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Antarctic glacial history since the Last Glacial Maximum: an overview of the record on land

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Abstract: This overview examines available circum-Antarctic glacial history archives on land, related to developments after the Last Glacial Maximum (LGM). It considers the glacial-stratigraphic and morphologic records and also biostratigraphical information from moss banks, lake sediments and penguin rookeries, with some reference to relevant glacial marine records. It is concluded that Holocene environmental development in Antarctica differed from that in the Northern Hemisphere. The initial deglaciation of the shelf areas surrounding Antarctica took place before 10 000 ^{14}C yrs before present (BP), and was controlled by rising global sea level. This was followed by the deglaciation of some presently ice-free inner shelf and land areas between 10 000 and 8000 yr BP. Continued deglaciation occurred gradually between 8000 yr BP and 5000 yr BP. Mid-Holocene glacial readvances are recorded from various sites around Antarctica. There are strong indications of a circum-Antarctic climate warmer than today 4700–2000 yr BP. The best dated records from the Antarctic Peninsula and coastal Victoria Land suggest climatic optimums there from 4000–3000 yr BP and 3600–2600 yr BP, respectively. Thereafter Neoglacial readvances are recorded. Relatively limited glacial expansions in Antarctica during the past few hundred years correlate with the Little Ice Age in the Northern Hemisphere.

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Introduction

The Antarctic Ice Sheet, presently containing 25–30 million km^3 of ice (Lovering & Prescott 1979, Drewry *et al.* 1982), is the worlds' largest glacial system and has existed intermittently for 30–40 million years, since the mid-Tertiary (Hambrey *et al.* 1989, Birkenmajer 1987, 1991, Barrett *et al.* 1991). It has been suggested that the Antarctic Ice Sheet has existed close to its present stable configuration for the past 14 million years (Shackleton & Kennett 1975, Sugden *et al.* 1993), although parts of the ice sheet have fluctuated substantially during the Quaternary.

The influences of Antarctic Ice Sheet fluctuations in the Quaternary history of global climate are not yet well understood. More than 98% of the Antarctic continent is today covered by glacier ice, and the potential on land for obtaining high-resolution geological data pertaining to its glacial history is poor. A fundamental question, given the suggested long-term stable glacial system in Antarctica, is what caused the glacial fluctuations observed in the records? Antarctic glaciers respond both to global sea level changes, mainly driven by Northern Hemisphere glacial fluctuations, and to Southern Hemisphere climate changes. A good understanding of the Late Quaternary glacial and climate history of Antarctica will also constrain the contribution of Antarctic ice to the global sea-level and marine oxygen-isotope records, and is important for understanding the relative timing of climate changes between the polar hemispheres (Denton *et al.* 1989, Clapperton & Sugden 1990, Andrews 1992, Colhoun *et al.* 1992, Moriwaki *et al.* 1992, Quilty 1992).

Studies of Late Quaternary climate changes in Antarctica have been focused on ice-core and marine records, as a consequence of the scarcity of chronologically well constrained geological data on land. However, the last two decades have seen increasingly more sophisticated data from ice free areas in Antarctica (Fig. 1), based on glacial stratigraphical and morphological investigations, studies of

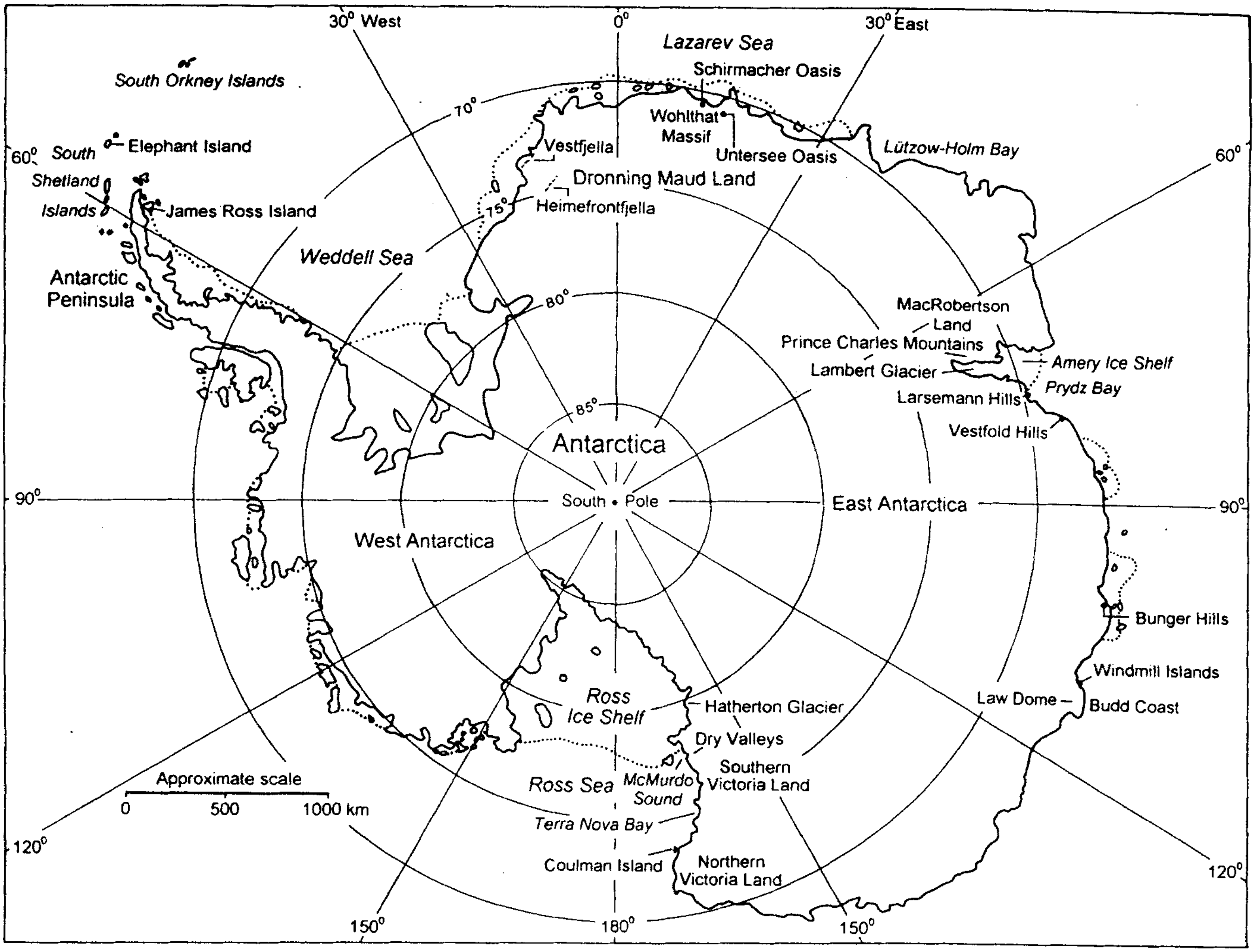

Fig. 1. Map of Antarctica showing the locations of the ice-free areas discussed in the text.

lake sediment and moss-bank cores, and of fossil penguin rookeries. Our knowledge primarily concerns the "postglacial" developments, i.e. from the Late Wisconsinan and Holocene, because those geological archives, although few and far between, are the best preserved.

The purpose of this paper is to review the glacial history in Antarctica since the Last Glacial Maximum (LGM), as seen through the geological data on land. The focus of the paper is on three major aspects:

a) Can a circum-Antarctic glacial history pattern be identified?

b) What does the record imply about the control of glaciation by sea-level and by climate?

c) What is the relative timing of the deglaciation in Antarctica and of the Northern Hemisphere large ice sheets?

Chronological control

The age control on glacial and climatic events since the LGM in Antarctica is primarily through ^{14}C dating. Marine materials like mollusc shells, marine mammal bones, penguin remains (bones, guano-debris) and snow petrel stomach oil or nest deposits are one source of datable materials. Terrestrial and lacustrine materials like mosses, lake sediment bulk samples, microbial mats, aquatic mosses and algal flakes have also been widely used for constraining environmental changes in time. Both marine and terrestrial/lacustrine materials often yield ages that appear too old in comparison with the conventional terrestrial-based radiocarbon time-scale (Björck *et al.* 1991a, Gordon & Harkness 1992). A requirement for understanding the dynamics of the Antarctic glacial system is to have a reliable chronology for both terrestrial and marine materials.

Dating terrestrial materials

Radiocarbon dates on terrestrial materials are mainly from peat deposits, primarily mosses, in moss-banks on the islands off the Antarctic Peninsula (e.g. Fenton 1980, Birkenmajer *et al.* 1985, Björck *et al.* 1991a, 1991b) and bulk sediments, microbial mats, aquatic mosses and algal flakes from Antarctic lake sediments and deltas (e.g. Stuiver *et al.* 1981, Pickard & Seppelt 1984, Pickard *et al.* 1986, Zale & Karlén 1989, Mäusbacher *et al.* 1989, Schmidt *et al.* 1990, Ingólfsson *et al.* 1992, Björck *et al.* 1993, 1996a).

Samples of moss-bank peat on Elephant Island (Fig. 1) have been found to give some of the most reliable ^{14}C ages in Antarctica (Björck *et al.* 1991a, 1991b), and are thus optimal for constructing a ^{14}C chronology for environmental changes in the Antarctic Peninsula region for the past 5000 years. There is a balance between atmospheric carbon content and the intake of carbon by the mosses; old groundwater or carbon from the bedrock will not influence the carbon content of the mosses and contamination by down-growth of roots from plants living on the surface is minimal. There are difficulties in correlating the peat sequences to other archives (glacial stratigraphical sections and lake sediment sequences), but Björck *et al.* (1991c, 1991d) were able to correlate between moss-bank deposits and lake sediments in the Antarctic Peninsula region, using tephrostratigraphy, thus gaining an important chronological control for the lake sediment archives.

