

Quaternary glacial and climate history of Antarctica

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1. Introduction

The Antarctic Ice Sheet, containing $25\text{--}30 \times 10^6 \text{ km}^3$ of glacial ice (Drewry *et al.*, 1982; Vaughan 2000), is the largest glacial system on Earth. The western part of the present Antarctic ice sheet (West Antarctic Ice Sheet, WAIS), west of the Transantarctic Mountains (Fig. 1), is largely marine-based, currently very dynamic and discharges huge volumes of ice to floating ice shelves through large ice-stream systems. To the east of the Transantarctic Mountains the current East Antarctic Ice Sheet (EAIS) is predominantly terrestrial, relatively stable and has limited discharge into surrounding ice shelves (Fig. 2). The WAIS and EAIS contain about $3.3 \times 10^6 \text{ km}^3$ and $26 \times 10^6 \text{ km}^3$ of ice, respectively (Drewry, 1983; Denton *et al.*, 1991). The Antarctic ice sheet, potentially containing 57–66 m of sea level equivalent (Denton *et al.*, 1991; Vaughan 2000), is a critical factor in regulating, modifying and forcing the global climate and oceanographic system. During most of the late Cenozoic the Antarctic ice sheet has driven global eustasy and deep-ocean circulation, and acted as a regulator of global climate (Anderson, 1999). However, the influences of Antarctic Ice Sheet fluctuations in the Quaternary history of global climate, after initiation of major glaciations in the Northern Hemisphere, 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 is what caused the glacial fluctuations observed in the records? Antarctic glaciers respond both to global sea level fluctuations, mainly controlled during the Quaternary by developments in the Northern Hemisphere, and to Southern Hemisphere climate changes. A good understanding of the Late Quaternary glacial and climate history of Antarctica will 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; Blunier *et al.*, 1998; Steig *et al.*, 1998).

Studies of Late Quaternary climate changes in Antarctica have been focused on ice-core (e.g. Jouzel *et al.*, 1987; Ciais *et al.*, 1994; Blunier *et al.*, 1998, Thompson *et al.*, 1998; Steig *et al.*, 1998, 2000; Petit *et al.*, 1999; Masson *et al.*, 2000) and marine records (e.g. Elverhøi,

1981; Anderson & Bartek, 1992; Shipp & Andersen, 1994; Licht *et al.*, 1996; Bentley & Anderson, 1998; Anderson, 1999; Domack *et al.*, 1991, 2001), as a consequence of the scarcity of chronologically well constrained geological data on land. However, the last two decades have seen increasingly more sophisticated evidence from ice-free areas in Antarctica, based on glacial stratigraphical and morphological investigations, studies of lake sediment and moss-bank cores, and of fossil penguin rookeries (reviews in Ingólfsson *et al.* (1998) and Bentley (1999)). Current knowledge primarily concerns the ‘postglacial’ developments, i.e. from 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 current knowledge of the glacial and climate history of Antarctica. Since large-scale glaciations there are not restricted to the Quaternary Period, the paper will first present a brief summary of the pre-Quaternary history, then proceed with a general overview of the Pleistocene record, and finish with more detailed reviews of the Late Quaternary (including the Holocene) glacial and climatic development, with focus on the Antarctic Peninsula and the Ross Sea/Victoria Land regions. The focus of the review is on the terrestrial record of the glacial and climate history of Antarctica, with reference to the marine record as given in excellent reviews by e.g. Anderson (1999) and Anderson *et al.* (2002).

2. Pre-Quaternary glacial history of Antarctica

The Antarctic ice sheet has existed for about 35–40 million years, since the mid-Tertiary, and it is widely believed that in East Antarctica it reached continental proportions by the latest Eocene–Early Oligocene (Hambrey *et al.*, 1989; Birkenmayer, 1987, 1991; Prentice & Mathews, 1988; Barrera & Huber, 1993; Barrett *et al.*, 1991; Denton *et al.*, 1991). The nature and timing of the initial glaciation of Antarctica is not well known, and the evolution and nature of the WAIS remains controversial (Wilson, 1995). Existing reconstructions of pre-Quaternary Antarctic ice volumes and ice-extent rely heavily on interpretations of Ocean Drilling Program (ODP) data from offshore cores, such as oxygen isotope ($\delta^{18}\text{O}$) concentrations, ice-rafted detritus (IRD) concentrations, sediment type and clay mineralogy, microfossil assemblages and occurrences of hiatuses (Fig. 3). The geological literature on Antarctic pre-Quaternary glacial history and palaeoceanography is vast,

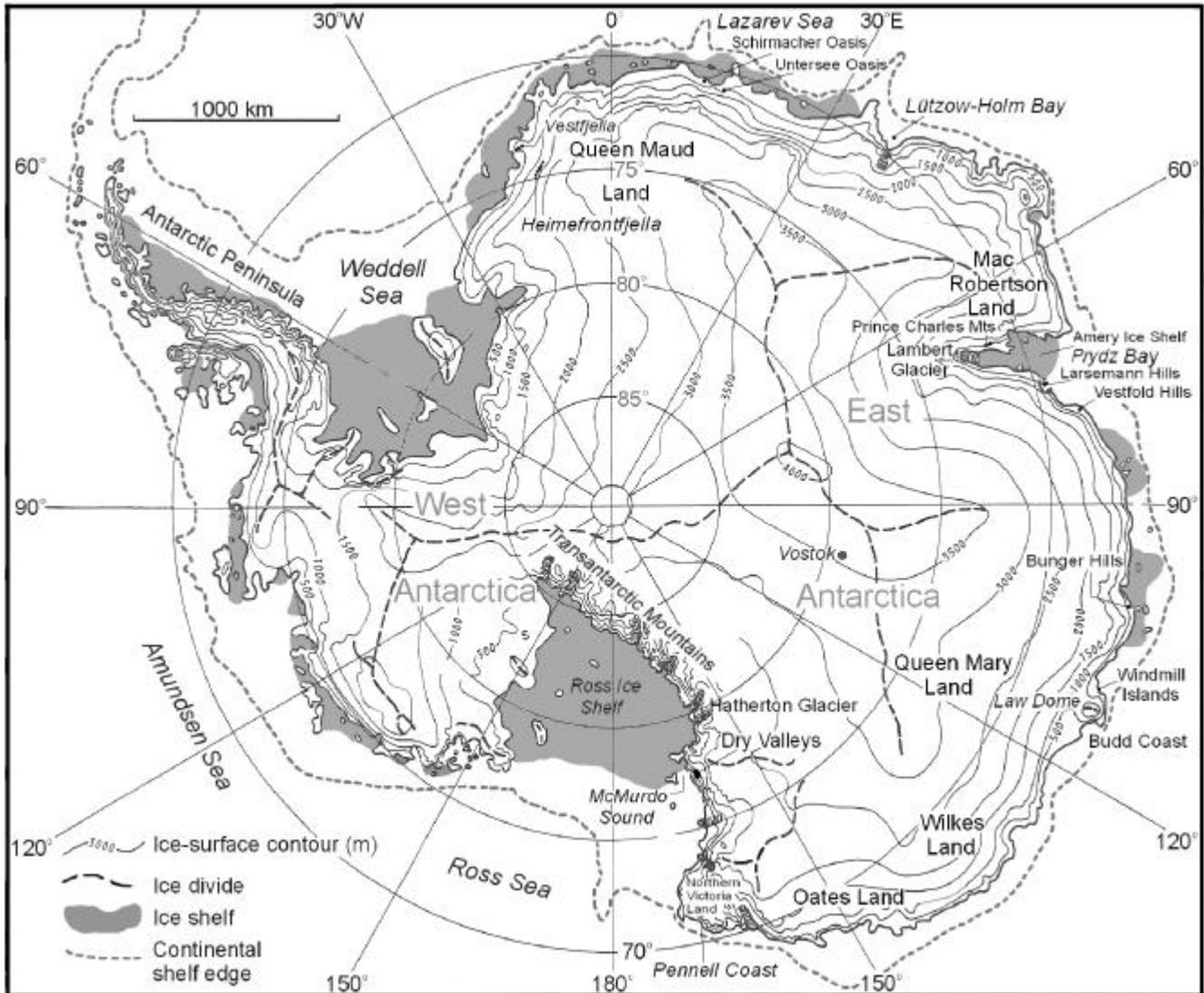


Figure 1. Map of Antarctica and its continental shelf.

and the following review relies heavily on overviews by Denton *et al.* (1991), Kennett & Hodell (1993) and Anderson (1999).

