Holocene. Domack et al. (1991) suggested a similar interpretation for mid-Holocene glacial advances in East Antarctica.

Holocene Climatic Optimum

Paleoclimatic records, with high resolution and good chronologies, are scarce from the Antarctic Peninsula region (Fenton, 1980, 1982; Barsch and Mäusbacher, 1986a; Zale and Karlén, 1989; Mäusbacher et al., 1989; Schmidt et al., 1990; Mäusbacher, 1991; Björck et al., 1991b, 1993, 1996a, 1996b; Wasell and Håkansson, 1992; Wasell, 1993; Zale, 1993, 1994; Håkansson et al., 1995; Yang and Hartwood, 1997; Domack et al., 2001a). Björck et al. (1996a) presented a paleoclimatic synthesis for the Holocene development in the Antarctic Peninsula region, based on multivariate analysis of a large and complex data set containing numerous stratigraphic variables in lake sediments and mossbank peats (Fig. 3). Their reconstruction indicates that the climate fluctuated from relatively mild and humid conditions prior to ca. 5 ka BP (characterized by high diversity of spores, aerobic conditions, and complete dominance of allochtonous components [C/N] in lake sediments) to more cold and arid conditions. Around 4.2 ka BP a gradual warming occurred, coupled with increasing humidity. These mild and humid conditions reached an optimum between 4 and 3 ka BP, after which a distinct and rapid climatic deterioration occurred.

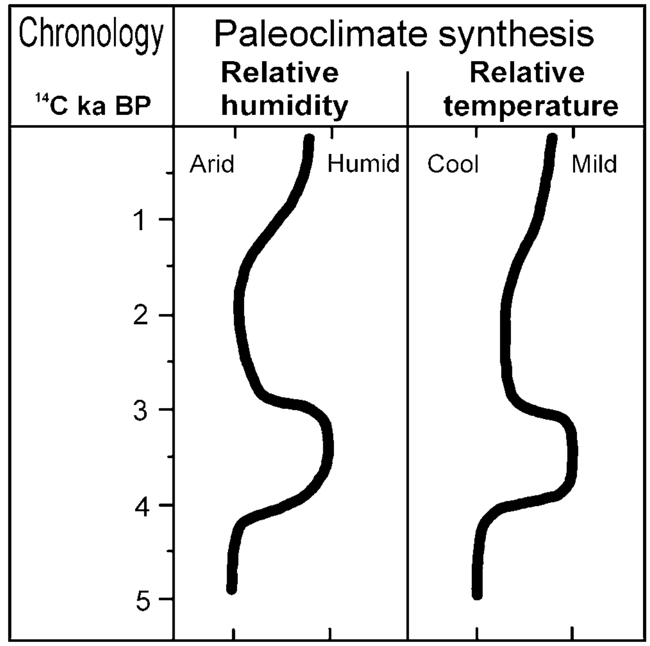

FIGURE 3. Paleoclimatic synthesis for the Antarctic Peninsula region. Modified after Björck et al. (1996a).

The climatic pattern for the last 5 ka is quite similar between the South Shetland Islands, in the maritime climate west of the peninsula, and James Ross Island in the colder and dryer climate in the western Weddell Sea, and Björck et al. (1996a) suggested that this might indicate that the primary factor controlling the climatic variations is the strength of the high-pressure cell over the Antarctic ice sheet.

The marine record from Lallemand Fjord (Fig. 1) recognizes a broad climatic pattern (Shevenell et al., 1996; Domack et al., 1995; Taylor et al., 2001), with an earlier relatively warm period between ca. 7.5–5.8 ka BP and a climatic optimum, reflected by high productivity in the fjord, between 4.2 and 2.7 ka BP. Domack et al. (2001a) defined a more prolonged Holocene climatic optimum in the Palmer Deep record, between ca. 8 and 3 ka BP, characterized by enhanced biological productivity and minimum IRD concentrations. The difference between the terrestrial record and the Palmer Deep record with regard to the duration of the Holocene climate optimum might be a reflection of paucity of terrestrial paleoclimatic data extending beyond 6–5 ka BP, or be partly caused by oscillations in humidity, reflected in the terrestrial record but not picked up by the marine record.

Neoglaciation 3–0, 1 ka BP : Climate Cooling and Increased Ice Volumes

The paleoclimatic synthesis of Björck et al. (1996a) recognized that after ca. 3 ka BP the climate became characterized by cold and dry conditions, which persisted until ca. 1.5 ka BP. Thereafter the climate was somewhat warmer and more humid, but still cold compared to the situation during the climatic optimum.

A number of investigations suggest that glaciers have oscillated and expanded in the Antarctic Peninsula region during the past ca.

2.5 ka (John and Sugden, 1971; John, 1972; Sugden and John, 1973; Curl, 1980; Birkenmajer, 1981; Clapperton and Sugden, 1988; Zale and Karlén, 1989; Clapperton, 1990; López-Martínez et al., 1996). In the South Shetlands, Curl (1980), Birkenmajer (1981), Clapperton and Sugden (1988), and Björck et al. (1996b) found evidence for glacial expansions in the form of readvance moraines covering Holocene raised beaches, and all suggested that this advance had coincided with the Little Ice Age (LIA) glacial expansion in the Northern Hemisphere. Lichenometric dating, using *Rhizocarpon geographicum* thalli, dates the advances to ca. A.D. 1240, 1720, and 1780–1822 (Curl, 1980; Birkenmajer, 1981). Whalebone found on top of a moraine on the front of Rotch Dome on Livingston Island dates the advance there to after ca. 0.3 ka BP, or after ca. A.D. 1650 (Björck et al. 1996b).

Barcena et al. (1998) studied siliceous microfossils from the Bransfield Strait. They found significant increase in sea-ice taxa after 3 ka BP. Fabres et al. (2000) presented a palaeoclimatic sedimentary record from the Bransfield Strait, extending back to almost 3 ka BP. They recognize a distinct neoglacial cooling as well as a LIA cold pulse. The marine record from Lallemand Fjord (Shevenell et al., 1996; Taylor et al., 2000) suggests decreased productivity in the fjord after ca. 3 ka BP due to cooling. Diatom abundances reflect more extensive and seasonally persistent sea ice after ca. 2.7 ka BP (Shevenell et al., 1996) and after ca. 0.4 ka BP an ice-shelf advance into the fjord is documented. Levanter et al. (1996) presented a multiproxy record from the Palmer Deep, and identified both short-term (200 yr) and long-term (2500 yr) cyclicity in the variability of stratigraphic parameters that they interpreted to signify responses to climatic fluctuations. Levanter et al. (1996) and Kirby et al. (1998) suggested, on the basis of organic productivity changes and magnetic lithostratigraphic changes, respectively, that ca. 2500 BP marked the termination of the Holocene climate optimum as reflected in the sediment core record. Domack et al. (2001a) define a neoglacial cooling in the Palmer Deep record, starting at ca. 3 ka BP and lasting until about 100 yr ago. The neoglacial conditions are characterized by greater concentrations of IRD, decreasing sediment accumulation, and decreased biological productivity. The timing of neoglacial onset, as reflected in the Palmer Deep record (Levanter et al., 1996; Kirby et al., 1998; Domack et al., 2001a), agrees well with other marine (Smith et al., 1999) and terrestrial (Björck et al., 1996a; Ingólfsson et al., 1998) records.