There are a number of sources of contamination when dating bulk sediments, microbial mats, aquatic mosses or algal flakes from Antarctic lake basins, often causing ages that are too old (e.g. Adamson & Pickard 1986, Stuiver *et al.* 1981, Squyres *et al.* 1991, Björck *et al.* 1991a, Melles *et al.* 1994, Zale 1994):

a) Old groundwater or an input of glacial meltwater depleted in ^{14}C, contaminating the submerged flora, can be a serious problem. Adamson & Pickard (1986) and Stuiver *et al.* (1981) found that the correction needed for reservoir effects in freshwater algae is 450–700 yrs.

b) Reduced gas exchange with the atmosphere due to a long (in extreme cases perennial or decadal) duration of the ice cover in lake and marine environments may lead to much older radiocarbon ages. This effect was described by Weiss *et al.* (1979) for the Weddell Sea and by Melles *et al.* (1997) for the marine basins (epishelf lakes) of Bunger Hills (Fig. 1). In the latter paper, modern reservoir effects of more than 2000 years were estimated, considerably higher than the marine reservoir effect of 1300 years estimated for the Vestfold Hills area (Adamson & Pickard 1986).

c) Contamination by the marine reservoir effect through input from sea mammals and birds to lake basins (Björck *et al.* 1991a, Zale 1994).

d) Supply of old carbon from soils or weathered carbon-bearing rocks. Stuiver *et al.* (1981) described two ^{14}C dates from the same delta bed in southern Victoria Land, one from terrestrial algae, the other from a well preserved valve of the scallop *Adamussium colbecki*. The shell date was 5050 ± 50 yr BP (corrected by 1300 ^{14}C yr for marine reservoir age) and the algae 5930 ± 200 yr BP, which Stuiver *et al.* (1981) thought could reflect contamination by carbonate from local marble bedrock.

e) Continuous erosion of lake bottom surface sediments due to bottom-freezing in winter, or oxidation of these surface sediments during periods of desiccation, are processes which could lead to erroneous ^{14}C dates (Björck *et al.* 1991a).

f) Recycling of old carbon in stratified Antarctic lakes (Squyres *et al.* 1991).

g) Longevity of organisms may also play a role. Some Antarctic freshwater and terrestrial algae can survive long periods of desiccation and repeated freeze-thaw cycles (Vincent *et al.* 1993). Cryptoendolithic algae in suitable rock types in Antarctica are thought to have very slow turnover times, on the order of 10 000 to 17 000 yrs (Nienow & Friedmann 1993, Johnson & Vestal 1991). These algae can be released to the ground when the rock erodes and then be blown or washed into other stratigraphic archives.

Radiocarbon dates for organic remains (microbial mats, algae, water mosses) in sediment cores sampled from the present Lake Hoare in Taylor Valley, McMurdo Dry Valleys, southern Victoria Land (Fig. 1), revealed that there can be large contamination problems (Squyres *et al.* 1991). These were expressed as very old ages of surface sediments (varying between 2000 and 6000 ^{14}C yrs) and samples obtained from depth in the cores yielding ages similar or younger than the surface material. Squyres *et al.* (1991) concluded that ^{14}C dates of Lake Hoare sediments were of limited value because of the high degree of contamination, and pointed out that probably the source carbon, which the organisms fix, is old, and that relatively long-term recycling of carbon in the lake could contribute to old apparent ^{14}C dates. A similar explanation is possible for ^{14}C ages of 24 000 and 35 700 yr BP from the base of marine inlet and lake cores in Bunger Hills, East Antarctica (Melles *et al.* 1997). In the fresh-water Lake Untersee (Dronning Maud Land, East Antarctica (Fig. 1)), a thick, perennial lake ice cover probably leads to a reservoir effect on radiocarbon dates as high as 11 000 years (M. Schwab, personal communication 1998).

Björck *et al.* (1991a, 1991d) concluded that the causes of erroneous ages often seem to be a combination of different contamination sources and processes and that great caution is needed when ^{14}C dates on Antarctic lacustrine samples are interpreted and evaluated. A primary control for the reliability of the dates is the stratigraphic consistency in the dated sequence. In addition, the $\partial^{13}C$-value should always be measured and used to correct the $^{14}C/^{12}C$ relationship. In some cases, the reported age could be incorrect by hundreds of years without such a correction (Björck *et al.* 1991a). Dates on aquatic moss samples, extracted from the bulk sediments, appear to be more reliable than dates on the bulk sediments themselves, or on algae. Björck *et al.* (1991d) found that out of 14 radiocarbon dates from a lake basin on Livingston Island, in the South Shetland Islands, only three determinations could be judged reliable, after controlling the dates by tephrochronological cross-correlations. Two of these were on aquatic mosses.

Dating marine materials

Radiocarbon concentration in the Southern Ocean is dominated by the upwelling of deep water from the Northern Hemisphere at the Antarctic Divergence. Deepwater is depleted in ^{14}C, and although mixing with "younger" surface water south of the Antarctic Convergence occurs, marine species which live in those waters have apparent radiocarbon ages that are older than 1000–1200 yrs (Broecker 1963, Björck *et al.* 1991a, Gordon & Harkness 1992). Other factors which influence spatial and temporal variability in the Antarctic radiocarbon reservoir are inputs of ^{14}C depleted CO_2 from melting ice, regional differences in the upwelling around Antarctica, perennial sea ice cover and local freshwater inputs into nearshore marine basins (Omoto 1983, Domack *et al.* 1989, Melles *et al.* 1994, Melles *et al.* 1997).

In the geological literature on Antarctica, different authors have taken different approaches to the marine reservoir correction. For example, Sugden & John (1973), Clapperton & Sugden (1982, 1988), Payne *et al.* (1989), Hansom & Flint (1989) and Clapperton (1990) subtracted 750 years from their Antarctic Peninsula radiocarbon dates, whereas Barsch & Mäusbacher (1986), working on the South Shetland Islands (Fig. 1) used an envelope of 850–1300 yrs. Ingólfsson *et al.* (1992) and Hjort *et al.* (1997) applied a sea correction of 1200 yrs, while Pudsey *et al.* (1994) used a reservoir correction of 1500 yrs. In East Antarctica, Adamson & Pickard (1986), Colhoun & Adamson (1992a) and Fitzsimons & Colhoun (1995) used a reservoir correction of 1300 years when dealing with the Late Quaternary glacial history in the Vestfold Hills and Bunger Hills areas in East Antarctica, while Hayashi & Yoshida (1994), working in the Lützow-Holm Bay area (Fig. 1), suggested a correction of 1100 years. Verkulich & Hiller (1994) ^{14}C dated stomach oil deposits in snow petrel colonies in Bunger Hills. They based their chronology for petrel colonization on conventional ^{14}C dates, but stated that a reservoir correction of 1300 years probably was appropriate. In the Victoria Land/Ross Sea area (Fig. 1), Stuiver *et al.* (1981) and Denton *et al.* (1989) based their chronology on uncorrected ^{14}C dates. Likewise, Baroni & Orombelli (1991) used conventional dates for their deglaciation chronology for Terra Nova Bay, but calibrated the conventional ages when bracketing a relative sea level curve for the area. Baroni & Orombelli (1994a) presented both uncorrected conventional and calibrated ^{14}C chronologies when dealing with the Holocene environmental history of Victoria Land, but Baroni & Orombelli (1994b) based their chronology of Holocene glacier variations in Terra Nova Bay on calibrated ^{14}C dates. Colhoun *et al.* (1992) used a correction of 1090 years for mollusc dates from the Ross Sea area, while Licht *et al.* (1996) used a reservoir correction of 1200 years for dates from the same area.

A number of investigations have assessed which correction of ^{14}C ages of Antarctic marine organisms has to be applied in order to establish a coherent radiocarbon chronology for Late Wisconsinan–Holocene glacial events. Circum-Antarctic studies generally show the average correction for reservoir age of marine mollusc shells to be 1100–1400 ^{14}C yr (Yoshida & Moriwaki 1979, Stuiver *et al.* 1981, Adamson & Pickard 1986, Björck *et al.* 1991a, Gordon & Harkness 1992, Domack 1992, Berkman 1994, Berkman & Forman 1996). Studies of pre-bomb seal, whale and penguin samples have yielded greater variability than the pre-bomb marine mollusc shells, ranging between 915 ± 75 and 1760 ± 55 yr (Curl 1980, Mabin 1985, Whitehouse *et al.* 1987, 1989, Baroni & Orombelli 1991, Gordon & Harkness 1992), suggesting that longevity and ecology of different species, as well as what tissue (flesh, bone, feathers, guano) is dated can significantly influence the correction factor needed (Mabin 1986, Baroni & Orombelli 1991.) Berkman & Forman (1996) suggested that reservoir corrections of 1300 ± 100 yr (molluscs), 1424 ± 200 yr (seals) and 1130 ± 130 yr (penguins) should be applied to Antarctic marine organisms.

In this paper we adopt 1300 ^{14}C yr as the best estimate for a circum-Antarctic correction for all ^{14}C dated marine organisms, for the sake of comparing glacial histories of the different areas. All ^{14}C ages given in the text have been corrected by that amount, no matter which reservoir correction or calibration was originally made by the authors cited as source of the data. All ages are in uncalibrated ^{14}C yr BP. At the same time we stress that there are still large uncertainties in the Antarctic marine reservoir effect.

Deglaciation and Holocene glacial history in Antarctica

The overall extent of ice cover in Antarctica during the LGM is not well known and some existing reconstructions are controversial. One maximum reconstruction suggests that the peripheral domes of the Antarctic Ice Sheet were 500–1000 m thicker than at present and that ice extended out to the continental shelf break around most of Antarctica (Denton 1979, Hughes *et al.* 1981, Clark & Lingle 1979, Denton *et al.* 1991, Zhang 1992). Other reconstructions indicate a smaller ice extent. Mayewski (1975) maintained that the West Antarctic Ice Sheet was only slightly, if at all, larger than it is today. Data from East Antarctica have been interpreted as indicating that ice either did not extend to the shelf edge (Colhoun & Adamson 1992a, Goodwin 1993) or that ice extended insignificantly farther out than today in some areas (Hayashi & Yoshida 1994). The timing of the LGM around Antarctica is likewise poorly known, but most authors assume it to have coincided with the timing of lowest global sea levels around 20 000–18 000 yr BP. The LGM in the western Ross Sea area has been dated to 20 000–17 000 yr BP (Stuiver *et al.* 1981, Anderson *et al.* 1992, Kellogg *et al.* 1996, Licht *et al.* 1996). Hall (1997) has recently dated the LGM in Taylor Valley, southern Victoria Land, to 14 600–12 700 yr BP. In East Antarctica, on the coast of Mac. Robertson Land and at Prydz Bay (Fig. 1) the LGM preceded 17 000 yr BP and 10 700 yr BP, respectively (Harris *et al.* in press, Domack *et al.* 1991a).