2.1. Onset of Antarctic glaciations.

The onset of glaciations in Antarctica in the mid-Tertiary was probably related to the breakup of Gondwana, poleward drift of Antarctica and the development of ocean passages around the continent (Kennett, 1977). Plate-tectonic and palaeoceanographic isolation of the Antarctic continent successively led to cooling and glaciations. Macro- and microfossil evidence from Late Cretaceous and Paleocene deposits on James Ross Island and Vega Island (Fig. 5) suggest that climate there was warm to cool temperate and humid (Askin, 1992). ODP results from the Queen Maud Land margin suggest temperate to subtropical Late Cretaceous climate (Kennett & Barker, 1990), and

ODP-data from northern Weddell Sea suggests bottom and surface waters there were relatively warm throughout most of Paleocene (Kennett & Stott, 1990; Robert & Kennett, 1994). There is no conclusive evidence for the existence of a large ice sheet on Antarctica during the Late Cretaceous or early Tertiary. Fluctuations in global sea level during late Paleocene might, however, record ice-sheet fluctuations in interior East Antarctica (Anderson, 1999).

Deep-sea oxygen isotope records show a steady increase in $\delta^{18}\text{O}$ concentrations during the Eocene, which have been interpreted to signify buildup of ice-sheet in Antarctica (Prentice & Mathews, 1988; Denton *et al.*, 1991; Abreu & Anderson, 1998). Seismic records of a major unconformity related to glacial diamictites in ODP drill cores from the East Antarctic continental shelf provide strong evidence for an ice sheet on East Antarctica by Late Eocene to Early Miocene (Fig. 3) (Barron *et al.*, 1991; Andersen, 1999). Birkenmajer (1988, 1991) described Eocene glacial deposits from King George Island (Fig. 5). These probably