A Distinct Warming during the Last 100 Years: Effects and Causes

A number of paleoclimatic records show that climatic development in the western Antarctic Peninsula region has moved from a relatively cold regime to an increasingly warm regime during the past 100 yr (review in Smith et al., 1999). Air-temperature records from the Antarctic Peninsula reveal a distinct and dramatic warming trend of 2–3°C for the past 50 yr (King, 1994, Stark, 1994; Smith et al., 1996; King and Harangozo, 1998). Polar ecosystem research and paleoecological records indicate ecological transitions that have occurred in response to this climate change (Fraser et al., 1992; Emslie, 1995; Trivelpiece and Fraser, 1996; Emslie et al., 1998). A number of the Antarctic Peninsula ice shelves have been retreating during much of the past century, with increasing speed since the late 1980s (e.g., Rott et al., 1996; Vaughan and Doake, 1996; Cooper, 1997; Scambos 2000). Skvarca et al. (1999) estimated the ice-shelf loss to amount to ca. 10,000 km^2 since the mid-1960s. The retreat appears to have been caused by the regional warming, but the mechanisms causing this climate change are poorly understood (Smith et al., 1999). It has been suggested that at least some of the ice shelves were not present earlier in the Holocene and that the recent collapse is probably not unprecedented during the last ca. 10 ka (Hjort et al., 2001; Pudsey and Evans, 2001). Domack et al. (2001b), Pudsey et al. (2001), and Pudsey and Evans (2001) suggested that the Larsen Ice Shelf and the Prince Gustav Channel Ice Shelf, respectively, had expanded in response to enhanced cooling and more persistent sea ice during the last ca. 2.5 ka. Pudsey and Evans (2001) concluded that the recent decay, although probably a response to regional warming, could not be viewed as an unequivocal indicator of anthropogenic climate pertubation.

Existing antarctic meteorological observations may provide some understanding of modern climatic variations. Many meteorological stations were established in 1957 (the International Geophysical Year), with highest density on the Antarctic Peninsula. Simple mean temperature trends such as reported by Vaughan et al. (2001) are therefore biased toward a relatively small part of the continent and provide only limited information about simultaneous continental antarctic air temperature variations. We present spatially interpolated air temperature variations in order to provide insight into the overall geographical pattern of late-20th-century antarctic temperature changes. Most of the homogenized meteorological data used in the analysis were obtained from the NASA Goddard Institute. Interannual surface air temperature variations are known to be substantial in polar regions, so we decided to reduce the influence of such short-term variations and to highlight trends by comparing 5-yr unweighted temperature means centered on 1960 and 1998, using data from 1958–1962 and 1996–2000, respectively. The results of the spatial surface temperature analysis are displayed in the accompanying map series (Fig. 4), showing annual and seasonal changes from 1960 to 1998. Existing meteorological data from islands in the South Atlantic and South Pacific were incorporated into the analysis, although the temperature maps shown here are truncated shortly beyond the Antarctic coast. Ship-based temperature observations were not included in the present analysis due to potential difficulties with such data.

For the winter season (JJA) the net temperature change 1960–1998 has been warming in coastal regions, including the Antarctic Peninsula and little change only in the central Antarctic (Fig. 4). Spring (SON) has experienced net warming in East Antarctic coastal regions, while the interior regions and the Antarctic Peninsula have cooled. Summer (DJF) has only experienced small changes during the observational period. The net result has been slight summer cooling in the Antarctic mainland and more pronounced warming in the Antarctic Peninsula. Except for the Antarctic Peninsula, there has been a general trend toward cooler autumn conditions from 1960 to 1998, especially affecting East Antarctic. The net result has been autumn warming in the Antarctic Peninsula and cooling in the Antarctic mainland.

These seasonal changes have affected the mean annual air temperature (MAAT) in various ways, although the net continental surface temperature change has been small for the period 1960–1998. There has been some warming affecting mainly coastal regions between the Ross Sea and the Antarctic Peninsula, while net cooling has affected interior areas and most of the East Antarctic Plateau.

The data shows that Antarctic surface air temperature records 1960–1998 reveal surface air temperature trends toward cooling in the interior and warming in the coastal regions. Doran et al. (2002) have described ecosystem response in areas affected by this cooling. A more detailed analysis shows the spatial and seasonal patterns of these trends to be complicated and to change with time; that is, the temperature relationship between specific locations is not temporally consistent within Antarctica. A warming trend has persisted within the Antarctic Peninsula during the observational period, with exception of the spring season. The cooling has been modest in coastal East Antarctic regions, but more pronounced near the Amundsen-Scott Base and at the South Pole. By this, an Antarctic Peninsula–Central Antarctic temperature opposition apparently has prevailed during much of the late 20th century. This situation has been difficult to simulate by General Circulation Models (Connolley et al., 1998) and is not yet fully understood (King and Harangonzo, 1998).

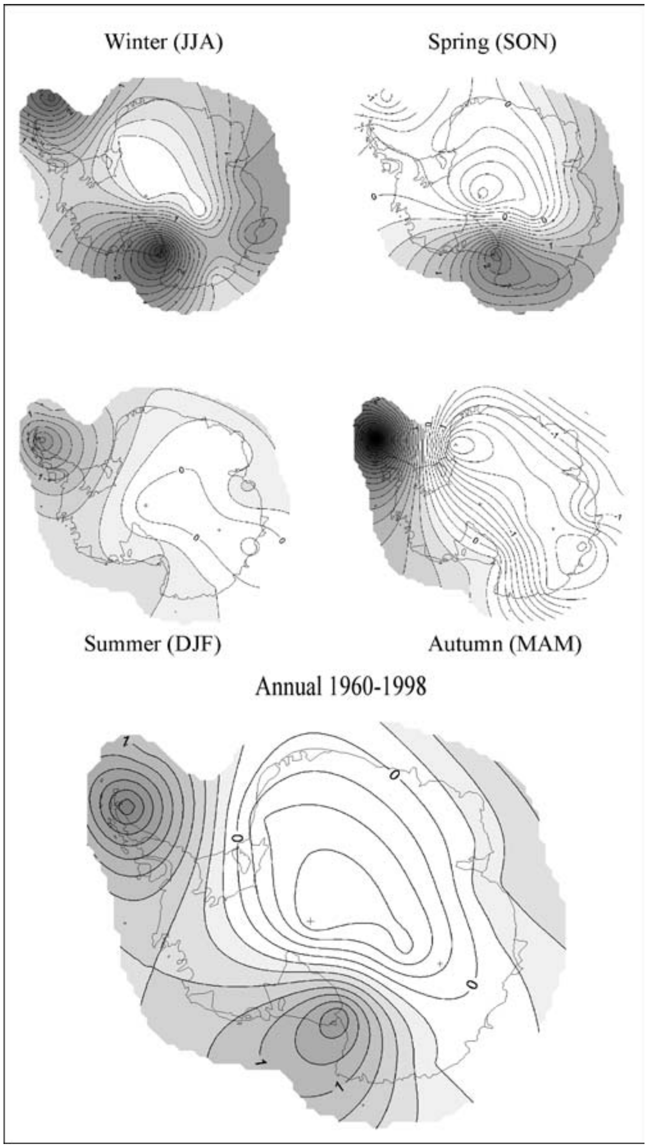

FIGURE 4. Seasonal and annual surface air temperature changes, 1960–1998. Gray indicates temperature increase; white indicates temperature decrease.