The Antarctic Peninsula region

The Antarctic Peninsula ice sheet (Fig. 2) is a part of the marine-based West Antarctic Ice Sheet, where sea level is a major control on ice volume. Sugden & Clapperton (1977) suggested, on the basis of bathymetric data showing glacial troughs incised into the submarine continental shelf, that during the LGM a number of ice caps formed on the South Shetland Islands, separated from the Antarctic Peninsula ice sheet by the deep Bransfield Strait. According to their reconstruction the main control for ice extension was sea level, and the presently 200 m deep submarine platforms around the islands and along the Antarctic Peninsula roughly coincide with the ice extension. There is no evidence that the South Shetland Islands were overridden by ice from the Antarctic continent during the LGM (John 1972, Sugden & Clapperton 1977), as suggested by Hughes (1975) and Hughes *et al.* (1981).

Evidence of more extensive ice cover than today exists all along the Antarctic Peninsula in the form of ice-abraded ridge crests at high altitudes, striated bedrock on presently ice-free islands, erratics and thin till deposits, as well as raised beach and marine deposits (John & Sugden 1971, Sugden & John 1973, Curl 1980, Clapperton & Sugden 1982, 1988, Ingólfsson *et al.* 1992). The timing of the onset of deglaciation in the Antarctic Peninsula area is not known. Sugden & John (1973) and Sugden & Clapperton (1977) suggested that it was triggered by rising sea levels some time after 14 000 yr BP. Banfield & Anderson (1995) found that the Bransfield basin was free of shelf ice as early as 14 000 yr BP, but recent investigations on the Antarctic Peninsula shelf (Pope & Anderson 1992, Pudsey *et al.* 1994) indicate initial ice retreat from outer and middle shelf areas shortly before 11 000 yr BP.

Deglaciation of the inner shelf and the presently ice-free land areas in the Antarctic Peninsula region was considerably delayed after 11 000 yr BP. Some inner shelf areas were deglaciated as late as 8000 to 6000 yr BP (Clapperton & Sugden 1982, Herron & Anderson 1990, Pudsey *et al.* 1994, Shevenell *et al.* 1996). The oldest ^{14}C dates, on fossil molluscs from raised marine deposits give minimum ages for the deglaciation of northern Antarctic Peninsula coastal areas (Fig. 2) as 8400 yr BP on the South Shetland Islands (Sugden & John 1973) and 7300 yr BP on James Ross Island (Hjort *et al.* (1997). At Alexander Island (Fig. 2) deglaciation occurred some time before 6000 yr BP (Clapperton & Sugden 1982). A number of studies on land suggest that once above the coastline, between 8400–6000 yr BP, glaciers retreated and disintegrated slowly towards positions at or inside their present margins (Barsch & Mäusbacher 1986, Mäusbacher *et al.* 1989, Mäusbacher 1991, Ingólfsson *et al.* 1992, Björck *et al.* 1993, 1996a, López-Martínez *et al.* 1996, Hjort *et al.* 1997). Data from Livingston Island (Fig. 2), in the South Shetlands, date the deglaciation of Byers Peninsula to 5000–3000 yr BP (Björck *et al.* 1996b). The deglaciation on King George Island was completed by *c.* 6000 yr BP (Martinez-Macchiavello *et al.* 1996).

There are indications of glacial readvance on King George Island (Sugden & John 1973, Mäusbacher 1991) and expansion of the ice shelf in George VI Sound, east of Alexander Island (Sugden & Clapperton 1981, Clapperton & Sugden 1982), some time after 6000 yr BP. A glacial readvance also occurred on James Ross Island, culminating between 5000 and 4500 yr BP (Rabassa 1983, Hjort *et al.* 1997), and indications of a post-5300 yr BP glacial expansion have been described from Brabant Island (Hansom & Flint 1989). Clapperton (1990) interpreted mid-Holocene glacier expansion in the Antarctic Peninsula region to reflect neoglacial cooling,

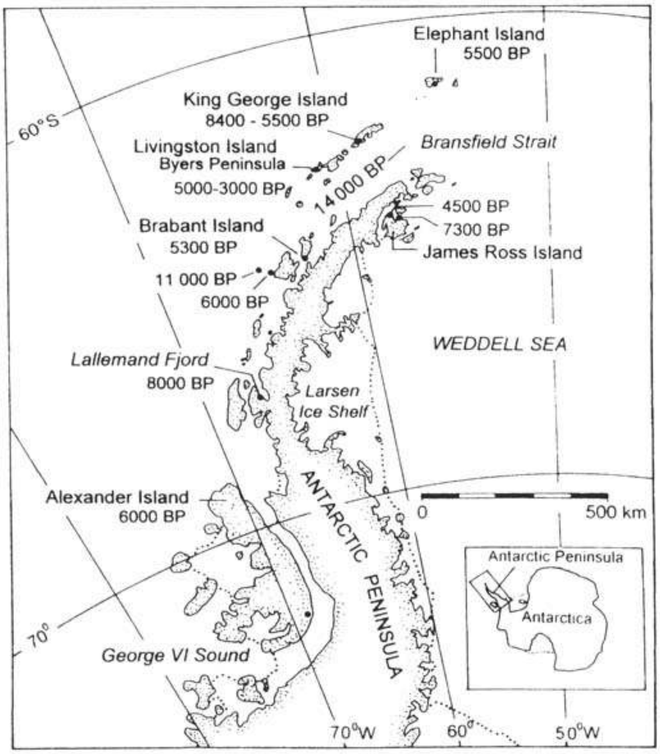

Fig. 2. The Antarctic Peninsula regions, with sites discussed in the text. Dots give approximate locations for ^{14}C dated samples giving minimum ages for the deglaciation.

whereas Ingólfsson *et* al. (1992) and Hjort *et al.* (1997) suggested that increases in precipitation might periodically have overcome increases in temperature in the generally cold climate, causing a cessation of mid-Holocene deglaciation or even glacial advance. However, recent nearshore marine data (Shevenell *et al.* 1996) may indicate a cooling background for these glacial readvances (Hjort *et al.* in press).

The present interglacial environment in the Antarctic Peninsula region dates back to 6000–5000 yr BP, when lake sediments started to accumulate in ice-free basins, and moss banks began to grow (see Fenton 1982, Barsch & Mäusbacher 1986, Mäusbacher *et al.* 1989, Mäusbacher 1991, Björck *et al.* 1991b, 1991d, 1993, 1996a, Yang & Harwood 1997). Palaeoclimatic studies based on numerous stratigraphical parameters in lake sediments and moss-bank peats in the northern part of the region (Björck *et al.* 1991b, 1993, 1996a) indicate that the climate fluctuated from relatively mild and humid conditions at 5000 yr BP to more cold and arid conditions. Around 4000 yr BP a gradual warming occurred, coupled with increasing humidity. These mild and humid conditions reached an optimum around 3000 yr BP, whereafter a distinct climatic deterioration occurred, again with colder and drier conditions. Cold and arid climate then persisted until 1500–1400 yr BP, and thereafter the climate has been somewhat warmer and more humid, but still cold compared to the climatic optimum. The climatic pattern for the last 5000 years is quite similar in the South Shetland Islands, in the maritime climate west of the peninsula and James Ross Island, and in the colder and drier polar climate in the western Weddell Sea. Björck *et al.* (1996a) suggested this might indicate that the primary factor controlling the climatic variations is the strength of the high-pressure atmosphere cell over the Antarctic ice sheet.

A number of investigations indicate that glaciers have oscillated and expanded somewhat in the Antarctic Peninsula region during the past 2500 years (John & Sugden 1971, Sugden & John 1973, Zale & Karlén 1989, Clapperton 1990, López-Martínez *et al.* 1996). In the South Shetland Islands, Curl (1980), Birkenmajer (1981), Clapperton & Sugden (1988) and Björck *et al.* (1996b) found evidence for glacial expansions in the form of readvance moraines on raised beaches, which they suggested coincided with the Little Ice Age glacial expansion in the Northern Hemisphere. Radiocarbon dating on whalebones found on some raised beaches give recent ages when a sea correction of 1300 yr is applied, but lichenometric dating, using *Rhizocarpon geographicum* thalli, dates the advances to 1240 AD (Birkenmajer 1981), 1720 AD (Curl 1980) and 1780–1822 AD (Birkenmajer 1981). Limiting dates of 1837 AD and 1880 AD were derived from lichenometry for two moraines in the South Orkney Islands (Lindsay 1973). A whalebone found on top of a moraine-ridge on Livingston Island dates a glacial advance there to after 1690 AD (Björck *et al.* 1996b).

A marine record from near-shore glaciomarine sediments in Lallemand Fjord (Fig. 2) indicates a pattern broadly similar to the record on land from the Antarctic Peninsula region (Shevenell *et al.* 1996). There, deglaciation of the inner shelf occurred somewhat before 8000 yr BP, and was followed by a period of open marine conditions with variable extent of sea ice between 8000 and 2700 yr BP. However, the data indicate a cooling between 6000 yr BP and 5000 yr BP, which might coincide with the mid-Holocene glacial readvances mentioned above. A climatic optimum, reflected by high productivity in the fjord, was recognized between 4200 and 2700 yr BP. After 2700 yr BP, a decrease in productivity and diatom abundance reflects more extensive and seasonally persistent sea ice. After 400 yr BP, ice shelf advance into the fjord was documented, correlating with the Little Ice Age (Shevenell *et al.* 1996, Domack *et al.* 1995).