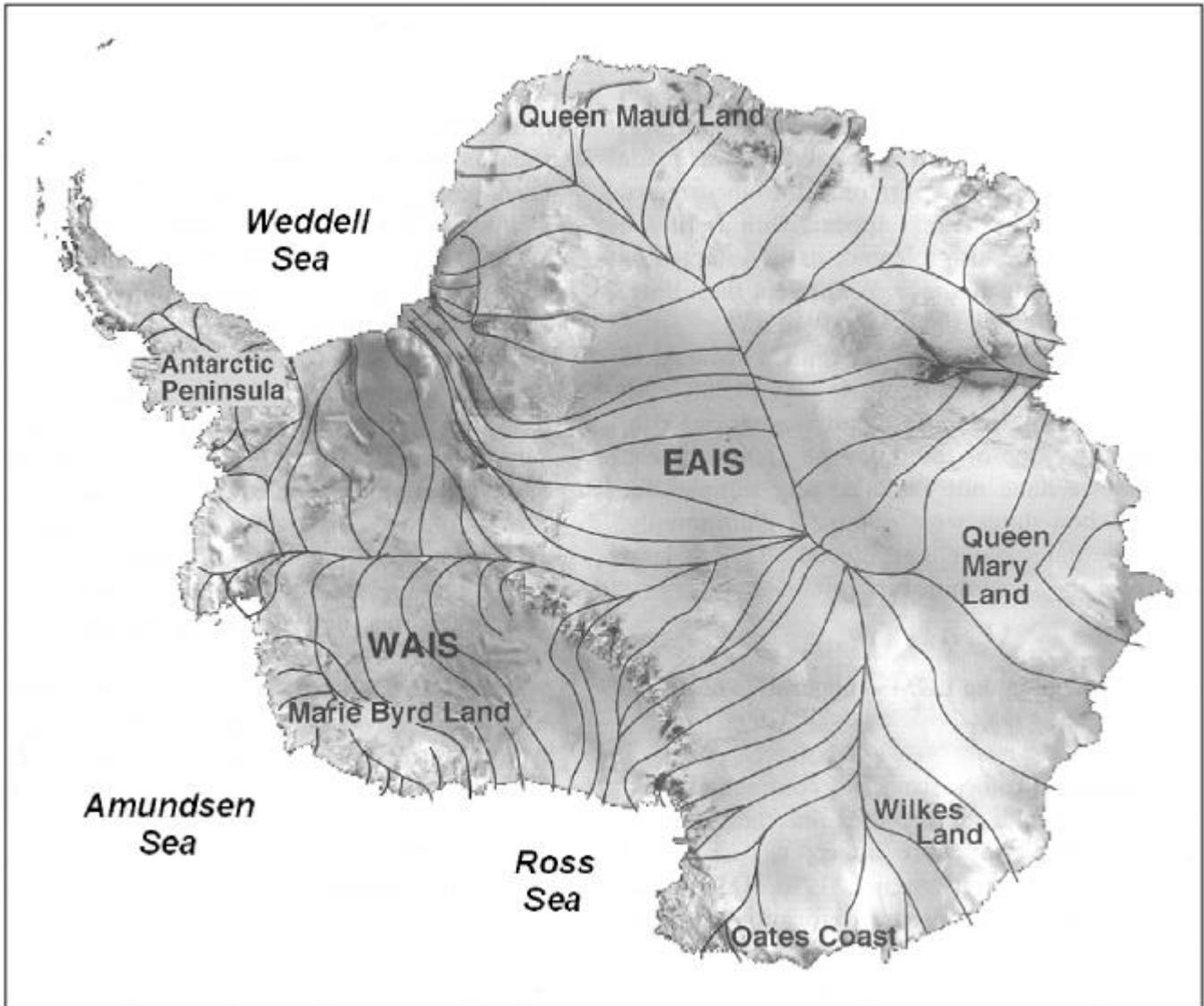


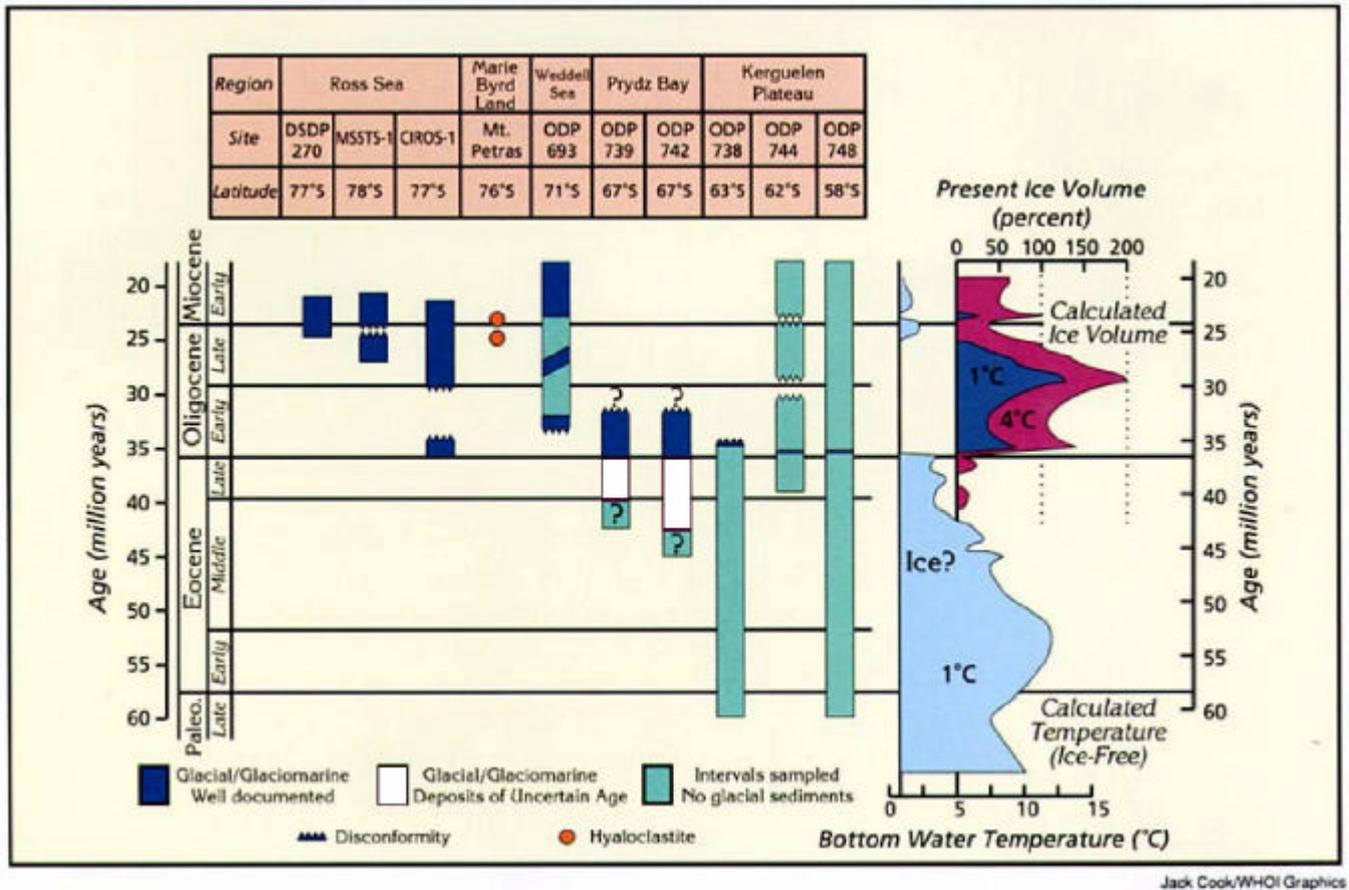
Figure 2. Drainage map of the Antarctic ice sheet. Modified after Anderson *et al.* (2002).

relate to a local glaciation rather than an extensive West Antarctic glaciation, since other data from that region suggest relatively warm Eocene conditions (Askin, 1992). There is strong evidence for an ice sheet on East Antarctica during the Oligocene (Denton *et al.*, 1991; Hambrey *et al.*, 1991), and ice spread into the western Ross Sea by the Late Oligocene (Hambrey, 1993; Wilson *et al.*, 1998). On West Antarctica, there existed at least mountain glaciers and localized ice caps (Birkenmajer, 1998; Anderson, 1999), but the existence of an Oligocene WAIS has not been proven.

2.2 Miocene – Extensive glaciations in both East and West Antarctica

Deep-sea oxygen isotope records show stepwise increase in $\delta^{18}\text{O}$ concentrations during the Miocene and into the

Pliocene, which have been interpreted to signify step-like build-up of ice volumes on Antarctica (Miller *et al.*, 1991). Results of ODP drillings, seismic records and terrestrial stratigraphical records provide strong evidence for the presence of large ice sheets on both West and East Antarctica during the Miocene (Hambrey *et al.*, 1991; Anderson, 1999). Miocene glacial and glaciomarine strata recorded in drill sites on the Ross Sea continental shelf and in the westernmost Ross Sea are interbedded with meltwater deposits and diatomaceous oozes, indicating shifts from temperate to sub-polar or polar climates throughout the Miocene (Anderson, 1991). The recent Cape Roberts Project provides evidence for the expansion of polythermal glaciers from the Transantarctic Mountains towards and beyond the Cape Roberts drill-site (Powell *et al.*, 1998), as well as subsequent ice recession during the younger part of the Miocene. Larter & Barker (1989, 1991) and Bart & Anderson (1995) provided data to suggest that



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Figure 3. The ODP record of Tertiary glaciations in Antarctica (downloaded from: <http://www.oceandrilling.org/Documents/Oceanus/Paleo/OceanusPal16HR.html>).

ice had expanded out on the Antarctic Peninsula continental shelf by Middle or Late Miocene, and there is also onshore evidence for large Miocene ice caps in the Antarctic Peninsula (Birkenmajer, 1988). Abreu & Anderson (1998) suggested that that the WAIS was probably grounded mostly above sea level, and therefore contributed to Miocene eustatic sea level changes.

2.3 Pliocene and the transition to the Pleistocene: controversies on ice volumes, ice sheet stability and continental climate

The deep-sea record from around Antarctica suggests significant Pliocene sea surface temperature fluctuations, and the oxygen isotope and eustatic records indicate that considerable ice volume fluctuations may have occurred (e.g. Prentice & Fastook, 1990; Denton *et al.*, 1991; Dowsett *et al.*, 1996; Flemming & Barron, 1996). The continental seismic shelf records from the Ross Sea region and Antarctic Peninsula continental shelf show a number of shelf-wide unconformities bounding till sheets, indicating high-frequency grounding events, taken to represent waxing and waning of continental glaciers during the Pliocene (Alonso *et al.*, 1992; Bart & Anderson, 1995). There is,

however, no clear evidence, in the form of delta deposits or river-valleys, showing significant meltwater discharge in connection with those oscillations (Anderson, 1999). Core data from western (near-shore) Ross Sea shows meltwater deposits interbedded with tills (Barrett *et al.*, 1992).

The Pliocene continental record is fragmentary and difficult to interpret in terms of climate and glaciation development. Consequently it has led to considerably different opinions as to the Miocene-Pliocene-Pleistocene stability of the Antarctic Ice Sheet (Wilson, 1995), and the nature, timing, extent and evolution of the ice sheets. Reconstructions of the history of the EAIS range from drastic ice-volume fluctuations, with the ice sheet fluctuating between extensive collapse and very extensive build-up and mountain overriding (Barrett *et al.*, 1992; Barrett, 1996), to suggestions that it has existed close to its present configuration for the past c. 14 million years (Shackleton & Kennett, 1975; Sugden *et al.*, 1993; Denton *et al.*, 1993). The greatest controversy focuses around evidence for a major deglaciation in East Antarctica and different interpretations of the fossil evidence from the so-called 'Sirius Group' (Mercer, 1972, 1981; Mayewski, 1975; Wilson, 1995). It consists mainly of tillites, which occur above 2000 m a.s.l. in the Transantarctic Mountains. The tillites suggest that some time during the Pliocene the