The observed spatial pattern of temperature variations, however, suggests that the temperature change may have been caused by a large-scale circulation pattern that exhibits long-term persistence. Van Loon and Williams (1977) linked mean winter surface temperature trends in Antarctica to slow (multiyear) variations in atmospheric long waves, suggesting that midlatitude large-scale circulation plays a significant role in the spatial variability of temperature over the continent. The present cooler conditions in central Antarctica may be associated with stronger zonal westerlies around the Antarctic continent, causing warmer conditions and less sea ice to prevail in the Antarctic Peninsula region (c.f. King and Harangonzo, 1998; Thompson and Solomon, 2002), due to its penetration north into the zone of westerlies.

Summary and Discussion

A reconstruction of the glacial and climatic history of the Antarctic Peninsula since the LGM recognizes the following fairly consistent general patterns (Fig. 5):

1. There is evidence of much more extensive ice cover in the Antarctic Peninsula region than today prior to ca. 14 ka BP. In the central and southern parts, the ice may have been >500 m thicker than today, while in the northern part estimates vary between 150 and 400 m. LGM glaciers from the Antarctic Peninsula generally grounded on the middle-outer shelf, at water depths of ca. 400 m, and were probably fringed by ice shelves extending over the outer shelf and the shelf break.
2. Taken together, the terrestrial and marine geological data on LGM ice extent and configuration can be interpreted to suggest that the LGM Antarctic Peninsula ice sheet was composed of a number of confluent, localized ice caps, domes, and ice stream systems advancing out from the spine of the Antarctic Peninsula toward the middle-outer continental shelf, rather than a concentric and dynamically coherent ice sheet.
3. Deglaciation occurred mainly during the time period >14–6 ka BP. Outer and middle shelf areas deglaciated between 14 and 8 ka BP, while most inner shelf areas, fjords, bays, and presently ice-free coastal land areas deglaciated prior to 8–6 ka BP.
4. The transition from glacial to interglacial conditions in the Antarctic Peninsula region, signaled by ice configuration becoming similar to or less than at present, by the onset of lake-sediment accumulation in ice-free basins, by moss-bank growth on the islands off the peninsula, and by penguin occupation of coastal rookeries, was broadly completed by 6 ka BP.
5. Mid-Holocene glacial- and ice-shelf readvances have been described from a number of sites.
6. Terrestrial paleoclimatic records, based on stratigraphical variables in lake and moss-bank archives, suggest a climate optimum, with relatively warm and humid conditions, between ca. 4 and 3 ka BP. The marine record from Palmer Deep recognizes a longer climate optimum, between ca. 8 and 3 ka BP.
7. Available data suggest that after ca. 3–2.5 ka BP a distinct cooling occurred, and that glaciers and ice shelves expanded during the late Holocene.
8. Air temperature records show a distinct and rapid warming in the Antarctic Peninsula region during the past 50 yr, which may be causing disintegration of some of the Antarctic Peninsula ice shelves.

One interpretation of the data is that Antarctic Peninsula deglaciation and Holocene climate development, prior to the neoglacial cooling after ca. 2.5 ka BP, were significantly lagging the Northern Hemisphere deglaciation and Holocene climate development (Hjort et al., 1998; Ingólfsson et al., 1998; Ingólfsson and Hjort, 1999). The late (ca. 4–2.5 BP) Holocene climate optimum recognized in the terrestrial archives of the Antarctic Peninsula is also recognized as a broadly synchronous environmental event in East Antarctica and Ross Sea/Victoria Land (e.g., Baroni and Orombelli 1994; Cunningham et al., 1999; Hall and Denton, 1999, 2000; Kulbe et al., 2001), which supports the notion of a circum-Antarctic climate optimum. This optimum is not recognized in the Antarctic ice cores. The ice-core records on Holocene climate variability from Byrd, Vostok, and Taylor Dome differ considerably in terms of temperature trends for the past 20 ka and through the Holocene (Blunier et al., 1998; Thompson et al., 1998; Petit et al., 1999; Steig et al., 1998, 2000) and cannot be used as proxies for environmental changes in the coastal areas of the continent. A study by Ciais et al. (1994) on Holocene temperature variations in Antarctica, as expressed in data from 6 ice cores, indicated only subtle temperature variations (±1°C) for the past 10 ka BP. That study places the Antarctic Holocene climate optimum between ca. 10 and 8 ka BP. A study on 11 ice core isotope records for reconstructing Holocene climate variability in Antarctica (Masson et al., 2000) likewise recognized a widespread early Holocene climate optimum (11.5–9 ka cal yr BP: ca. 10–8.5 ka ^{14}C years).