A broad synthesis of the glacial history and related Holocene environmental development in the Antarctic Peninsula region is shown in Fig. 3.

East Antarctica

The extent of ice during the LGM in East Antarctica is not well known and partly controversial. A maximum

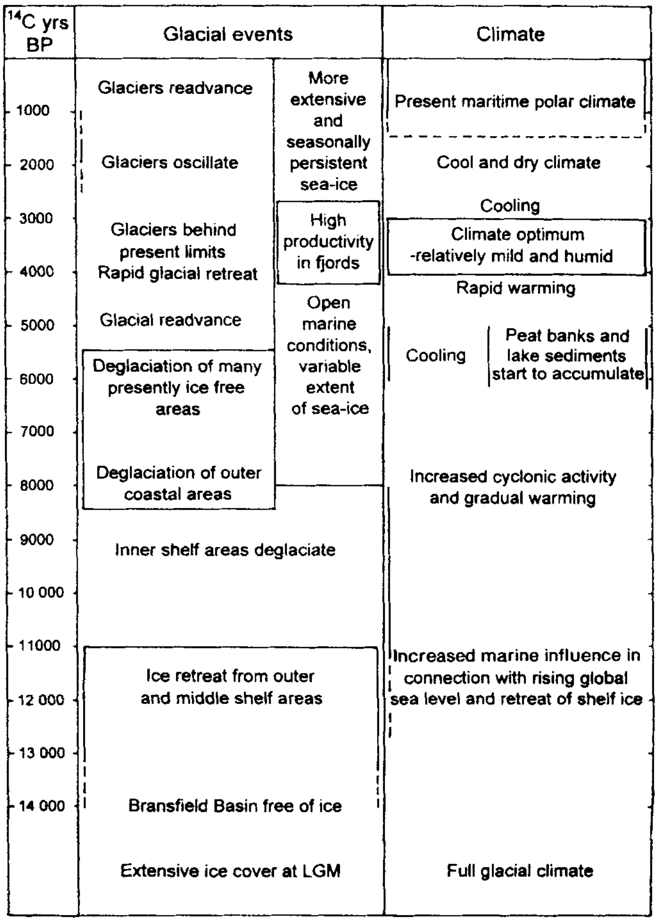

Fig. 3. Broad synthesis of the Antarctic Peninsula glacial development and associated environmental changes since the LGM.

reconstruction extends ice to the shelf break (Denton 1979, Hughes *et al.* 1981) while a minimalist view suggests only modest glacial oscillations and that some coastal oases remained ice free (Omoto 1977, Hayashi & Yoshida 1994, Adamson *et al.* 1997). Other data from East Antarctica suggest that ice extended off the present coast, but not all the way to the shelf edge (Colhoun & Adamson 1992a, Fitzsimons & Domack 1993, Goodwin 1993). The ice-free areas in East Antarctica, discussed below, are the Vestfjella and Heimefrontfjella nunatak ranges in western Dronning Maud Land, the Schirmacher and Untersee oases, Lützow-Holm Bay, Larsemann Hills and Vestfold Hills in Prydz Bay, Bunger Hills and Windmill Islands (Fig. 1).

Sedimentological data and ^{14}C dates from the Weddell Sea suggest that a grounded ice sheet extended to the shelf break off *western Dronning Maud Land* at 20 000 yr BP (Elverhøi 1981). Jonsson (1988) used observations of glacial striae to reconstruct ice thickness at the LGM in northern Vestfjella, and concluded that the ice there had been 360 m thicker than today, but that the highest nunataks had been ice-free. Lintinen & Nenonen (1997) concluded that during the LGM the ice sheet was at least 700 m thicker in northern Vestfjella than today, while in Heimefrontfjella it was less than 200 m thicker. ^{14}C dates of stomach oil deposits from nesting sites of snow petrels (*Pagodroma nivea*) give 7400 yr BP as the minimum age for deglaciation of the southern Vestfjella nunataks and the lower altitudes of Heimefrontfjella and record continuous avian occupation since then.

Schirmacher Oasis (Fig. 1) is an ice-free area located *c.* 100 km south of the Lazarev Sea. It covers an area of 34 km^2. Another 100 km to the south, *Untersee Oasis* (Fig. 1) forms the eastern rim of the unglaciated Wohlthat Massif, surrounding Lake Untersee.

At least four generations of moraines have been differentiated in the Untersee Oasis (Stackebrand 1995). They were interpreted as glaciation and deglaciation stages, representing a succession from a total submergence by the inland ice towards the present ice-free setting. The last total ice coverage probably predates the LGM (Hiller *et al.* 1988). This is indicated by Late Wisconsinan ^{14}C ages on snow petrel stomach oil deposits from high-altitude locations in the Untersee Oasis. Lake Untersee (169 m deep and 11.4 km^2 in area) perhaps existed during the Late Pleistocene, but at least since earliest Holocene time (M. Schwab, personal communication 1998).

The last glacial submergence of Schirmacher Oasis is reflected in a sparse coverage of morainic material, fractured rocks, striated surfaces and roches moutonnées at many places in the oasis (Richter & Bormann 1995). ^{14}C dating of lake sediments suggests that deglaciation of Schirmacher Oases started before 3500 yr BP (M. Schwab, personal communication 1998). Hence, deglaciation here may have occurred significantly later than in Untersee Oasis to the south (see above), and also later that on the Lazarev Sea shelf to the north, where it was under way by 9500 yr BP (Gingele *et al.* 1997).

No indications are known from Untersee or Schirmacher oases for significant Holocene readvances of ice margins. However, grain-size distribution and radiocarbon ages of sediments on the Lazarev Sea shelf could indicate that ice tongues to the east may have readvanced some time between 8000 and 2000 yr BP (Gingele *et al.* 1997).

The only available information on the Holocene climate development of the region comes from Lake Untersee. Today it has a perennial ice cover, up to 5 m thick, but its sediment composition indicates that during the early Holocene the ice cover was only semi-permanent (M. Schwab, personal communication 1998). That might have been due to a warmer climate.

Ice-free areas fringing *Lützow-Holm Bay* (Fig. 1) occur on a number of islands and headlands, the largest one being 61 km^2. Signs of glaciation occur as discontinuous glacial drift and erratics, striated bedrock surfaces and streamlined glacial bedforms, and extensive raised beaches (Yoshikawa & Toya 1957, Hirakawa *et al.* 1984, Yoshida 1983, Igarahi *et al.* 1995, Maemoku *et al.* 1997, Sawagaki & Hirakawa 1997). Omoto (1977) and Hayashi & Yoshida (1994) concluded that ice retreat from the Lützow-Holm Bay area occurred before 30 000 yr BP, and that it had not been ice-covered during the LGM. Igarashi *et al.* (1995) concluded that the last major deglaciation in the area dated back to the last interglacial, on the basis of the occurrence of old (42 000–33 000 yr BP) fossil shells in raised beach deposits. Maemoku *et al.* (1997) also described old (46 000–32 000 yr BP) fossil shells from Lützow-Holm Bay in undisturbed raised beach deposits, and interpreted these data to indicate the area had not been overridden by glaciers during the LGM. The location of the ice-margin during the LGM is unknown (Igarashi *et al.* 1995, Maemoku *et al.* 1997). Raised beaches of Holocene age, 7000 to 2000 yr BP (Igarashi *et al.* 1995) do, however, occur. The Holocene marine limit is at 25 m a.s.l., indicating considerable regional isostatic response to decreased ice volume. Maemoku *et al.* (1997) suggested the East Antarctic ice sheet might have covered the southern part of Lützow-Holm Bay during the LGM. Nothing is known about the Holocene climate development in this area.

The Amery Ice Shelf (Fig. 1), the largest in East Antarctica, has been present throughout the Holocene since retreat from the LGM terminal moraines at mid-shelf in Prydz Bay (Harris & O'Brien, personal communication 1997). It has been suggested that the small (50 km^2) oasis of *Larsemann Hills* in Prydz Bay (Fig. 1) was either wholly or partly ice free during the LGM (Burgess *et al.* 1994, 1997) or completely covered by 200–500 m thick continental ice (Gillieson 1991). Erratic boulders are scattered throughout the Hills, but till deposits, moraine ridges and glacial striae occur only sparsely (Burgess *et al.* 1994). A number of sediment cores has been retrieved from lakes in the Larsemann Hills. There are serious problems with radiocarbon dating of the sediments, expressed as reversed stratigraphical age successions (Burgess *et al.* 1994). While most dated samples give mid-late Holocene ages, a single date gave 9400 yr BP and another 25 000 yr BP.

There is no evidence of postglacial raised beaches in the Larsemann Hills, in contrast to extensively developed beaches in the Vestfold and Bunger Hills (Fig. 1). Their absence is puzzling, both if the oasis was ice free during the LGM or if it was gradually deglaciated in Holocene times.

During the LGM the whole 400 km² of the *Vestfold Hills* oasis in Prydz Bay (Fig. 4) was covered by ice (Adamson & Pickard 1983, 1986). Evidence of glacial overriding include glacial striae, erratics, till deposits and moraine ridges, as well as raised beaches at altitudes below 10 m a.s.l. (Pickard 1985, Zhang 1992). The orientation of glacial striae is uniform, showing ice movement towards WNW across the oasis at a time of complete ice coverage (Adamson & Pickard 1983, 1986). According to Domack *et al.* (1991a, 1991b), open marine conditions existed on the shelf, some 30 km off Vestfold Hills, at 10 700 yr BP. The last deglaciation of Vestfold Hills has been determined by ^{14}C dates on molluscs from raised marine deposits and moraines, on marine algal sediments from numerous lake basins, and on fossil mosses (Adamson & Pickard 1983,1986, Pickard 1985, Pickard & Seppelt 1984, Pickard *et al.* 1986, Bronge 1992). A prominent feature of the Vestfold Hills is its numerous lakes. There are about 300 lakes, from freshwater to saline and hypersaline. Many were formerly marine inlets and became isolated by isostatic uplift following the glacial retreat. The oldest ^{14}C dates, giving minimum ages for the initial deglaciation and the incursion of marine water onto coastal areas, as well as for the initiation of aquatic moss growth, are between 8600–8400 yr BP (Pickard & Seppelt 1984, Fitzsimons & Domack 1993, Roberts & McMinn in press). According to Adamson & Pickard (1986) and Pickard *et al.* (1986), ice retreat thereafter was slow or stepwise, averaging 1–2 m yr^{-1}, with 20% of the land area exposed by 8000 yr BP, 50% by 5000 yr BP, and the ice margin reaching its present position in the last 1000 yrs. Fitzsimons & Domack (1993) and Fitzsimons & Colhoun (1995) maintain that the margin of the Sørsdal Glacier (Fig. 4), in the southern part of the Vestfold Hills was at or south of its present position by 8600 yr BP and has been relatively stable since then. The data of Fitzsimons & Domack (1993) show that deglaciation of at least part of the Vestfold Hills occurred earlier than previously thought.