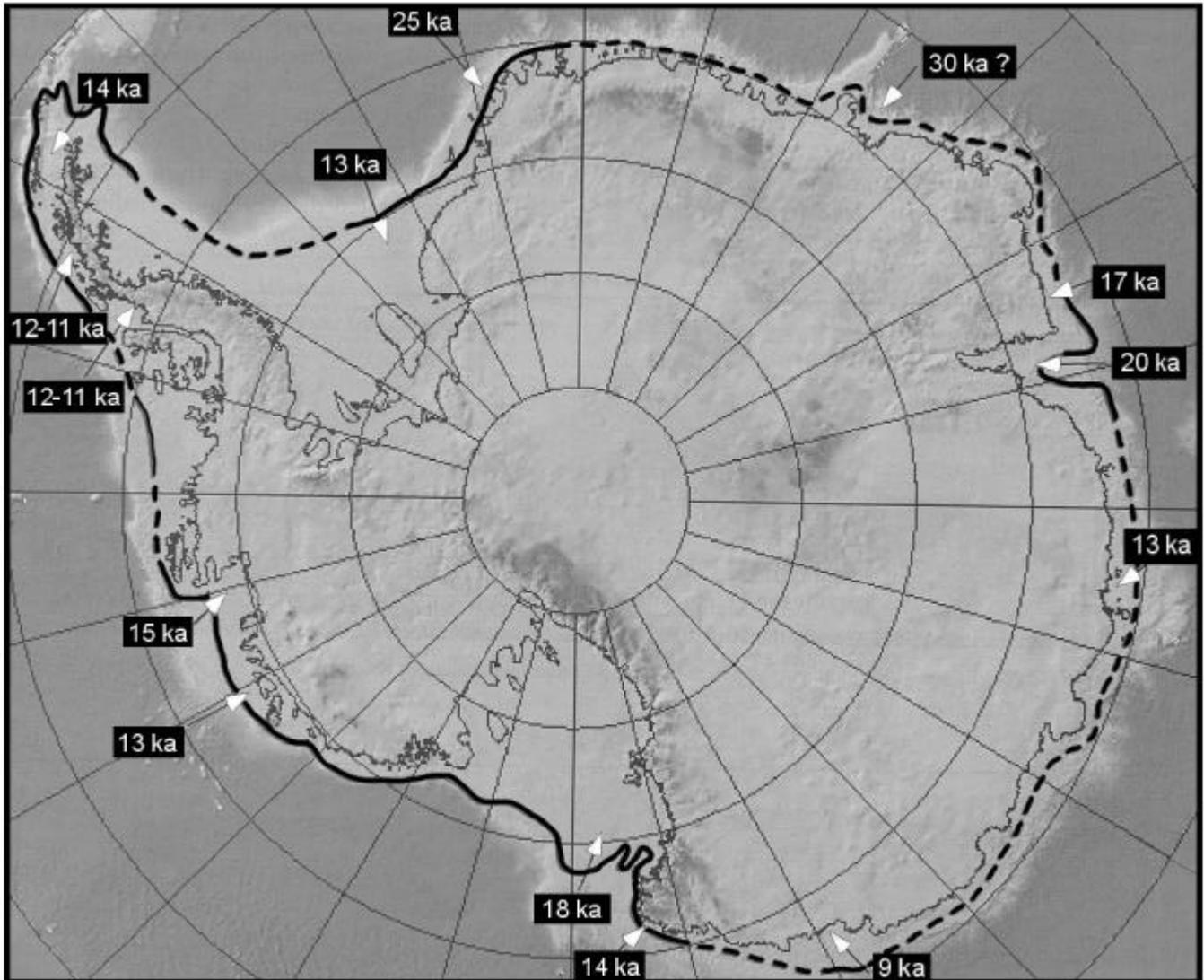


Figure 4. A reconstruction of grounding line positions of the Antarctic LGM ice sheet. Oldest constraining minimum ages for initial ice retreat are given. Slightly modified after Anderson *et al.* (2002).

Transantarctic Mountains were covered by an ice sheet considerably thicker than the present (Mayewski, 1975; Denton *et al.*, 1991). However, included in the Sirius sediments are palaeosols and plant remains, both leaves and *in situ* roots of southern beech, *Nothofagus* sp., at c. 1800 m a.s.l., suggesting a much-warmer-than-present situation. One major point of controversy is the occurrence of Late Pliocene-Early Pleistocene diatoms in the strata. Do they originate from interior basins of East Antarctica, implying a non-glacial situation there, or have they been reworked (windblown) into the strata? There are numerous other aspects to this discussion (for reviews see Denton *et al.*, 1991; Wilson, 1995; Barrett, 1997; Anderson, 1999), and it has not been resolved if the ice sheet underwent such massive changes in ice volumes as for interior East Antarctica to become ice free, alternating with very extensive glacial overriding of the Transantarctic Mountains. There is also discussion on the Pliocene

climate; how warm or cold conditions do the *Nothofagus* remains represent? Hill & Trustwell (1993) proposed that the *Nothofagus* vegetation suggested summer temperatures of up to +8°C whilst Francis (1995) suggested that could be compared with the modern willow-shrub vegetation of the Arctic tundra, signifying conditions with annual mean temperatures below freezing, permafrost conditions, but seasonal moisture supply. This implies considerably warmer Pliocene interglacial conditions than at present, and raises the question whether the present polar conditions became established first in the Early Pleistocene, after 2.5 Ma (Webb & Harwood, 1991)?

The Pliocene history of the WAIS is scantily known. Reconstructions are heavily dependent on the East Antarctic records, i.e. whether or not they encompass a major deglaciation of interior East Antarctica or not (Denton *et al.*, 1991). Based on seismic records from the Antarctic Peninsula continental shelf there were repeated