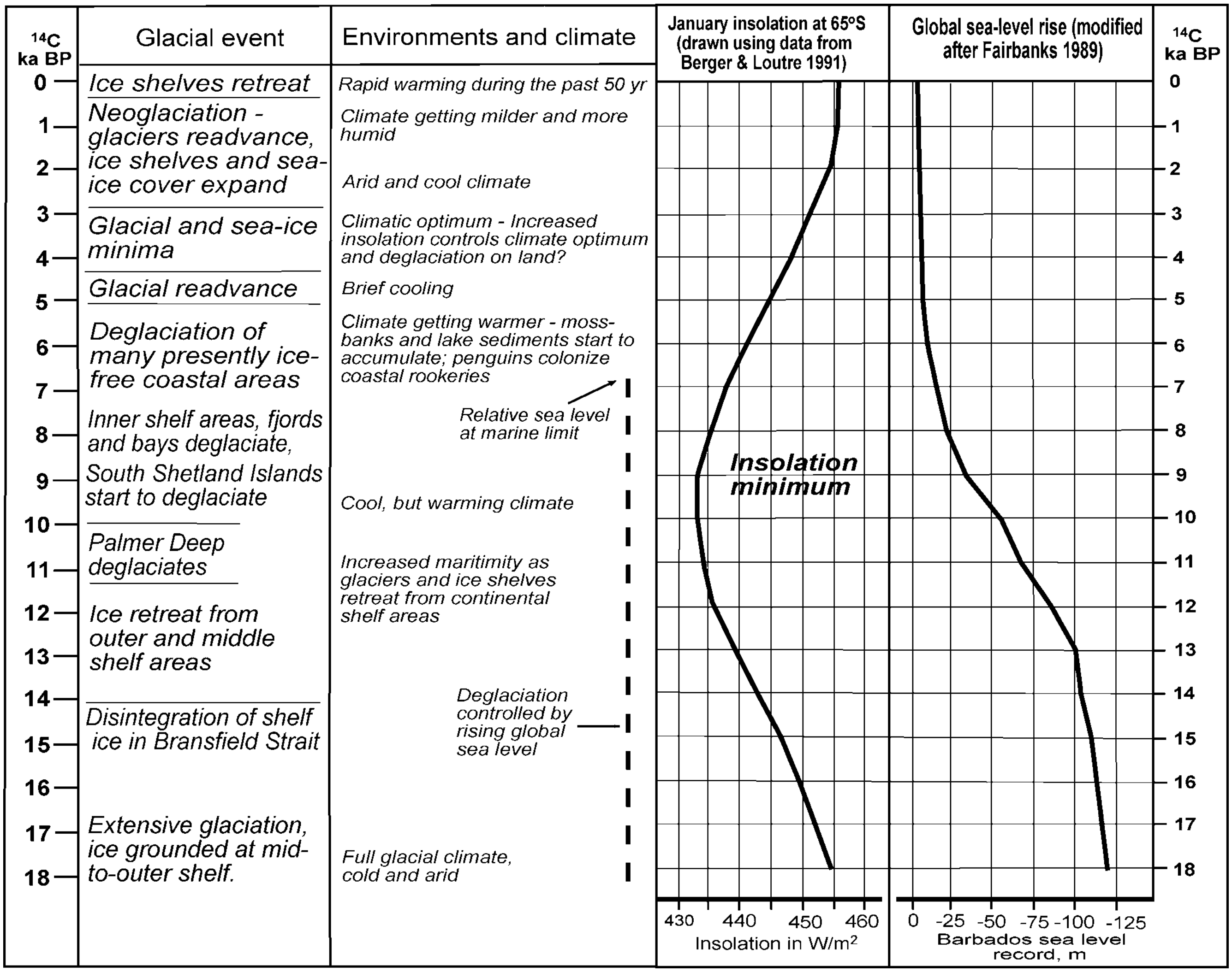

FIGURE 5. A tentative synthesis of glacial and environmental changes in the Antarctic Peninsula region since the LGM.

The large Antarctic system allows for significant regional variability in the Holocene development of climate and glaciation. Ice-core records from the high-altitude inland continental plateau may not capture well signals of environmental change in the peripheral maritime areas of the system. White and Steig (1998) suggested that the inland-plateau ice cores might not be entirely representative of Antarctic climate development, and that records from far more locations are needed to complete the story. The early Holocene ice core temperature optimum (Masson et al., 2000) occurs at a time of summer insolation minimum at 65°S (Berger, 1978; Berger and Loutre, 1991), which probably precludes insolation forcing for the suggested temperature rise. Alternative explanations (Masson et al., 2000) are that the isotope records of Antarctic inland ice cores might be showing responses to changes in local ice-sheet elevations, in connection with rapid sea-level rise and deglaciation of the shelf areas in early Holocene causing drawdown of ice from the plateaus, or decreased summer sea-ice coverage causing the source of precipitation to move closer to the continent, rather than a circum-Antarctic Holocene temperature optimum in early Holocene. Some of the differences between the different records may reflect lag in response to climate forcing between the atmospheric, terrestrial, and oceanic systems.

One of the challenges of paleoclimatic research is to understand the phase relationships between climatic changes in high-latitude Northern and Southern hemispheres and identify mechanisms that can reconcile the different dynamical behavior of glaciers and climate. In order to respond to this challenge, well-dated, high-resolution paleoclimate records from Antarctica are essential. In our opinion, the following should be taken into consideration when developing future research strategies:

1. Defining the LGM glacial situation in both time and space: The extent of ice on the continental shelf areas and timing of the last maximum glaciation is poorly known. We know from a few radiocarbon dates that ice extended out toward the shelf edges some time before 14 ka BP, but it is not known if the Wisconsinan maximum ice extent on the Antarctic Peninsula coincided with the last global glacial maximum 20–18 ka BP.
2. Developing well-constrained deglaciation chronologies for the continental shelf areas: The deglaciation of the shelf areas is constrained by relatively few dates on the western (Pacific) shelf of the peninsula, while a deglaciation chronology for the eastern (Weddell Sea) shelf is still lacking.
3. Better constraining the onshore LGM ice-sheet configuration and thickness, as well as obtaining more high-resolution chronologies for the deglaciation of presently ice-free areas and Holocene environmental changes. Developing programs for cosmogenic exposure dating of erratics and bedrock surfaces can help confine LGM ice thickness in different parts of the Antarctic Peninsula.
4. More data highlighting the relative sea-level history is needed: There exist no well-dated sea-level curves from the region, which can be used for better understanding the geometries of past ice-sheet loads.

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References Cited

Anderson, J. B., Andrews, B. A., Bartek, L. R., and Truswell, E. M., 1991a: Petrology and palynology of glacial sediments: implications for subglacial geology of the eastern Weddell Sea, Antarctica. *In* Thomson, M. R. A., Crame, J. A., and Thomson, J. W. (eds.), *Geological Evolution of Antarctica*. Cambridge: Cambridge University Press, 231–235.

Anderson, J. B., Kennedy, D. S., Smith, M. J., and Domack, E. W., 1991b: Sedimentary facies associated with Antarctica's floating ice masses. *In* Anderson, J. B., and Ashley, G. M. (eds.), Palaeoclimatic interpretation of glacial marine deposits. *Geological Society of America Special Paper,* 256: 1–25.

Anderson, J. B., Shipp, S. S., and Siringan, F. P., 1992: Preliminary seismic stratigraphy of the northwestern Weddell Sea continental shelf. *In* Yoshida, Y., Kaminuma, K., and Shiraishi, K. (eds.), *Recent Progress in Antarctic Earth Sciences.* Tokyo: Terra Scientific Publishing Company, 603–612.

Anderson, J. B., Shipp, S. S., Lowe, A. L., Wellner, J. S., and Mosola, A. B. 2002: The Antarctic Ice Sheet during the Last Glacial Maximum and its subsequent retreat history. *Quaternary Science Reviews,* 21: 49–70.

Aristarain, A. J., Pinglot, J. F., and Pourchet, M., 1987: Accumulation and temperature measurements on the James Ross Island, Antarctic Peninsula, Antarctica. *Journal of Glaciology,* 33: 385–407.

Banfield, L. A., and Anderson, J. B., 1995: Seismic facies investigation of the Late Quaternary glacial history of Bransfield Basin, Antarctica. *Antarctic Research Series,* 68: 123–140.

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

Barsch, D., and Mäusbacher, R., 1986a: Beiträge zur Vergletscherungsgeschichte und zur Reliefentwicklung der Südshetland Insel. *Zeitschrift für Geomorphologie,* 61: 25–37.