Domack *et al.* (1991b) found evidence on the shelf for a middle Holocene readvance of floating ice tongues some time within the interval 7300–3800 yr BP. There is no dated evidence for this from the Vestfold Hills, but Adamson & Pickard (1986) suggested that moraine ridges on Broad Peninsula (Fig. 4) may have formed during a middle Holocene glacial advance. A late Holocene ice advance, called the Chelnok glaciation, is poorly dated but probably occurred some time between 2000 and 1000 yr BP (Pickard *et al.* 1984, Adamson & Pickard 1986, Zhang 1992). Fitzsimons & Colhoun (1995) found evidence for minor (≤ 500 m) late Holocene (post 700 yr BP) ice marginal fluctuations in the form of a discontinuous series of ice-cored moraine ridges. These may correlate with the Little Ice Age.

There is no detailed record for the Holocene climatic development in Vestfold Hills. Pickard *et al.* (1984, 1986) concluded that the "post-glacial" climate of Vestfold Hills had been very stable and similar to today's. Zhang (1992), however, on the basis of marine fossil assemblages and geomorphological criteria, proposed that a climatic optimum had occurred there sometime between 6200 and 3700 yr BP. In the geomorphological record, Pickard (1982) found evidence that the prevailing wind direction had been stable for the past 4000 years. The climatic implications of the Chelnok advance are unclear (Pickard *et al.* 1984, Adamson & Pickard 1986).

Björck *et al.* (1996a) re-interpreted the lake sediment data of Pickard *et al.* (1986) in terms of palaeoclimate development of the Vestfold Hills. They suggested that the deglaciation of the Watts Lake area (Fig. 4) prior to 4700 yr BP occurred in an arid and cold environment (low lake levels, high salinity), followed by a relatively warm and humid climate between 4700 and 3000 yr BP which caused intense melting of stagnant ice in the vicinity of the lakes. Massive input of fresh water into the basins caused salinity to fall and lake levels to rise.

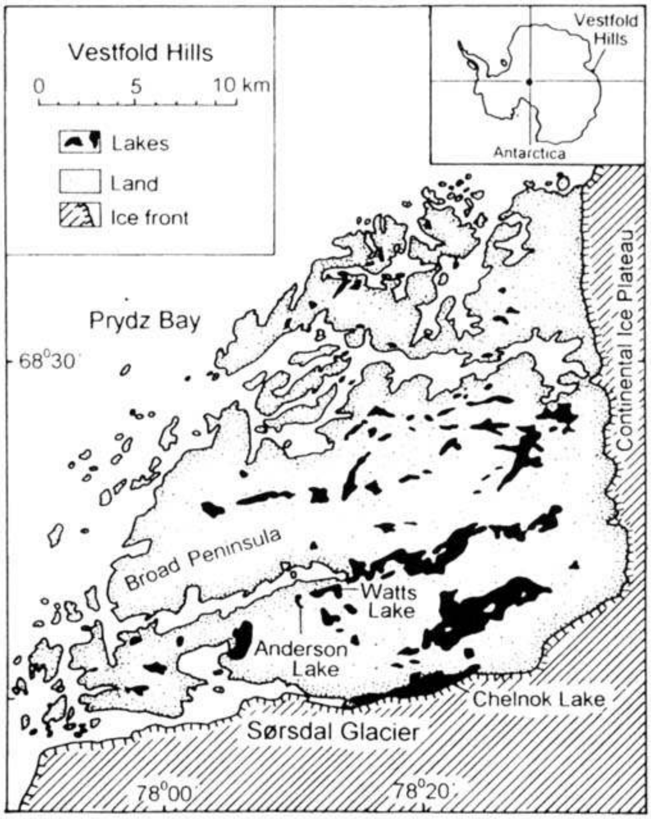

Fig. 4. The Vestfold Hills oasis in Prydz Bay, East Antarctica.

After 3000 yr BP the climate again turned arid and cold, with decreased fresh water input and lowered lake levels. Björck *et al.* (1996a) thus inferred a late Holocene (4700–3000 yr BP) climate optimum in the Vestfold Hills data.

Roberts & McMinn (1996, in press) used transfer functions for the reconstruction of past lakewater salinity from fossil diatom assemblages. They found that since 5200 yr BP, Anderson Lake (Fig. 4) in the Vestfold Hills had undergone cycles of varying salinity. The data show a long period with relatively low salinity, possibly indicating warmer and wetter conditions in the time interval *c.* 4200–2200 yr BP, if a constant sedimentation rate of 0.007 cm y^{-1} since 5200 yr BP is assumed. The glacial history of Vestfold Hills since the LGM is summarized in Fig. 5.

Bunger Hills (Fig. 6) form the most extensive oasis of deglaciated hills and marine inlets in East Antarctica, with a total size of 952 km^2, of which 482 km^2 are land (Wisniewski 1983). There are signs of extensive glaciation in the form of discontinuous but locally thick glacial drift deposits, striated bedrock surfaces, roches moutonnées and erratics, as well as extensive raised beaches. Colhoun & Adamson (1992a) and Augustinus *et al.* (1997) suggested that there is evidence that at some stage in the past the Antarctic ice sheet completely submerged most of the oasis, and that ice flow from south-east to north-west was independent of the local topography. Augustinus *et al.* (1997) proposed that this extensive glaciation may have predated the LGM. Limited glacial erosion, thick glacial deposits in the north-western part of the hills and the relatively low altitude of the postglacial marine limit (9–7 m a.s.l.) suggested to Colhoun & Adamson (1992a) and Colhoun (1997) that the LGM ice sheet was not very thick and consequently did not extend far onto the continental shelf. Melles *et al.* (1997) found till in 18 sediment cores retrieved from different basins within the oasis. They concluded that probably the whole oasis was buried by glaciers during the LGM.

^{14}C dates on total organic carbon from lacustrine and marine sediments indicate that the initial deglaciation of the southern part of the Bunger Hills oasis dates back to between 10 000–8000 yr BP (Bolshiyanov *et al.* 1990, 1991, Melles *et al.* 1994, Melles *et al.* 1997), and radiocarbon dates of stomach oil deposits at nest sites of snow petrels show occupation as early as 9500 yr BP (Bolshiyanov *et al.* 1991, Verkulich & Hiller 1994). The glacial retreat was partly controlled by the rise of sea level, which caused a relatively rapid collapse of the ice sheet margin (Colhoun & Adamson 1992a, Verkulich & Melles 1992). Melles *et al.* (1997) conclude that the first phase of deglaciation was also associated with climatic warming, indicated by high diatom concentrations in the sediments and a large meltwater input to the basins. By 7700 yr BP the sea had flooded all major inlets in Bunger Hills (Colhoun & Adamson 1991, Melles *et al.* 1997). The breeding colonies of snow petrels expanded continually, following ice retreat and the down-wasting of dead ice (Verkulich & Hiller 1994), with the most intense

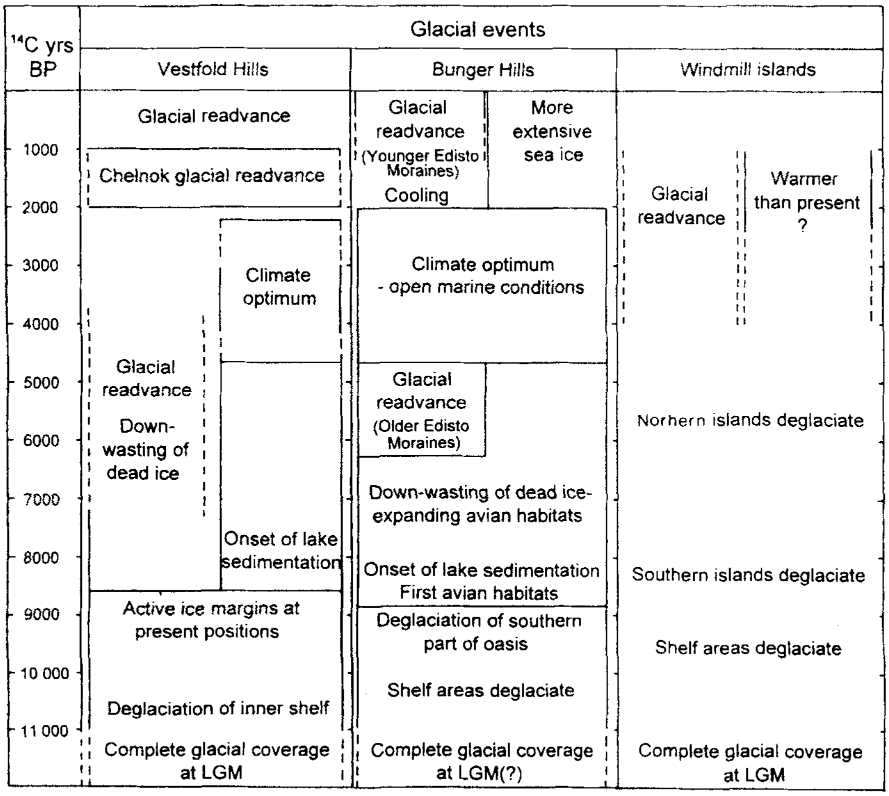

Fig. 5. Summary of the most complete East Antarctic records on glacial and climatic development since the LGM.

phases of colonization between 6700 yr BP and 4700 yr BP. Before 5600 yr BP, when the sea stood at the marine limit at 9–7 m, glaciers were at or behind their present margins (Colhoun & Adamson 1992a, Colhoun & Adamson 1992b). Melles *et al.* (1997) dated initiation of lacustrine sedimentation in the northern part of Bunger Hills to 6000 yr BP.