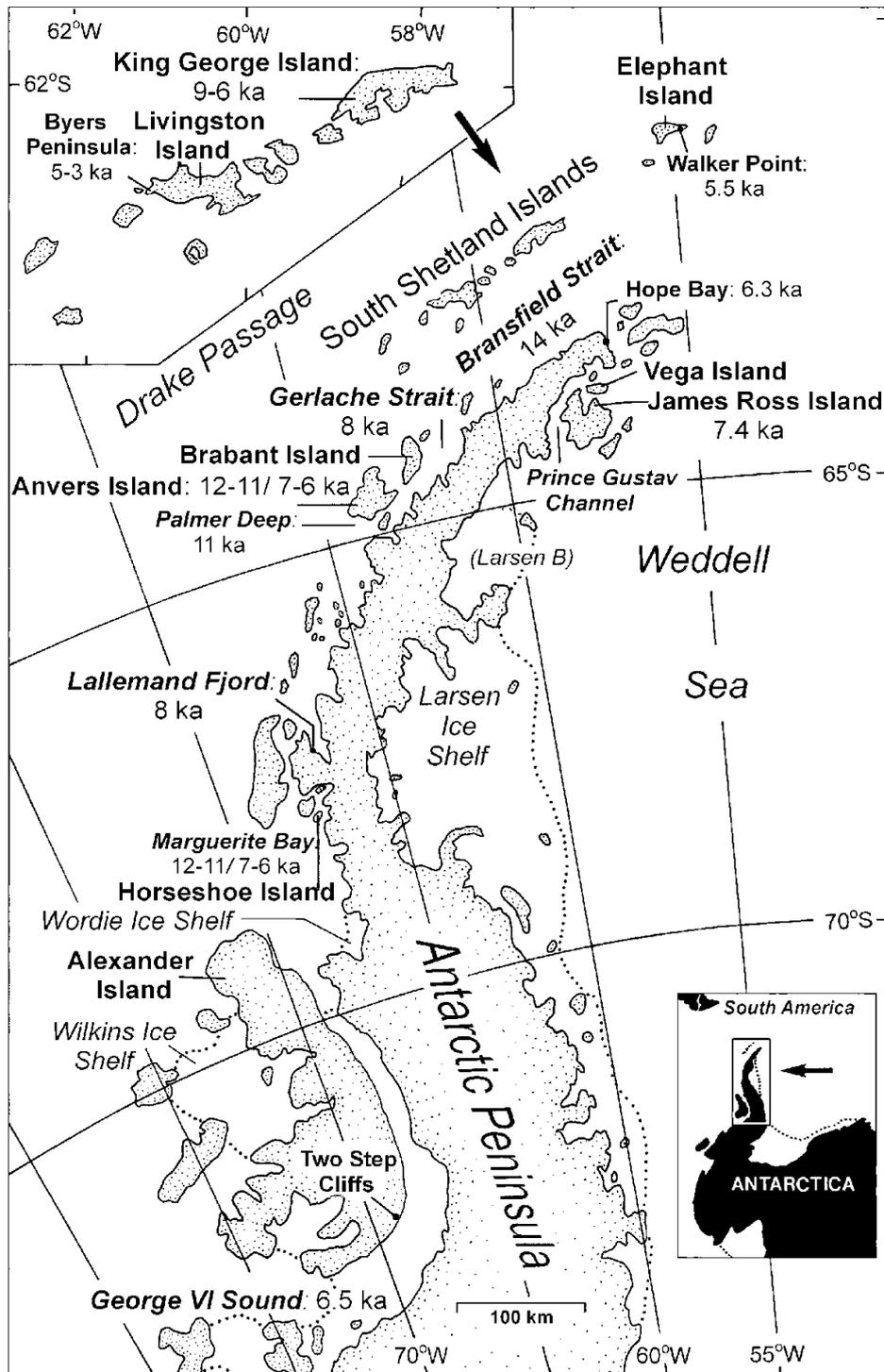


Figure 5. Location map for the Antarctic Peninsula. Minimum ages for ice sheet retreat.

fluctuations in West Antarctic ice volumes (Bart & Anderson, 1995) during the Pliocene-Pleistocene.

The combined continental and shelf evidence suggests that during late Pliocene and perhaps Early Pleistocene glacial episodes, ice sheets advanced to the edge of the continental shelf around Antarctica, with ice extent similar to or greater than during the last (Wisconsinan) glaciation, and that those glaciations were interrupted by interglacial episodes that were warmer than the present. Denton *et al.*

(1991) suggested that the deep-sea $\delta^{18}\text{O}$ record could be interpreted to indicate that the overall Antarctic ice volume was never significantly less than today during the Pliocene.

3. The Pleistocene in Antarctica – a brief overview

The continental Pleistocene record of glacial and climatic events in Antarctica is fragmentary, at best. In East

Antarctica, outcrops with Pleistocene deposits are confined to few localities in the Transantarctic Mountains and at some few-and-far-between coastal sites. Data from the sites in the Transantarctic Mountain have been interpreted to record a major cooling in the Early Pleistocene (Denton *et al.*, 1986; Denton *et al.*, 1989). Lithostratigraphical and sedimentological evidence from Pleistocene drift deposits, as well as geomorphological evidence in the form of trimlines and glacial striae at high altitudes above sea level indicate that the EAIS overrode the Transantarctic Mountains during several episodes of extensive glaciations (Denton *et al.*, 1991). The West Antarctic continental record is equally fragmentary, but available stratigraphical and geomorphological evidence suggest that the WAIS was considerably thicker than at present on several occasions during the Early and Middle Pleistocene, and probably grounded on the shelf edge (Denton *et al.*, 1991; Anderson, 1999). The continental data corroborates with seismo-stratigraphic profiles and core data from the shelf areas of the Antarctic Peninsula, Weddell Sea and the Ross Sea. These data are interpreted to show numerous stacked till sheets, interbedded with thin and discontinuous glaciomarine and meltwater sediments, indicating subglacial and grounding zone-proximal deposition (McKelvey, 1991; Shipp *et al.*, 1994; Bart & Anderson, 1995).

The available data strongly suggests that the Antarctic ice sheet was considerably thicker than present on several occasions during the Pleistocene, with circum-Antarctic ice grounded out on the continental shelf edge. The debate concerning when relatively stable polar conditions were established in Antarctica does not concern the Pleistocene; most workers agree that a major cooling was at least established in Antarctica by the Pliocene-Pleistocene transition (Anderson, 1999). The controlling mechanism for the waxing and waning of the Antarctic ice sheet is not well understood: isotope records from deep-sea sediment cores off Antarctica (e.g. Mackensen *et al.*, 1989; Grobe & Mackensen, 1992) and long Antarctic ice cores (e.g. Jouzel *et al.*, 1987; Petit *et al.*, 1997) show strong correlations with the global isotope record (Prell *et al.*, 1986). This has been taken to indicate that there are linkages between Northern and Southern Hemisphere ice sheet developments during the Quaternary. Denton *et al.* (1991) concluded that there is an overall in-phase relationship between glacial cycles in the Northern and Southern hemispheres, perhaps driven by falling and rising global sea levels regulated by build-up and decay of Northern Hemisphere ice sheets (Hollin, 1962; Thomas & Bentley, 1978).

4. Late Quaternary glacial development in Antarctica

The late Quaternary stability and fluctuations of the marine-based ice sheet in West Antarctica have been debated during the past decades. Mercer (1968) suggested that the WAIS had repeatedly collapsed during Pleistocene interglacials, and Scherer *et al.* (1998) suggested, on the

basis of the existence of marine sediments below present-day West Antarctic ice streams, that the WAIS had collapsed at least once during late Pleistocene. The oceanic $\delta^{18}\text{O}$ record suggests that between 150-120 ka BP there was less global ice volume than at present, and it has been suggested that collapse of the WAIS caused global sea level to be 5-6 m higher than present during Marine Isotope Substage 5e (Mercer, 1978). If this signifies the collapse or decay of the WAIS (Mercer, 1968), or if the water-source was the Greenland ice (Emiliani, 1969; Cuffey & Marshall, 2000), or it came from melting of the Northern Hemisphere ice sheets and partial melting of the Antarctic ice sheet (Denton *et al.*, 1971; Hughes, 1987), remains unresolved.

5. Methods for reconstructing Antarctic glacial history since Last Glacial Maximum (LGM)

5.1 Terrestrial evidence for the extent of LGM ice and subsequent deglaciation

The terrestrial evidence for the extent of LGM ice around Antarctica comes from a number of archives:

- Extensional and directional evidence on land: evidence of more extensive ice cover than today is present on almost every ice-free lowland area around Antarctica and its outlying islands, in the form of glacial drift, erratics and striations. Directional evidence in general suggests ice flow offshore towards the shelf areas around Antarctica.
- Evidence on ice thickness and ice movements at the glacial maximum: these include ice-abraded ridge crests, trimlines, striations, erratics and glacial drift on nunataks and coastal mountains. In general these suggest considerable thickening of LGM ice in the coastal regions compared to the present.
- Lithostratigraphical data: good stratigraphical sections are extremely rare on Antarctica. High permafrost table prevents digging or excavating, and there is a lack of natural exposures in an environment where eroding rivers are almost totally lacking and extensive sea-ice cover and perennial ice-foot shelter most beaches from erosion. The stratigraphical data available, from natural sections and boreholes, mostly reflects environmental development subsequent to the last deglaciation.
- Raised beaches: raised beaches around Antarctica can generally be regarded as isostatic fingerprinting of earlier expanded ice volumes compared to present. Raised marine beach deposits that post-date the last major glaciation in Antarctica have been described from many ice-free areas on the continent and nearby islands. Because of the late deglaciation of coastal areas and still extensive glaciation of the continent, the elevation

of Holocene marine limit (ML) is much lower than corresponding limits generally are in the Northern Hemisphere. The highest ML described on Antarctica is from Horseshoe Island in Marguerite Bay, off the Antarctic Peninsula, at 55 m a.s.l. (Hjort & Ingólfsson, 1990). The ML on the outer coast around Antarctica is generally below 30-35 m a.s.l., whereas it usually is below 15 m a.s.l. in coves and bays more proximal to the ice sheet (Ingólfsson *et al.*, 1998; Berkman *et al.*, 1998). The age of the Holocene ML around Antarctica is not synchronous, but reflects differences in the timing of regional and local deglaciation. At most sites, however, it is constrained by minimum dates between 7-5 ka BP.