Barsch, D., and Mäusbacher, R., 1986b: New data on the relief development of the South Shetland Islands, Antarctica. *Interdisciplinary Science Reviews,* 11: 211–218.

Bart, P. J., and Anderson, J. B., 1996: Seismic expressions of depositional sequences associated with expansion and contraction of ice sheets on the Northwestern Antarctic Peninsula Continental Shelf. *Geological Society Special Publication,* 117: 171–186.

Bentley, M. J., 1999: Volume of Antarctic ice at the Last Glacial Maximum, and its impact on global sea level change. *Quaternary Science Reviews,* 18: 1569–1595.

Bentley, M. J., and Anderson, J. B., 1998: Glacial and marine geological evidence for the ice sheet configuration in the Weddell Sea–Antarctic Peninsula region during the Last Glacial Maximum. *Antarctic Science,* 10: 309–325.

Bentley, M. J., Kubik, P. W., and Sugden, D. E., 2000: Geomorphological evidence and cosmogenic 10Be/26 Al exposure ages for the Last Glacial Maximum configuration of the Antarctic Peninsula Ice Sheet. *In* Hjort, C. (ed.), *Glaciation of the Weddell Sea basin.* ANTIME-Workshop, Abisko, Sweden, 16–19 September 2000. Lund, Department of Quaternary Geology, 5.

Berger, A., 1978: Long-term variations of daily insolation and Quaternary climatic change. *Journal of Atmospheric Sciences,* 35: 2362–2367.

Berger, A., and Loutre, M. F., 1991: Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews,* 10: 297–317.

Berkman, P. A., and Forman, S. L., 1996: Pre-bomb radiocarbon and the reservoir correction for calcareous marine species in the Southern Ocean. *Geophysical Research Letters,* 23: 363–366.

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, Ó., López-Martínez, J., Lyons, W. B., Mabin, M.C. G., Quilty, P. G., Taviani, M., and Yoshida, Y., 1998: Circum-Antarctic coastal environmental shifts during the Late Quaternary reflected by emerged marine deposits. *Antarctic Science,* 10: 345–362.

Birkenmajer, K., 1981: Lichenometric dating of raised marine beaches at Admiralty Bay, King George Island (South Shetland Islands, West Antarctica). *Bulletin de lÁcademie Polonaise des Sciences,* 29: 119–127.

Björck, S., Hjort, C., Ingólfsson, Ó., and Skog, G., 1991a: Radiocarbon dates from the Antarctic Peninsula region—problems and potential. *Quaternary Proceedings,* 1: 55–65.

Björck, S., Malmer, N., Hjort, C., Sandgren, P., Ingólfsson, Ó., Wallén, B., Smith, R. I. L., and Liedberg-Jönsson, B., 1991b: Stratigraphic and paleoclimatic studies of a 5,500 years old moss bank on Elephant Island, Antarctica. *Arctic and Alpine Research,* 23: 361–374.

Björck, S., Håkansson, H., Olsson, S., Barnekow, L., and Jansens, J., 1993: Paleoclimatic studies in South Shetland Islands, Antarctica, based on numerous stratigraphic variables in lake sediments. *Journal of Paleolimnology,* 8: 233–272.

Björck, S., Håkansson, H., Olsson, S., Ellis-Evans, C., Humlum, O., and Lirio, J.M., 1996a: Late Holocene palaeoclimatic records from lake sediments on James Ross Island, Antarctica. *Palaeogeography, Palaeoclimatology, Palaeoecology,* 113: 195–220.

Björck, S., Hjort, C., Ingólfsson, Ó., Zale, R., and Ising, J., 1996b: Holocene deglaciation chronology from lake sediments. *In* López-Martínez, J., Thomson, M. R. A., and Thomson, J. W. (eds.), *Geomorphological Map of Byers Peninsula, Livingston Island.* BAS GEOMAP Series, Sheet 5-A, 1:25 000, with supplementary text, 65 pp. Cambridge, British Antarctic Survey.

Blunier, T., Chappellaz, J., Schwander, J., Dallenbach, A., Stauffer, B., Stocker, T. F., and Raynaud, D., 1998: Asynchrony of Antarctic and Arctic climate change during the last glacial period. *Nature,* 394: 793–743.

Brachfeld, S. A., Banerjee, S. K, Guydo, Y., and Acton, G. D., 2002: A 13,200 year history of century to millennial-scale paleoenvironmental change magnetically recorded in the Palmer Deep, western Antarctic Peninsula. *Earth and Planetary Science Letters,* 194: 311–326.

Canals, M., Urgeles, R., and Calafat, A. M., 2000: Deep sea-floor evidence of past ice streams off the Antarctic Peninsula. *Geology,* 28: 31–34.

Carrara, P., 1979: Former extent of glacial ice in Orville Coast region, Antarctic Peninsula. *Antarctic Journal of the United States,* 14: 45–46.

Carrara, P., 1981: Evidence for a former large ice sheet in the Orville Coast–Ronne Ice Shelf area, Antarctica. *Journal of Glaciology,* 27: 487–491.

Ciais, P., Jouzel, J., Petit, J. R., Lipenkov, V., and White, J. W. C., 1994: Holocene temperature variations inferred from six Antarctic ice cores. *Annals of Glaciology,* 20: 427–436.

Clapperton, C. M., 1990: Quaternary glaciations in the Southern Ocean and Antarctic Peninsula area. *Quaternary Science Reviews,* 9: 229–252.

Clapperton, C. M., and Sugden, D. E., 1982: Late Quaternary glacial history of George VI Sound area, West Antarctica. *Quaternary Research,* 18: 243–267.

Clapperton, C. M., and Sugden, D. E., 1988: Holocene glacier fluctuations in South America and Antarctica. *Quaternary Science Reviews,* 7: 185–198.

Connolley, W. M., and O'Farrell, S. P., 1998: Comparison of warming trends over the last century around Antarctica from three coupled models. *Annals of Glaciology* 27: 565–570.

Cooper, A. P. R., 1997: Historical observations of Prins Gustav Ice Shelf. *Polar Record,* 33: 285–294.

Cunningham, W. L., Leventer, A., Andrews, J. T., Jennings, A. E., and Licht, K. J., 1999: Late Pleistocene-Holocene marine conditions in the Ross Sea, Antarctica: evidence from the diatom record. *The Holocene,* 9: 129–139.

Curl, J. E., 1980: A glacial history of the South Shetland Islands, Antarctica. *Institute of Polar Studies Report*, 63. Columbus: Institute of Polar Studies, Ohio State University. 129 pp.

Denton, G. H., and Marchant, D. R., 2000: The geological basis for a reconstruction of a grounded ice sheet in McMurdo Sound, Antarctica, at the Last Glacial Maximum. *Geografiska Annaler,* 82: 167–211.

Doake, C. S. M., Corr, H. J. F., Rott, H., Skvarca, P., and Young, N. W., 1998: Breakup and conditions for stability of the northern Larsen Ice Shelf, Antarctica. *Nature,* 391: 778–780.