Evidence for the mid-late Holocene glacial and climatic evolution of Bunger Hills is, however, somewhat controversial. Bolshiyanov *et al.* (1991) suggested that the area was re-glaciated several times during the Holocene, causing damming of tributary valleys with periodic lake sediment deposition. They based their conclusions on lake sediment thickness, as well as on fluctuations in the growth rate of aquatic mosses and algae, and on changing salinity conditions as reflected by diatom assemblages and the geochemistry of lake sediments. Verkulich & Hiller (1994) found no evidence of any major Holocene glacial advance in the Bunger Hills and Colhoun & Adamson (1992a), Fitzsimons & Colhoun (1995) and Fitzsimons (1997) concluded that ice margins at the southern boundary of Bunger Hills had been fairly stable since the last deglaciation. At the western margin, however, glacier expansions of a few hundred metres resulted in the formation of the older Edisto moraines, post-dating 6200 yr BP (Colhoun & Adamson 1992a).

Verkulich & Melles (1992) and Melles *et al.* (1997) studied sediment cores from freshwater lakes and from marine basins in the Bunger Hills. Melles *et al.* (1997) reported high and stepwise increasing biogenic production between 4700 and 2000 yr BP, taken to indicate increased temperatures and correlated with the Antarctic Peninsula climate optimum at about the same time. Melles *et al.* (1997) also concluded that the glacial advance forming the older Edisto moraines predated the 4700 yr BP warming. This constrains the Edisto advance to some time between 6200–4700 yr BP, which is coincident with the cooling and glacial readvance in the Antarctic Peninsula area (see above). Colhoun & Adamson (1992a, 1992b) agreed with Rozycki's (1961) interpretation of beach morphology, suggesting that wave action may have been more important in the mid-Holocene, and that the extent and duration of sea ice has increased in the late Holocene.

Melles *et al.* (1997) found indications of climatic deterioration after 2000 yr BP and of subsequent warming to an intermediate level until the present. Colhoun & Adamson (1992a) described a glacial advance during the last few centuries, leading to the formation of the younger Edisto moraines. Melles *et al.* (1997) concluded that the younger Edisto moraines postdate 1100 yr BP. Marine shell fragments collected from the moraines were dated to 200 yr BP (Colhoun & Adamson 1992a), suggesting a Little Ice Age glacial event. A summary of the Bunger Hills data on deglaciation and associated environmental changes is given in Fig. 5.

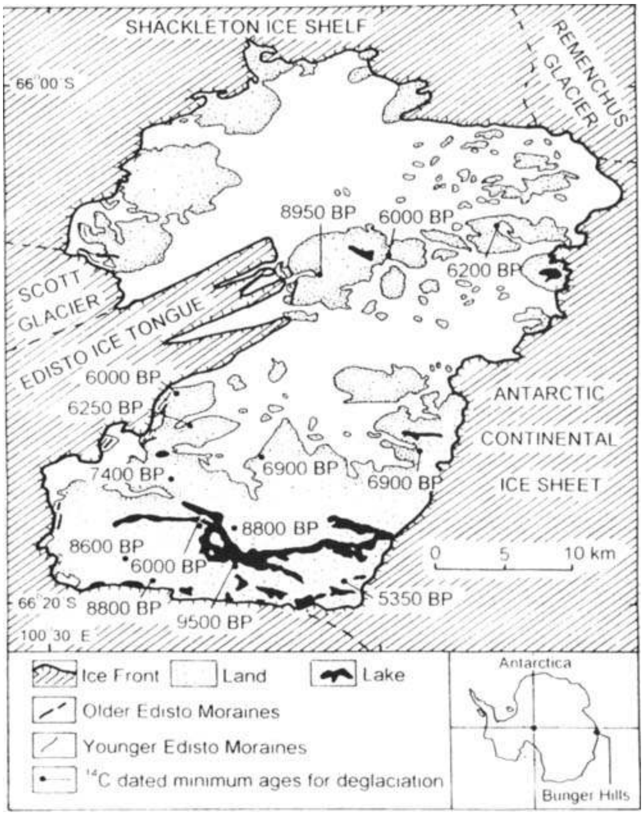

Fig. 6. The icebound Bunger Hills oasis, East Antarctica. Dots give locations of some selected sites where ^{14}C dated samples give minimum ages for deglaciation (modified from Colhoun & Adamson 1992a and Verkulich & Hiller 1994).

Windmill Islands (Fig. 1) are a group of low islands and peninsulas, forming a major oasis on Budd Coast in East Antarctica. Evidence of an extensive late Pleistocene ice cover are glacial polishing and striae on the gneiss bedrock, roches moutonnées and erratics, as well as raised beaches at altitudes below 32 m a.s.l. (Goodwin 1993). A shallow veneer of unconsolidated sediments occurs on the islands. Cameron *et al.* (1959) interpreted this as reworked till, but Goodwin (1993) found subglacially deposited fine sediments almost totally lacking. He argued that the best indicator for glacial overriding during the LGM were the raised beaches, bearing witness to isostatic rebound in connection with deglaciation. Goodwin (1993) calculated that the Late Pleistocene–early Holocene ice thickness over the Windmill Islands and the inner shelf had been <200 m and <400 m, respectively, and that ice had extended 8–15 km off the present coast.

Radiocarbon dating on bulk samples from basal lake sediments provides minimum estimates for the deglaciation of the Windmill Islands (Goodwin 1993). These indicate that the southern part of the islands were deglaciated before 8000 yr BP (Fig. 5), while the northern islands were only deglaciated some time before 5500 yr BP. Very little is known about the post-glacial Holocene climate development at Windmill Islands. Goodwin (1993) interpreted the onset of algal growth in the lakes to indicate warmer conditions than at present between 2000 and 1000 yr BP. He also found indications of higher lake levels during that period.

There is evidence for a readvance of the Law Dome ice sheet margin onto part of the Windmill Islands some time between 4000 and 1000 yr BP (Goodwin 1996). The overriding advance of the ice margin incorporated frozen coastal sediments from raised beach, lacustrine and proglacial environments together with slabs of marine ice from a palaeo-ice shelf, during the marginal transition from fringing ice shelf to grounded ice sheet. Goodwin (1996) attributed the readvance to a positive mass balance on the Law Dome caused by high precipitation rates during the Holocene.

Summary of the East Antarctic data

There is a broad pattern in the glacial histories from East Antarctica (Fig. 7). All studies infer relatively moderate ice thickness during the LGM, with ice extending off the present shore, but not very far onto the shelf. Some areas may have remained wholly or partly ice free throughout the LGM (Schirmacher and Untersee oases, Larsemann Hills, Lützow-Holm Bay). Ice retreat from the shelf areas was under way by 11 000–10 000 yr BP. The oldest deglaciation dates show marine and lacustrine environments, as well as avian habitats, developing in coastal areas between 10 000–8000 yr BP. The deglaciation history of Vestfold Hills is controversial. One reconstruction specifies it as successive and slow and mainly post-dating 8000 yr BP. In the other reconstruction, deglaciation was more or less complete by 8600 yr BP. There are indications from Vestfold Hills of a mid-Holocene glacial advance. The combined data from Bunger Hills show an early initial deglaciation phase, and retreat of the continental ice sheet margin to its present position by 10 000 yr BP. The retreat was coupled with collapse of ice over the marine inlets while the land was still depressed below present sea level around the Pleistocene–Holocene transition. Most of the downwasting of dead ice may have occurred successively after 8800 yr BP. There are indications of a mid-Holocene glacial readvance: expansion of the Edisto Glacier occurred after 6200 yr BP and before 4700 yr BP. The Windmill Islands were successively deglaciated between 8000 and 5500 yr BP. There are indications from both Vestfold Hills and Bunger Hills of a climate warmer and wetter than the present in the interval *c.* 4700–2000 yr BP, and the Windmill Islands record may tentatively be interpreted as indicating the same for the 4000–1000 yr BP interval. The East Antarctic data also show minor fluctuations of the ice margins over the past few hundred years (Goodwin in press), possibly correlative with the Little Ice Age advances in the Northern Hemisphere.

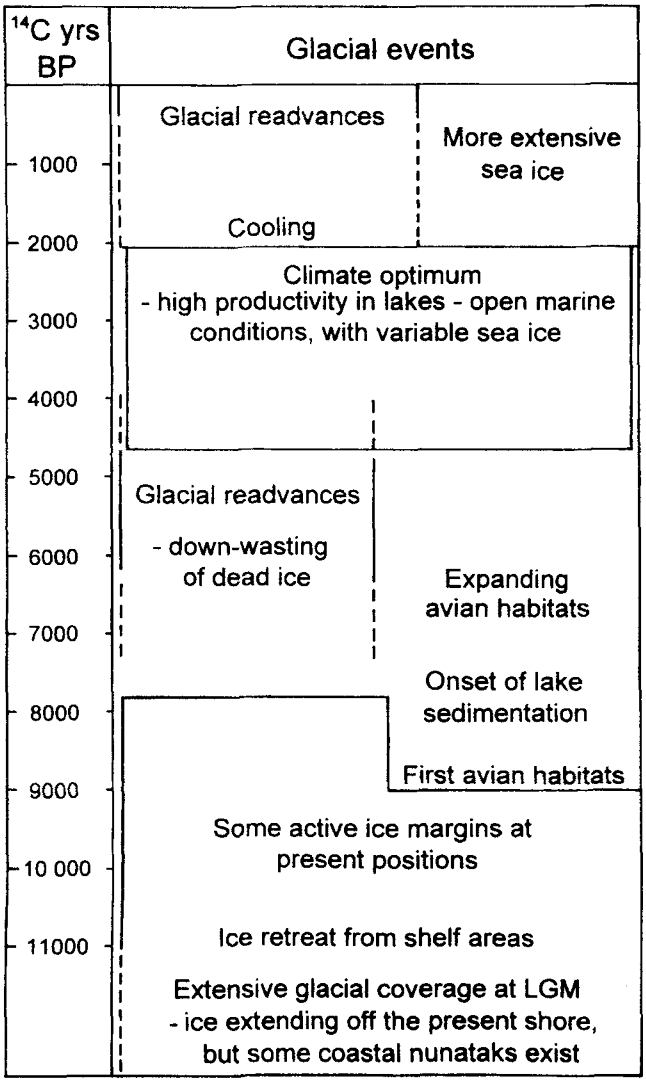

Fig. 7. Broad synthesis for the glacial and climatic development in East Antarctica since the LGM.