5.2 Marine geological evidence for former ice-extent

Archives containing data on former ice-extent (Anderson, 1999; Anderson *et al.* 2002) include:

- Shelf bathymetry; glacial troughs, submarine valleys, 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.
- 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.
- 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.

5.3 Ice-core evidence on glacial history

Antarctic ice cores provide long, low-resolution proxy records of ice volume changes for the past *c.* 450,000 years (Petit *et al.*, 1997), and inferred information on summer temperatures and ice thickness (Lorius *et al.*, 1985, 1993). The long Vostok record shows a strong correlation with the global isotopic record, which has been taken to indicate strong linkages between Northern and Southern Hemisphere ice sheets (Petit *et al.*, 1997). There are, however, large discrepancies between Late Quaternary glacial and climate history as reconstructed from geological evidence from coastal terrestrial or offshore records, and ice core records (Jouzel *et al.*, 1987; Petit *et al.*, 1997; Ingólfsson *et al.*, 1998, Ingólfsson & Hjort, in press; Anderson, 1999). This summary focuses on the stratigraphical terrestrial and offshore geological record as

the most reliable proxies for the glacial history of Antarctica.

5.4 Chronological control

The age control on glacial and climatic events since the LGM in Antarctica is primarily through radiocarbon dating. Marine materials such as 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 dated 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 (*cf.* Björck *et al.*, 1991a; Gordon & Harkness, 1992; Andrews *et al.*, 1999; Domack *et al.*, 2000). 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, 1996; Doran *et al.*, 1999; Denton & Hughes, 2000). Samples of moss-bank peat on Elephant Island (Fig. 5) have been found to give some of the most reliable radiocarbon ages in Antarctica (Björck *et al.*, 1991a, 1991b), and are thus optimal for constructing a radiocarbon 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 on 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 resulting in 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 years.
- b) Reduced gas exchange with the atmosphere resulting from the long (in extreme cases perennial or decadal) duration of the ice coverage 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 Bunge 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 radiocarbon 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 BP (corrected by 1300 radiocarbon years for marine reservoir age) and the algae 5930 ± 200 BP, which Stuiver *et al.* (1981) thought could reflect contamination by carbonate from local marble bedrock. Doran *et al.* (1999) concluded that dates of algae that grew on proglacial lake bottoms may require corrections because of the relict dissolved inorganic carbon introduced into the lake by inflow from glacier margins, but that this was not a problem when dating near-shore microbial mats in deltas fed by meltwater streams.
- e) Continuous erosion of lake bottom surface sediments resulting from bottom-freezing in winter, or oxidation of these surface sediments due to periods of desiccation, are processes which could lead to erroneous radiocarbon 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 years (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 stratigraphical situations.

- h) Blanketing snow and ice on perennially frozen lakes could hinder lacustrine sedimentation, causing hiatus between the deglaciation and onset of biogenic deposition, resulting in too young minimum dates for deglaciation (Gore, 1997).

Radiocarbon dates from organic remains (microbial mats, algae, water mosses) in sediment cores sampled from the present Lake Hoare in Taylor Valley, Dry Valleys, southern Victoria Land (Fig. 1), have revealed that there occur large contamination problems (Squyres *et al.*, 1991). These were expressed as very old ages of surface sediments (varying between, 2000 and 6000 radiocarbon years) and samples obtained from deeper in the cores yielded ages similar or younger than the surface material. Squyres *et al.* (1991) concluded that radiocarbon dates from 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 ages of radiocarbon dates. A similar explanation is possible for radiocarbon ages of 24 ka and 35.7 ka BP from the base of marine inlet and lake cores in Bunge Hills, East Antarctica (Melles *et al.*, 1997). In the fresh-water Lake Untersee (Queen 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 ka years (M. Schwab, personal communication, 1998, quoted in Ingólfsson *et al.*, 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 radiocarbon 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 $\delta^{13}\text{C}$ -value should always be measured and used to correct the $^{14}\text{C}/^{12}\text{C}$ 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 Shetlands, 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 deepwater 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 years (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 radiocarbon-depleted CO₂ 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, while Barsch & Mäusbacher (1986), working on the South Shetland Islands (Fig. 5) used an envelope of 850-1300 years. Ingólfsson *et al.* (1992) and Hjort *et al.* (1997) applied a sea correction of 1200 years, while Pudsey *et al.* (1994) used a reservoir correction of 1500 years. 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) radiocarbon-dated stomach oil deposits in snow petrel colonies in Bunger Hills. They based their chronology for petrel colonisation on conventional radiocarbon dates, but stated that a reservoir correction of 1300 years probably was appropriate. Bird *et al.* (1991) applied a correction for reservoir effects for marine sediments of 2200 years when studying evolution of marine inlets in the Vestfold Hills oasis, East Antarctica. In the Victoria Land/Ross Sea area Stuiver *et al.* (1981) and Denton *et al.* (1989) based their chronology on uncorrected radiocarbon 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 radiocarbon 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 radiocarbon 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. Gingele *et al.* (1997) used a reservoir correction of 1550 years when reporting accelerator mass spectrometer (AMS) radiocarbon ages from the Lazarev Sea. Considerable efforts are being devoted to solving problems concerning radiocarbon dating of Antarctic marine sediments, including dating acid-insoluble organic matter and foraminiferal calcite (Andrews *et al.*, 1999; Domack *et al.*, 2000, 2001a).

A number of investigations have assessed which correction of radiocarbon ages of Antarctic marine organisms has to be applied in order to establish a coherent radiocarbon chronology for Late Wisconsinan-Holocene glacial events. Circumantarctic studies generally show the average correction for reservoir age of marine mollusc shells to be 1100-1400 radiocarbon years (Yoshida & Moriwaki, 1979; Stuiver *et al.*, 1981; Omoto, 1983; 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 years (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 material (flesh, bone, feathers, guano) is dated can significantly influence the correction factor required (Mabin, 1986; Baroni & Orombelli, 1991). Berkman & Forman (1996) suggested that reservoir corrections of 1300±100 years (molluscs), 1424±200 years (seals) and 1130±130 years (penguins) should be applied to Antarctic marine organisms.

This paper adopts 1300 radiocarbon years as the best estimate for a circumantarctic correction for all radiocarbon-dated marine organisms, for the sake of comparing glacial histories of the different areas. All radiocarbon ages of fossil marine organisms 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 radiocarbon years BP, reported as kilo-annum before present (ka BP). At the same time it should be stressed that there are still large uncertainties in the Antarctic marine reservoir effect.

6. Last Glacial Maximum in Antarctica

The overall extent of ice cover in Antarctica during LGM is not well known and some existing reconstructions are controversial (reviews in Denton *et al.*, 1991; Ingólfsson *et al.*, 1998; Anderson *et al.*, 2002). 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 shelf break around most of Antarctica (Denton, 1979; Hughes *et al.*, 1981; Clark & Lingle, 1979; Denton *et al.*, 1991; Zhang, 1992). Other reconstructions indicate less ice extent. Mayewski (1975) maintained that the WAIS was only somewhat, 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 extent was insignificantly further than today in some areas (Hayashi & Yoshida, 1994). A recent study by Anderson *et al.* (2002) found that the East and West Antarctic ice sheets have not advanced and retreated in concert, and whereas the LGM WAIS advanced for the

Figure 6. Crystalline erratics from the Antarctic Peninsula on northern James Ross Island
(Photograph: Ó. Ingólfsson, 1989).



most part to the outer shelf, the EAIS did not expand to the continental shelf edge during the LGM (Fig. 4).