Domack, E. W., Jull, A. J. T., and Nako, S., 1991: Advance of East Antarctic outlet glaciers during the Hypsithermal: implications for the volume state of the Antarctic ice sheet under global warming. *Geology,* 19: 1059–1062.

Domack, E. W., 1992: Modern carbon-14 ages and reservoir corrections for the Antarctic Peninsula and Gerlache Strait area. *Antarctic Journal of the United States,* 27: 63–64.

Domack, E. W., Ishman, S. E., Stein, A. B., McClennen, C. E., and Tull, A. J. T., 1995: Late Holocene advance of the Müller Ice Shelf, Antarctic Peninsula: sedimentological, geochemical and palaeontological evidence. *Antarctic Science,* 7: 159–170.

Domack, E. W., Jakobson, E. A., Shipp, S., and Anderson, J. B., 1999: Sedimentologic and stratigraphic signature of the late Pleistocene/Holocene fluctuations of the West Antarctic Ice Sheet in the Ross Sea: a new perspective. Part 2. *Geological Society of America Bulletin,* 111: 1486–1516.

Domack, E. W., Hall, B. L., and Hayes, J. M., 2000: Accurate Antarctic dating technique sought by Quaternary community. *EOS,* 80(49): 591 and 596.

Domack, E. W., Levanter, A., Dunbar, R., Taylor, F., Brachfield, S., Sjunneskog, C., and ODP Leg 178 Scientific Party, 2001a: Chronology of the Palmer Deep site, Antarctic Peninsula: a Holocene palaeoenvironmental reference for the circum-Antarctica. *The Holocene,* 11: 1–9.

Domack, E. W., Levanter, A., Gilbert, R., Brachfeld, S., Ishman, S., Camerlenghi, A., Gavahan, K., Carlson, D., and Barkoukis, A., 2001b: Cruise reveals history of Holocene Larsen Ice Shelf. *EOS,* 82 (2): 13 and 16–17.

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., and Parsons, A. N., 2002: Antarctic climate cooling and terrestrial ecosystem response. *Nature* 517.

Elliot, D. H., 1981: Glacial geology of Seymour Island. *Antarctic Journal of the United States,* 16(5): 66–67.

Emslie, S. D., 1995: Age and taphonomy of abandoned penguin rookeries in the Antarctic Peninsula. *Polar Record,* 31: 409–418.

Emslie, S. D., 2001: Radiocarbon dated abandoned penguin colonies in the Antarctic Peninsula region. *Antarctic Science,* 13: 289–295.

Emslie, S. D., Fraser, W., Smith, R. C., and Walker, W., 1998: Abandoned penguin colonies and environmental change in the Palmer Station area, Anvers Island, Antarctic Peninsula. *Antarctic Science,* 10: 257–268.

Fabres, J., Calafat, A., Canals, M., Barcena, M. A., and Flores, J. A., 2000: Bransfield Basin fine-grained sediments: late-Holocene sedimentary processes and Antarctic oceanographic conditions. *The Holocene,* 10: 703–718.

Fairbanks, R. G., 1989: A 17,000-year glacio-eustatic sea level record. *Nature,* 342: 637–642.

Fenton, J. H. C., 1980: The rate of peat accumulation in Antarctic moss banks. *Journal of Ecology,* 68: 211–228.

Fenton, J. H. C., 1982: The formation of vertical edges on Antarctic moss peat banks. *Arctic and Alpine Research,* 14: 21–26.

Fraser, W. R., Trivelpiece, W. Z., Ainley, D. G., and Trivelpiece S.G., 1992: Increases in Antarctic penguin populations: Reduced competition with whales or a loss of sea ice due to environmental warming. *Polar Biology,* 11: 525–531.

Gordon, J. E., and Harkness, D. D., 1992: Magnitude and geographic variation of the radiocarbon content in Antarctic marine life: implications for reservoir corrections in radiocarbon dating. *Quaternary Science Reviews,* 11: 697–708.

Håkansson, H., Olsson, S., and Björck, S., 1995: Diatom records and sediment chemistry of a shallow glaciolacustrine basin on James Ross Island, Antarctica. *PACT,* 50: 417–430.

Hall, B. L., and Denton, G. H., 1999: New relative sea-level curves for the southern Scott Coast, Antarctica: evidence for Holocene deglaciation of the Ross Sea. *Journal of Quaternary Science,* 14: 641–650.

Hall, B. L., and Denton G. H., 2000: 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. *Geografiska Annaler,* 82: 305–336.

Hansom, J. D., and Flint, C. P., 1989: Short notes. Holocene ice fluctuations on Brabant Island, Antarctic Peninsula. *Antarctic Science,* 1: 165–166.

Harden, S. L., DeMaster, D. J., and Nittrouer, C. A., 1992: Developing sediment geochronologies for high-latitude continental shelf deposits: a radiochemical approach. *Marine Geology,* 103: 69–97.

Herron, M. J., and Anderson, J. B., 1990: Late Quaternary glacial history of the South Orkney Plateau, Antarctica. *Quaternary Research,* 33: 265–275.

Hiemstra, J. F., 2001: *"Dirt Pictures" reveal the past extent of grounded Antarctic Ice Sheet.* Unpublished Ph.D. dissertation. Amsterdam, University of Amsterdam, 231 pp.

Hjort, C., and Ingólfsson, Ó., 1990: Studies of the Glacial History in the Antarctic Peninsula Area. *In* Karlqvist, A. (ed.), *Swedish Antarctic Research Programme 1988/89—A Cruise Report.* Stockholm: Swedish Polar Research Secretary, 76–80.

Hjort, C., Björck, S., Ingólfsson, Ó., and Möller, P., 1998: Holocene deglaciation and climate history of the northern Antarctic Peninsula region—a discussion of correlations between the Southern and Northern Hemispheres. *Annals of Glaciology,* 27: 110–112.

Hjort, C., Ingólfsson, Ó., Möller, P., and Lirio, J.M., 1997: Holocene glacial history and sea-level changes on James Ross Island, Antarctic Peninsula. *Journal of Quaternary Science,* 12: 259–273.

Hjort, C., Bentley, M. J., and Ingólfsson, Ó., 2001: Holocene and pre-Holocene temporary absence of the George VI Ice Shelf, Antarctic Peninsula. *Antarctic Science,* 13: 296–301.

Hjort, C., Ingólfsson, Ó., Bentley, M. J., and Björck, S. in press: The late Pleistocene and Holocene glacial and climate history of the Antarctic Peninsula region as documented by land and lake sediment records—a review. *In* Domack, E. W., Leventer, A., Burnett, A., Kirby, and Bindschadler, R. (eds.), *Antarctic Peninsula Climate Variability: A Historical and Paleoenvironmental Perspective.* American Geophysical Union, Antarctic Research Series XX: 000–000.

Hollin, J. T., 1962: On the glacial history of Antarctica. *Journal of Glaciology,* 4: 173–195.

Ingólfsson, Ó., Hjort, C., Björck, S., and Smith, R. I. L., 1992: Late Pleistocene and Holocene glacial history of James Ross Island, Antarctic Peninsula. *Boreas,* 21: 209–222.