The Ross Sea area and coastal Victoria Land

The Ross ice drainage system comprises about 25% of the surface of the Antarctic Ice Sheet. Since the dawn of geological research in Antarctica there has been a discussion about the fluctuations of ice in the Ross Sea and Victoria Land (Fig. 8) in space and time (reviews in Stuiver *et al.* 1981 and Denton *et al.* 1989, 1991). Reconstructions of the LGM ice flowlines for the drainage of the East and West Antarctic ice sheets to the Ross Sea are conflicting (Drewry 1979, Denton *et al.* 1989, Clapperton & Sugden 1990, Kellogg *et al.* 1996), but most studies suggest that the Ross Sea embayment was largely filled by a low surface profile, marine based ice sheet. Glacial drift deposits (Ross Sea drift), containing kenyte erratics from Ross Island, show that during the LGM the West Antarctic ice sheet thickened and grounded in the Ross Sea and McMurdo Sound, pushing lobes of ice onto coastal southern Victoria Land and damming the Dry Valleys (Stuiver *et al.* 1981, Denton *et al.* 1989, 1991). Farther north along the Victoria Land coast, major outlet glaciers, in e.g., Terra Nova Bay, drained the East Antarctic ice sheet and coalesced with the marine based ice in the Ross Sea (Denton *et al.* 1989, Orombelli *et al.* 1991). Although reconstructions of ice

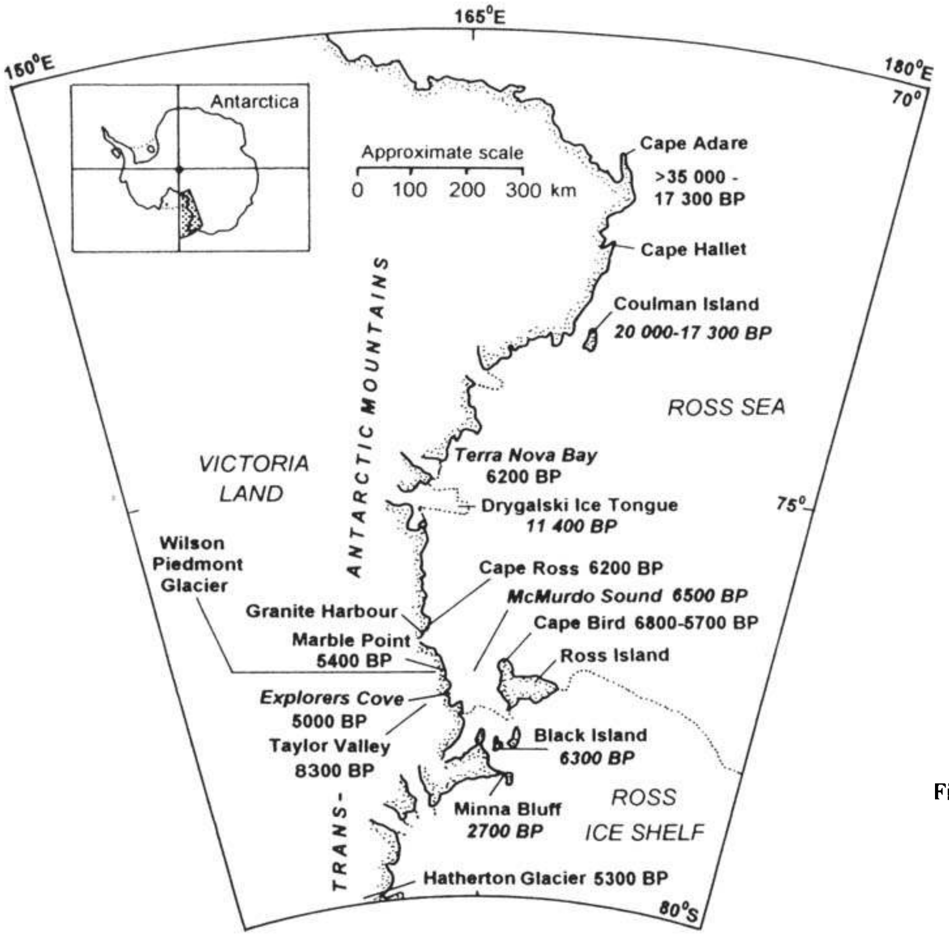

Fig. 8. The Ross Sea–Victoria Land region, with minimum ages for deglaciation. Radiocarbon ages in italic refer to grounding line recession.

extent in the Ross Sea during the LGM agree that ice was considerably expanded compared to the present (see Drewry 1979, Stuiver *et al.* 1981, Anderson *et al.* 1984, Denton *et al.* 1989, 1991, Clapperton & Sugden 1990, Orombelli *et al.* 1991, Kellogg *et al.* 1996, Licht *et al.* 1996), there remains uncertainty about the maximum position of the grounded ice. Licht *et al.* (1996) recognized tills in the western Ross Sea, but only in areas south of Coulman Island. Kellogg *et al.* (1996) suggested that grounded LGM ice extended to the shelf break off Cape Adare, while Denton *et al.* (1989), Anderson *et al.* (1992), Baroni & Orombelli (1994a), Shipp & Anderson (1994) and Licht *et al.* (1996) placed the LGM-grounding line along the northern Victoria Land coast in the vicinity of Coulman Island, between 74°S and 73°S. Anderson *et al.* (1992) dated the grounding line there to 17 300 yr BP whereas Licht *et al.* (1996) dated it to *c.* 20 000 yr BP. Penguin rookeries at Cape Adare and Cape Hallet were probably occupied between >35 000 and 17 300 yr BP (Baroni & Orombelli 1994a). Since these can exist only where there is access to open water in summer, they show that the coastal north-western Ross Sea was free of glacial ice during the LGM.

The terrestrial data on the timing and pattern of deglaciation derive from studies of lacustrine sediments in the Dry Valleys, of glacial landforms, drift deposits and raised beaches along Victoria Land, as well as from ornithogenic soils in both presently occupied and abandoned penguin rookeries (Stuiver *et al.* 1981, Denton *et al.* 1989, 1991, Clapperton & Sugden 1990, Baroni & Orombelli 1991, 1994a, Orombelli *et al.* 1991, Hall 1997). According to Stuiver *et al.* (1981) the LGM in the Dry Valleys culminated between 24 000 and 17 000 yr BP, and numerous ^{14}C dates of glacial lacustrine sediments in Taylor Valley date were taken to indicate deglaciation between *c.* 16 000 and 13 000 yr BP (Stuiver *et al.* 1981, Denton *et al.* 1989). Denton *et al.* (1991) and new data by Hall (1997) now suggest that this has to be seriously revised, and that the LGM in Taylor Valley dates to 14 600–12 700 yr BP, and ice was within a few hundred metres of its maximum position as late as 10 800 yr BP.

A number of ^{14}C dates constrain the marine-based ice sheet retreat in the Ross Sea from the LGM-grounding line after 17 000 yr BP. Minimum ages for the deglaciation of the area between 75°S and 76°S come from offshore sediments close to the present Drygalski ice tongue, and are *c.* 11 400 yr BP (Licht *et al.* 1996). At the LGM, the Terra Nova Bay region was occupied by coalescing outlet glaciers, draining the East Antarctic Ice Sheet and joining the marine based Ross Ice Sheet (Baroni & Orombelli 1991, Orombelli *et al.* 1991). Glacial drift deposits related to the LGM indicate that the ice surface in the bay was about 400 m above the present sea level. Even though the grounding line of the ice sheet had retreated to south of Terra Nova Bay by 11 400 yr BP, the break-up of the shelf ice was probably considerably delayed. The minimum age for the deglaciation of Terra Nova Bay is some time before 6200 yr BP, when the coasts there became free of ice and marine habitats could begin to develop (Baroni & Orombelli 1991, Orombelli *et al.* 1991). The oldest penguin rookeries in the Bay date back to 5800 yr BP.

Denton *et al.* 1991 and Hall (1997) date the presence of the Ross Sea Ice Sheet in Explorers Cove, southern Victoria Land, damming lakes along the valley threshold, at 8300 yr BP. Date of penguin rookeries on Ross Island and in the McMurdo Sound area give minimum ages for the deglaciation in the southern Ross Sea. Rookeries at Cape Bird were occupied by 6800–5700 yr BP (Speir & Cowling 1984, Heine & Speir 1989). Marine fossils collected from debris bands on the McMurdo Ice Shelf date the grounding line recession of the Ross Sea ice sheet to a position south of Black Island by 6300 yr BP (Kellogg *et al.* 1990), which is in line with the penguin data. ^{14}C dates on fossil marine molluscs from raised beaches along the southern Victoria Land coast, McMurdo Sound and the islands in the south-western Ross Sea, as well as additional dates on penguin occupation in McMurdo Sound, all give minimum ages for the deglaciation as *c.* 6300–6100 yr BP (Stuiver *et al.* 1981, Speir & Cowling 1984, Denton *et al.* 1989, Colhoun *et al.* 1992, Baroni & Orombelli 1994a, Kellogg *et al.* 1996). A relative sea-level curve for the coast north of Explorers Cove indicates unloading of grounded ice by 6300 yr BP and provides one minimum estimate for the deglaciation of coastal southern Victoria Land (Hall 1977). It coincides well with the oldest ^{14}C date on a bivalve from a core in McMurdo Sound, which gives 6500 yr BP as the minimum age of grounding line retreat from this area (Licht *et al.* 1996). The marine limit becomes gradually younger southwards, which might indicate lagging of shelf ice recession behind grounding line recession. The age of the marine limit is 6200 yr BP in Terra Nova Bay, 5400 yr BP at Marble Point/South Stream and 5000 yr BP in the Explorers Cove area (Stuiver *et al.* 1981, Denton *et al.* 1989, Orombelli *et al.* 1991, Berkman 1997, Hall 1997).