The timing of the last maximum glacial expansion around Antarctica is likewise poorly known, and though most authors assume it to have coincided with the timing of lowest global sea levels during MIS 2, around, 20-18 ka BP, circum-Antarctic field data suggest considerable regional differences in glacial maximum extension between >20-14 ka BP (Stuiver *et al.*, 1981; Domack *et al.*, 1995; Anderson *et al.*, 1992; Kellogg *et al.*, 1996; Ingólfsson *et al.*, 1998; Hall & Denton, 1999, 2000a; Denton & Hughes, 2000; Denton & Marchant, 2000). Maximum ages constraining ice-sheet grounding in the western Ross Sea are >24 ka BP (Denton & Hughes, 2000) and maximum extension of ice in the western Ross Sea area has been dated to *c.* 23-17 ka BP (Stuiver *et al.*, 1981; Anderson *et al.*, 1992; Kellogg *et al.*, 1996; Licht *et al.*, 1996; Denton & Hughes, 2000). Hall (1997) and Hall & Denton (2000a) have recently dated the maximum extent of ice in Taylor Valley, southern Victoria Land, to 14.6-12.7 ka BP. In East Antarctica, on the coast of Mac Robertson Land and at Prydz Bay (Fig. 1) the last maximum ice extent occurred before 17 ka BP and 10.7 ka BP, respectively (Domack *et al.*, 1991a, Taylor & McMinn, 2001). It has been suggested that the last maximum ice extent in the Antarctic Peninsula region occurred later than 30 ka BP (Sugden & Clapperton, 1980) and prior to 14-12 ka BP (Banfield & Anderson, 1995, Domack *et al.*, 2001a). Recent interpretations, however, suggest that the most recent Wisconsinan ice sheet maximum there may pre-date the, 20-18 ka BP LGM: Bentley & Anderson (1998) suggested that ice volumes were considerably greater before 35 ka BP compared to after 35 ka BP, and likewise, Anderson & Andrews (1999) suggest that ice extent towards the continental slope in the

Weddell Sea occurred considerably earlier than 26 ka BP. 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. A reconstruction of late Quaternary sea-ice distribution around Antarctica suggests that during LGM (MIS 2) the Antarctic winter sea-ice was extended towards the present marine polar front zone (Antarctic Convergence), whereas summer sea-ice was expanded into the area of the present winter sea-ice edge (Gersonde & Zielinski, 2000).

6.1 LGM-extent of ice in the Antarctic Peninsula region – terrestrial evidence

The Antarctic Peninsula (Figs. 2 and 5) ice sheet is a part of the marine-based WAIS, 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 shelf, that during 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 modern -200 m shallow submarine platforms around the islands and along the Antarctic Peninsula roughly coincide with the ice extent. 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 (Fig. 6) 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), whereas ancient glacial trimlines are rarely described. 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 & Sugden, 1982; Waitt, 1983; Bentley & Anderson, 1998; Bentley, 2000) and in the South Shetland Islands (John & Sugden, 1971; Sugden & John, 1973; Curl, 1980). Carrara (1979, 1981) and Waitt (1983) concluded that striations and erratics on nunataks on the southern Antarctic Peninsula showed that some time in the past they were overridden by a considerably thickened ice sheet, at least 500 m thicker than today (Waitt, 1983; Bentley & Anderson, 1998). Bentley *et al.* (2000) suggested that in the southern and central parts of the Antarctic Peninsula, the Wisconsinan ice sheet some time prior to 35 ka BP had expanded to form two ice domes and probably merged with an expanded WAIS in the Weddell Sea. Based on cosmogenic exposure ages from erratics, they suggested ice-sheet lowering in the southern part of the Antarctic Peninsula, indicating deglaciation, already by *c.* 16 ka BP. Bentley & Anderson (1998) and Bentley (1999) reconstructed the LGM configuration of an Antarctic Peninsula ice sheet, with ice flowing out towards the shelf edges from an ice-divide stretching N-S along the axis of the peninsula.

A number of additional investigations interpret former ice-thickness on the northern part of the Antarctic Peninsula region and the islands in the western Weddell Sea (e.g. 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 NW Weddell Sea had, at least once, been overridden by glaciers advancing from SW-W, from the Antarctic Peninsula, depositing tills and crystalline erratics up to 300-400 m a.s.l. Hjort *et al.* (in press) suggested that this period of extensive glaciation was contemporaneous with Bentley & Anderson's (1998) LGM-grounding line position some, 200 km further east, but was probably older than 30 ka BP. They suggested that the ice flowing off the peninsula during the LGM was thinner, channeled through the straits and troughs, and did not attain much greater thickness there than *c.* 150 m a.s.l.

Raised beaches on the Antarctic Peninsula can generally be regarded as isostatic remnants of earlier expanded ice volumes compared to present. Raised marine beach deposits, which post-date 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

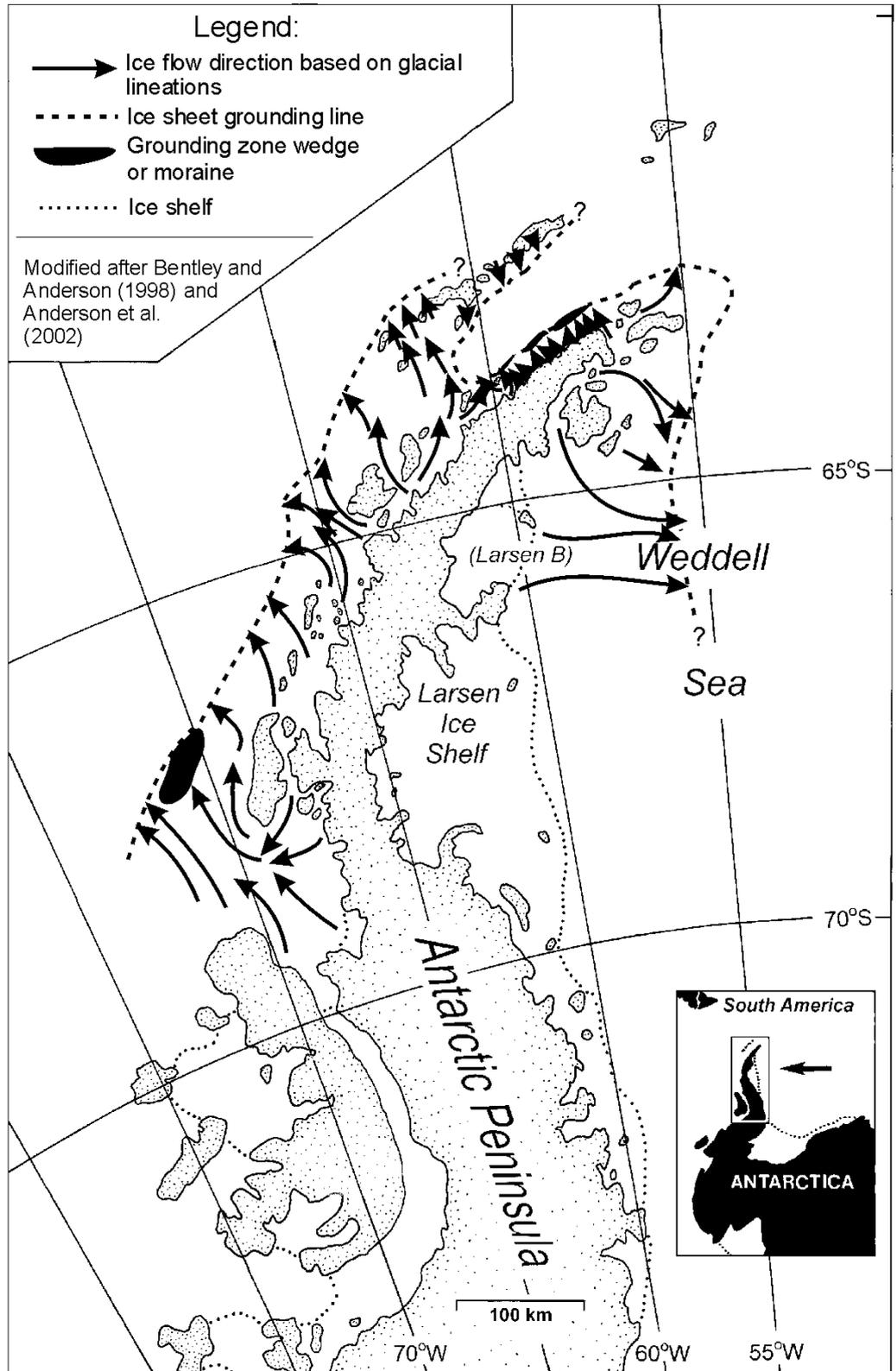
& Sugden, 1971; Curl, 1980; Birkenmajer, 1981; Hjort & Ingólfsson, 1990; Ingólfsson *et al.*, 1992; López-Martínez *et al.*, 1996; Hjort *et al.*, 1997). They imply expanded ice volumes prior to the early-mid Holocene, compared to the present. Since there are as yet no well-constrained relative sea-level 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 (ML) generally lies at altitudes between 15-20 m a.s.l., with the highest ML on James Ross Island at 30 m a.s.l. (Hjort *et al.*, 1997). There it is dated to *c.* 7.5 ka BP. The ML rises towards south, and is highest on Horseshoe Island in Marguerite Bay, at *c.* 55 m a.s.l. (Hjort & Ingólfsson, 1990). The Horseshoe Island ML is not absolutely dated, but is constrained by minimum ages for the deglaciation of Marguerite Bay as being older than 7-6 ka BP (Kennedy & Anderson, 1989; Harden *et al.*, 1992; Pope & Anderson, 1992; Emslie, 2001). The north-south gradient in altitude of the marine limits agrees with the conclusion of Bentley & Anderson (1998) and Anderson *et al.* (2002) that there were greater LGM ice volumes in the central and southern part of the region than in the northern part.