Ingólfsson, Ó., Hjort, C., Berkman, P., Björck, S., Colhoun, E., Goodwin, I. D., Hall, B., Hirakawa, K., Melles, M., Möller, P., and Prentice, M., 1998: Antarctic glacial history since the Last Glacial Maximum: an overview of the record on land. *Antarctic Science,* 10: 326–344.

Ingólfsson, Ó., and Hjort, C., 1999: The Antarctic contribution to Holocene global sea-level rise. *Polar Research,* 18: 323–330.

John, B. S., 1972: Evidence from the South Shetland Islands toward a glacial history of West Antarctica. *Institute of British Geographers, Special Publication,* 4: 75–89.

John, B. S., and Sugden, D. E., 1971: Raised marine features and phases of glaciation in the South Shetland Islands. *British Antarctic Survey Bulletin,* 24: 45–111.

Jones, V. I., Hodgson, D. A., and Chepow-Lusty, A., 2000: Palaeolimnological evidence for marked Holocene environmental changes on Signy Island, Antarctica. *The Holocene,* 10: 43–60.

Kellogg, T. B., Hughes, T., and Kellogg, D. E., 1996: Late Pleistocene interactions of East and West Antarctic ice-flow regimes:

evidence from the McMurdo Ice Shelf. *Journal of Glaciology,* 42: 486–500.

Kennedy, D. S., and Anderson, J. B., 1989: Glacial-marine sedimentation and Quaternary glacial history of Marguerite Bay, Antarctic Peninsula. *Quaternary Research,* 31: 255–276.

King, J. C., 1994: Recent climate variability in the vicinity of the Antarctic Peninsula. *International Journal of Climatology,* 14: 357–369.

King, J. C., and Harangozo, S. A., 1998: Climate change in the western Antarctic Peninsula since 1945: observations and possible causes. *Annals of Glaciology,* 27, 571–575

Kirby, M. E., Domack, E. W., and McClennen, C. E., 1998: Magnetic stratigraphy and sedimentology of Holocene glacial marine deposits in the Palmer Deep, Bellingshausen Sea, Antarctica: implications for climatic change. *Marine Geology,* 152: 247–259.

Kulbe, T., Melles, M., Verkulich, S. R., and Pushina, Z. V., 2001: East Antarctic climate and environmental variability over the last 9400 years inferred from marine sediments in the Bunger Oasis. *Arctic, Antarctic, and Alpine Research,* 33: 223–230.

Larter, R. D., and Barker, P. F., 1991: Neogene interaction of tectonic and glacial processes at the Pacific margin of the Antarctic Peninsula. *In* Macdonald, D. I. M. (ed.), *Sedimentation, Tectonics and Eustasy: International Association of Sedimentologists Special Publication,* 12, 165–186.

Larter, R. D., and Vanneste, L. E., 1995: Relict subglacial deltas on the Antarctic Peninsula outer shelf. *Geology,* 23: 33–36.

Levanter, A., Domack, E. W., Ishman, S. E., Brachfeld, S., McClennen, C. E., and 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. *Geological Society of America Bulletin,* 108: 1642–1644.

Licht, K. J., Jennings, A. E., Andrews, J. T., and Williams, K. M., 1996: Chronology of Late Wisconsin ice retreat from the western Ross Sea, Antarctica. *Geology,* 24: 223–226.

López-Martínez, J., Thomson, M. R. A., Arche, A., Björck, S., Ellis-Evans, J. C., Hathway, B., Hernández-Cifuentes, F., Hjort, C., Ingólfsson, Ó., Ising, J., Lomas, S., Martínez de Pisón, E., Serrano, E., Zale, R., and King, S., 1996: *Geomorphological Map of Byers Peninsula, Livingston Island.* BAS GEOMAP Series, Sheet 5-A, 1:25,000, with supplementary text. Cambridge, British Antarctic Survey, 65 pp.

Martin, P. J., and Peel, D. A., 1978. The spatial distribution of 10 m temperatures in the Antarctic Peninsula. *Journal of Glaciology,* 20: 311–317.

Mäusbacher, R., 1991: Die Jungkvartäre Relief- und Klimageschichte im Bereich der Fildeshalbinsel, Süd-Shetland-Inseln, Antarktis. *Heidelberger Geographische Arbeiten,* 89: 207 pp.

Mäusbacher, R., Müller, J., and Schmidt, R., 1989: Evolution of postglacial sedimentation in Antarctic lakes. *Zeitschrift für Geomorphologie,* 33: 219–234.

Martinez-Macchiavello, J. C., Tatur, A., Servant-Vildary, S., and Del Valle, R., 1996. Holocene environmental change in a marine-estuarine-lacustrine sediment sequence, King George Island, South Shetland Islands. *Antarctic Science,* 8: 313–322.

Oppenheimer, M., 1998: Global warming and the stability of the West Antarctic Ice Sheet. *Nature,* 393: 325–332.

Pope, P. G., and Anderson, J. B., 1992: Late Quaternary glacial history of the northern Antarctic Peninsula's western continental shelf: evidence from the marine record. *American Geophysical Union, Antarctic Research Series,* 57: 63–91.

Petit, J. R., Jouzel, J., Raynaud, D., Barkov, N.I., Barnola, J.-M., Basile, I., Bender, M., and Chappallaz, J., 1999: Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature,* 399: 429–436.

Pudsey, C. J., Barker, P. F., and Larter, R. D., 1994: Ice sheet retreat from the Antarctic Peninsula Shelf. *Continental Shelf Research,* 14: 1647–1675.

Pudsey, C. J., Domack, E. W., Evans, J., Morris, P., and Del Valle, R. A., 2001: Bathymetry and acoustic facies beneath the former Larsen-a and Prince Gustav Ice Shelves, NW Weddell Sea. *Antarctic Science,* 13: 312–322.

Pudsey, C. J., and Evans, J., 2001. First survey of Antarctic sub-ice shelf sediments reveals mid-Holocene ice shelf retreat. *Geology,* 29: 787–790.

Rabassa, J., 1983: Stratigraphy of the glacigenic deposits in northern James Ross Island, Antarctic Peninsula. *In* Evenson, E., Schlüchter, C., and Rabassa, J. (eds.), *Tills and Related Deposits.* Rotterdam: A. A. Balkema Publishers, 329–340.

Robin, G deQ., and Adie, R. J., 1964. The ice cover. *In* Priestley, R. E., Adie, R. J., and Robin, G. (eds.), *Antarctic Research.* London: Butterworth, 100–117.

Rott, H., Rack, W., Nagler, T., and Skvarca, P., 1998: Climatically induced retreat and collapse of northern Larsen Ice Shelf, Antarctic Peninsula. *Annals of Glaciology,* 27: 86–92.

Scambos, T. A., 2000: The link between climate warming and break-up of ice shelves in the Antarctic Peninsula. *Journal of Glaciology,* 46: 516–530.