Bockheim *et al.* (1989) concluded that Holocene ice-surface lowering of the Hatherton Glacier in the Transantarctic Mountains, corresponding in time with grounding line recession in the south-western Ross embayment, occurred before *c.* 5300 yr BP. Numerous ^{14}C dates on marine macrofossils recovered from dirt bands in shelf ice in southern McMurdo Sound indicate that the grounding line had retreated to an unknown position south of Minna Bluff by 2700 yr BP (Kellogg *et al.* 1990). It is not known when the McMurdo Ice Shelf attained its present grounding line position, and also unknown is whether the grounding line is presently stable, advancing or retreating (Kellogg *et al.* 1996).

There is some information available on mid- to late-Holocene glacier variations in Victoria Land. After the deglaciation in Terra Nova Bay, the ice shelves entering the bay were less extensive than today (Orombelli *et al.* 1991, Baroni 1994, Baroni & Orombelli 1994b). The ice margins stood 2–5 km inside their present margins between 6200 and 5300 yr BP. A readvance across raised beaches took place some time after 5300 yr BP (Baroni & Orombelli 1994b). There was then a renewed withdrawal phase between *c.* 1000 and 500 yr BP, correlated by Baroni (1994) and Baroni & Orombelli (1994b) with the Northern Hemisphere Medieval Warm Period (*c.* 1000–1300 AD). Moraine ridges containing fossil marine

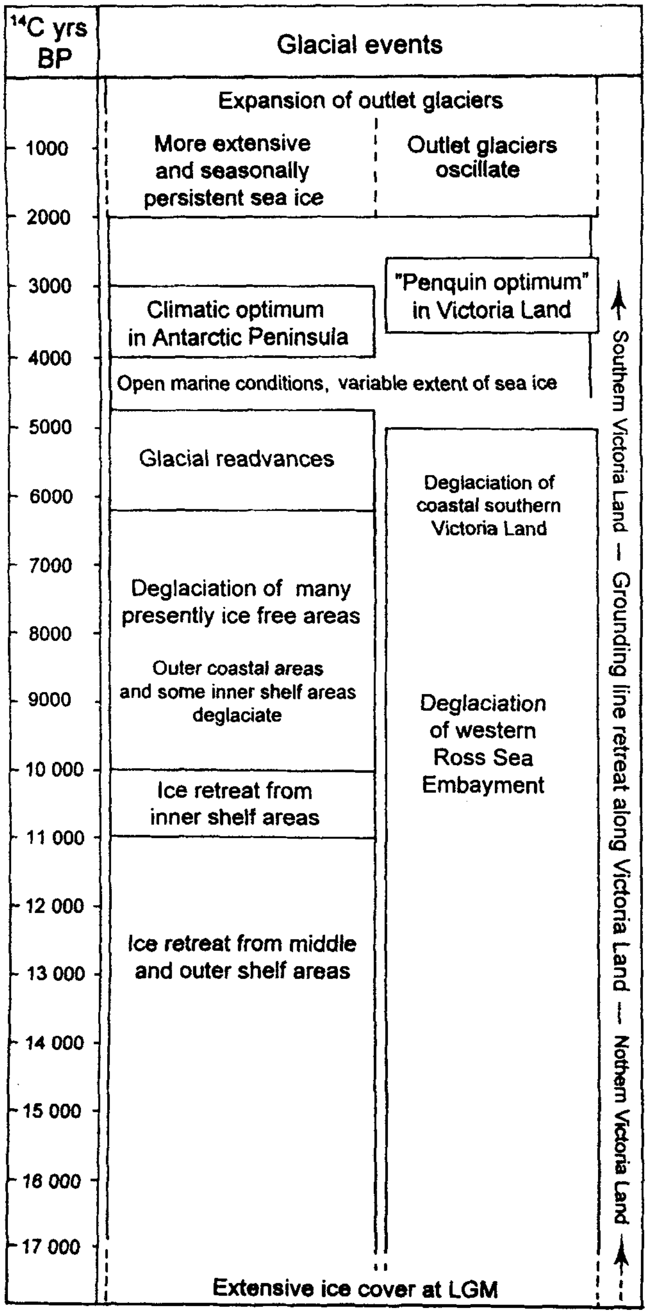

Fig. 9. Broad synthesis for Antarctic glacial and associated environmental development since the LGM.

shells then show a glacial readvance in Terra Nova Bay after 500 yr BP, which Baroni & Orombelli (1994b) suggested might correspond to the Little Ice Age glacial expansion in the Northern Hemisphere. Möller (1995) described a system of minor, fresh-looking thrust moraines in Granite Harbour, Victoria Land, which he suggested were formed by repeated oscillations during general retreat of the ice front. The youngest thrust moraine post-dates 1910 AD, when the British Terra Nova Expedition surveyed the area, and Möller (1995) concluded that the moraine ridge system was formed in connection with a Little Ice Age glacial expansion.

Alpine glaciers in the Dry Valleys probably did not contribute to the Ross Sea Ice Sheet, and some of these are presently at their maximum frontal positions since the LGM (Stuiver *et al.* 1981). The Wilson Piedmont Glacier in southern Victoria Land was contiguous with the Ross Ice Sheet at the LGM (Hall 1997) and merged with it north of Explorers Cove (Fig. 8). The Wilson Piedmont glacier retreated inside the coast between 5700 and 5000 yr BP (Hall 1997). It readvanced in late Holocene times, and at a number of locations it advanced across raised beaches which date back to 5400–5000 yr BP (Nichols 1968, Stuiver *et al.* 1981, Denton *et al.* 1989, Hall 1997).

The best information on palaeoclimatic development in Victoria Land during the latter part of the Holocene comes from the spatial and temporal distribution of penguin rookeries (Baroni & Orombelli 1994a). The location of Adélie penguin colonies 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 & Orombelli (1994a) documented the continuous presence of Adélie penguins after *c.* 7000 yr BP, but the greatest diffusion of rookeries occurred between 3600 and 2600 yr BP. They termed this interval the "penguin optimum", and concluded that it was a period of particularly favourable environmental conditions. It was followed by a sudden decrease in the number of penguin rookeries shortly after 2600 yr BP, particularly in southern Victoria Land, attributed to an increase in sea ice. A mid-late Holocene decrease in sea ice and longer seasons with open water in southern Victoria Land is supported by the observations of Nichols (1968). He described raised beaches, clearly of high-energy type, at many sites south of Granite Harbour. At many of these sites today the ice-foot rarely breaks up and recent beaches are of low-energy type. All beaches described by Nichols (1968) are younger than 5400 yr BP (Stuiver *et al.* 1981, Berkman 1997). Abandoned Adélie penguin rookeries, occupied during the penguin optimum, occur at Cape Ross and Marble Point (Baroni & Orombelli 1994a), where Nichols (1968) described high energy beach ridges. The raised beach deposits at Marble Point have been dated to 5400 yr BP (Stuiver *et al.* 1981, Berkman 1997).

Summary

Antarctic glaciers range in size from continental ice sheets to small outlet glaciers, with varied sensitivity and response time to climatic and other environmental changes. However, a broad pattern can be discerned in the glacial histories from the coastal sites around Antarctica discussed in this study (Fig. 9), from the LGM towards the present:

a) Ice extended offshore around most of Antarctica at the LGM, although some oases may have remained ice free.

b) Ice retreat from the LGM positions was under way by 17 000–14 000 yr BP, and by 11 000–10 000 yr BP initial deglaciation of some inner shelf areas as well as a few outer coastal land areas had occurred. The ice retreat from the LGM positions around Antarctica was probably eustatically controlled. Melting Northern Hemisphere glaciers caused global sea level to rise and Antarctic grounding lines and ice fronts gradually retreated from outer shelf areas towards the continent (Hollin 1962, Stuiver *et al.* 1981).

c) Deglaciation of some shallow inner shelf areas and of most presently ice-free land areas, with the exception of some East Antarctic oases, occurred between *c.* 10 000–5000 yr BP. After the receding glaciers had reached the edges of most of the presently ice free land areas or shallow inner shelf areas, by 10 000–8000 yr BP, the rate of deglaciation slowed considerably. It was not until 5000 yr BP that most Antarctic glaciers had retreated to, or behind, their present positions.

d) Mid-Holocene glacial readvances are described from several areas around Antarctica. These are mostly dated by maximum ages only, but on James Ross Island in the Weddell Sea the glacial readvance culminated between 5000 and 4500 yr BP and in the Bunger Hills in East Antarctica between 6200–4700 yr BP.

e) Although the initial phase of deglaciation in Antarctica, up to 10 000 yr BP was probably controlled by the global sea level rise, the slower phase of gradual deglaciation between 10 000 and 5000 yr BP was most likely controlled by gradual interglacial warming in Antarctica. This warming seems to have been temporarily halted by a cooling spell between 6500–4700 yr BP, reflected in the mid-Holocene glacial readvances and also documented in some lake- and marine records. The renewed warming thereafter peaked in a hypsithermal event, visible in the records between 4700 and 2000 yr BP. The best documented records put it between 4000 and 3000 yr BP, in the Antarctic Peninsula region, and between 3600 and 2600 yr BP in coastal Victoria Land. Many records indicate less sea ice in coastal areas and higher beach energies than at present in the mid-Holocene, which probably coincides with this relatively warm event. It also roughly correlates in time with peak summer insolation in the south (Budd & Smith 1987, Budd & Rayner 1990).

f) The hypsithermal event was followed by a general cooling. Outlet glaciers have expanded around Antarctica for the past *c.* 500 years. Goodwin (in press) suggested that late Holocene Antarctic ice-volume expansion might be equivalent to a *c.* 1 m global sea level lowering.

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