Schmidt, R., Mäusbacher, R., and Müller, J., 1990: Holocene diatom flora and stratigraphy from sediment cores of two Antarctic lakes (King George Island). *Journal of Paleolimnology,* 3: 55–74.

Shevenell, A. E., Domack, E. W., and Kernan, G. M., 1996: Record of Holocene palaeoclimate change along the Antarctic Peninsula: evidence from glacial marine sediments, Lallemand Fjord. *Papers and Proceedings of the Royal Society of Tasmania,* 130: 55–64.

Skvarca, P., Rack, W., Rott, H., and Donangelo, T.L., 1999: Climatic trends and the retreat and disintegration of ice shelves on the Antarctic Peninsula: an overview. *Polar Research,* 18: 151–157.

Sloan, B. J., Lawer, L. A., and Anderson, J. B., 1995: Seismic stratigraphy of the Palmer Basin. *Antarctic Research Series,* 68: 235–260.

Smith, R. C., Stammerjohn, S. E., and Baker, K.S., 1996: Surface air temperature variations in the western Antarctic Peninsula region. *In* Ross, R. M., Hoffman, E. E., and Quetin, L. B. (eds.), *Foundations for Ecological Research West of the Antarctic Peninsula.* Washington, American Geophysical Union, 105–121.

Smith, R. C., Ainley, D., Baker, K., Domack, E., Emslie, S., Fraser, S., Kennett, Levanter, A., Mosley-Thompson, E., Stammerjohn, S., and Vernet, M., 1999: Marine ecosystem sensitivity to climate change. *BioScience,* 49 (5): 393–404.

Stark, P., 1994: Climatic warming in the central Antarctic Peninsula area. *Weather,* 49: 215–220.

Steig, E. J., Brook, E. J., White, J. M. A., Sucher, C. M., Bender, M. L., Lehman, S. J., Morse, D. L., Waddington, E. D., and Clow, G.D., 1998: Synchronous climate changes in Antarctica and the North Atlantic. *Science,* 282: 92–95.

Steig, E. J., Morse, D. L., Waddington, E. D., Stuiver, M, Grootes, P. M., Mayewski, P. A., Twickler, M. S., and Whitlow, S. I., 2000: Wisconsinan and Holocene climate history from an ice core at Taylor Dome, western Ross Embayment, Antarctica. *Geografiska Annaler,* 82A(2–3): 213–232.

Stuiver, M., Denton, G. H., Hughes, T. J., and Fastook, J.L., 1981: History of the marine ice sheet in West Antarctica during the last deglaciation: A working hypothesis. *In* Denton, G. H., and Hughes, T. J. (eds.), *The Last Great Ice Sheets.* New York: John Wiley and Sons, 319–436.

Sugden, D. E., and Clapperton, C. M., 1980: West Antarctic Ice Sheet fluctuations in the Antarctic Peninsula area. *Nature,* 286: 378–381.

Sugden, D. E., and Clapperton, C. M., 1981: An ice-shelf moraine, George IV Sound, Antarctica. *Annals of Glaciology,* 2: 135–141.

Sugden, D. E., and John, B. S., 1973: The ages of glacier fluctuations in the South Shetland Islands, Antarctica. *In* van Zinderen Bakker, E. M. (ed.), *Palaeoecology of Africa, the Surrounding Islands and Antarctica*, 8. Cape Town: Balkema, 141–159.

Taylor, F., Whitehead, J., and Domack, E., 2000: Holocene paleoclimate change in the Antarctic Peninsula: evidence from diatom, sedimentary and geochemical record. *Marine Micropaleontology,* 41: 25–43.

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

Thompson, L. G., Davis, M. E., Mosley-Thompson, E., Sowers, T. A., Henderson, K. A., Zagorodnov, V. S., Lin, P. N., Mikhalenko,

V. N., Campen, R. K., Bolzan, J. F., Cole-Dai, J., and Francou, B., 1998: A 25,000-year tropical climate history from Bolivian ice cores. *Science,* 282: 1858–1864.
Trivelpiece, W. Z., and Fraser, W. R., 1996: The breeding biology and distribution of Adélie penguins: Adaptations to environmental variability. *In* Ross, R. M., Hoffman, E. E., and Quetin, L. B. (eds.), *Foundations for Ecological Research West of the Antarctic Peninsula.* Washington: American Geophysical Union, 273–285.
Van Loon, H., and Williams, J., 1977. The connection between trends of mean temperature and circulation at the surface: Part IV. Comparison of the surface changes in the Northern Hemisphere with the upper air and with the Antarctic in winter. *Monthly Weather Review,* 105: 636–647.
Vaughan, D. G., and Doake, C. S. M., 1996: Recent atmospheric warming and retreat of ice shelves on the Antarctic Peninsula. *Nature,* 379: 328–331.
Vaughan, D. G., Marshall, G. J., Connolley, W. M., King, J. C., and Mulvaney, R., 2001: Devil in the Detail. *Science* 293: 1777–1778.
Waitt, R. B., 1983: Thicker West Antarctic ice sheet and peninsula ice cap in late-Wisconsin time—sparse evidence from northern Lassiter Coast. *Antarctic Journal of the United States,* 18: 91–93.
Wasell, A., 1993: Diatom stratigraphy and evidence of Holocene environmental changes in selected lake basins in the Antarctic and South Georgia. Stockholm University: *Department of Quaternary Geology Research Reports,* 23: 1–15.
Wasell, A., and Håkansson, H., 1992: Diatom stratigraphy in a lake on Horseshoe Island, Antarctica: a marine-brackish-fresh water transition, with comments on the systematics and ecology of the most common diatoms. *Diatom Research,* 7: 157–194.
White, W. C., and Steig, E. J., 1998: Timing is everything in a game of two hemispheres. *Nature,* 394: 717–718.
Yang, S. Y., and Harwood, D. M., 1997: Late Quaternary environmental fluctuations based on diatoms from Yanou Lake, King George Island, Fildes Peninsula, Antarctica. *In* Ricci, C. A. (ed.), *The Antarctic Region: Geological Evolution and Processes.* Siena: Universtà degli Studi di Siena, 853–859.
Yoon, H. I., Park, B. K., Kim, Y., and Kim, D., 2000: Glaciomarine sedimentation and its palaeoceanographic implications along the fjord margins in the South Shetland Islands, Antarctica, during the last 6000 years. *Palaeogeography, Palaeoclimatology, Palaeoecology,* 157: 189–211.
Zale, R., 1993: Lake sediments around the Antarctic Peninsula, archives of climatic and environmental change. *GERUM Naturgeogr.,* 17: 1–21.
Zale, R., 1994a: Changes in size of the Hope Bay Adélie penguin rookery as inferred from Lake Boeckella sediment. *Ecography,* 17: 297–304.
Zale, R., 1994b: ^{14}C age correction in Antarctic lake sediments inferred from geochemistry. *Radiocarbon,* 36: 173–185.
Zale, R., and Karlén, W., 1989: Lake sediment cores from the Antarctic Peninsula and surrounding islands. *Geografiska Annaler,* 71(A): 211–220.

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