6.2 LGM extent of ice in the Antarctic Peninsula area – marine evidence

Ice extent on the western continental shelf - Kennedy and Anderson (1989) and Anderson *et al.* (1991a) 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 grounded-ice-to-recessional-ice scenario for Marguerite Bay, based on microscopic analyses of glacial 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 mega-scale glacial lineations in the seawards portion of the trough, extending into a prominent grounding zone wedge on the mid-shelf.

Pudsey *et al.* (1994) published side-scan sonar records offshore Anvers Island showing 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 ¹⁴C 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

Figure 7. LGM palaeo-drainage and grounding line positions off the Antarctic Peninsula. Drawn after Anderson et al. (2002).



the Palmer Deep also suggests that the shelf close to Anvers Island was covered by ice prior to ca. 11 ka BP.

Banfield & 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 & Anderson (1998) suggest that these ridges are moraines, marking the LGM grounding line (Fig.

7). A recent swath bathymetric documentation by Canals *et al.* (2000), showed that during a past glacial maxima a major ice stream had flowed off the north-western side of the Antarctic Peninsula, depositing an >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 ¹⁴C-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 northwards into the Bransfield basin.

Bentley & Anderson (1998) and Anderson *et al.* (2002) reviewed available data on LGM grounding line positions. They concluded that LGM ice flowing west from the Antarctic Peninsula had generally grounded on the middle-outer shelf, at water depths of ca. 400 m (Figs. 4 and 7), and that Antarctic Peninsula ice had met with a local ice cap over the South Shetland Islands.

Ice extent in the northwestern Weddell Sea – Because of the inaccessibility, the geological evidence from the western Weddell Sea is scarcer than from west of the peninsula. Sea ice has long effectively prevented the acquisition of marine geological information south of 65°S, but some is forthcoming after successful cruises to Prince Gustav Channel and Larsen Ice Shelf in, 2000 (Pudsey & Evans, 2001; Pudsey *et al.*, 2001; Domack *et al.*, 2001b). 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 & Anderson (1998) suggested, on the basis of high-resolution seismic records and sediment cores, that Prince Gustav Channel and adjoining bays contain only 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 mid-shelf. They suggested that the unconformity marked the extension of Antarctic Peninsula ice sheet, and that it had grounded on the outer continental shelf, some 200 km east of the peninsula, during recent glaciation (Figs. 4 and 7). However, the age of the unconformity is uncertain since it is not constrained by any available dates (Anderson *et al.*, 2002) but Bentley & Anderson (1998) and Anderson *et al.* (2002) correlated it to the LGM.

Taken together, the marine and terrestrial geological data on LGM ice flow and drainage can be interpreted to suggest that the LGM Antarctic ice sheet was composed of a number of confluencing, localized domes and ice stream systems, rather than a concentric and dynamically coherent ice sheet.

6.3 LGM extent of ice in East Antarctica

The extent of ice during the LGM in East Antarctica is not well known and partly controversial. This is mainly due to lack of detailed marine geological surveys aimed at constraining the LGM configuration of ice on the East Antarctic continental shelf (Anderson *et al.*, 2002). A maximum reconstruction extends ice to the shelf break (Denton, 1979; Hughes *et al.*, 1981; Stuiver *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). Denton *et al.* (1991) suggested that surface elevations in the interior did not differ significantly from the present during the LGM, but that there was a considerable thickening of marginal domes in East Antarctica. 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). Anderson *et al.* (2002) reviewed the data used for constraining the extension of the LGM EAIS, and concluded that ice had at most advanced to positions on the middle or mid-outer continental shelf, not to the shelf break, and in some areas there had hardly been any LGM ice expansion compared to present ice configuration (Fig. 4).

Eastern Weddell Sea and Queen Maud Land – The Weddell Sea is a major discharge centre for ice, draining the West and East Antarctic ice sheets as well as the eastern side of the Antarctic Peninsula (Fig 1). The extent of the WAIS on the Weddell Sea continental shelf is poorly constrained, and attempts to radiocarbon date glacial marine sediments from the Weddell Sea shelf have been hampered by paucity of sediment cores and general absence of dateable carbonate materials in the cores (Anderson *et al.*, 2002). A number of piston cores taken from the Weddell Sea continental shelf penetrate tills below few metres of glaciomarine sediments (Anderson *et al.*, 1980; Fütterer & Melles 1990; Anderson *et al.*, 1991b). A till provenance study by Anderson *et al.* (1991b) suggested that the till on the shelf had been deposited by a glacier coming from East Antarctica. High-resolution seismic profiles record a mid-shelf grounding zone wedge that rests above a glacial unconformity that extends to the outer shelf (Elverhøi & Maisey, 1993). Kristoffersen *et al.* (2000) described two submarine ridge complexes on the shelf, parallel to the shelf edge, and reported a radiocarbon date of *c.* 19 ka BP constraining grounded ice extension to the mid-shelf area at the LGM.

Seismic data, sedimentological data and radiocarbon dates from the Weddell Sea have been interpreted as suggesting that a grounded ice sheet extended to the shelf break off Queen Maud Land prior to *c.* 21 ka BP (Elverhøi, 1981; Melles, 1991). Grobe *et al.* (1993) presented a sedimentary record from the eastern Weddell Sea, where they concluded that the Antarctic Ice Sheet had extended to the shelf edge during the LGM. Weber *et al.* (1994) studied submarine fan deposits in front of the continental slope in

the southeastern Weddell Sea. They suggested that prior to *c.* 20 ka BP, laminated sediments were deposited under conditions of very high sedimentation rates (250 cm ka⁻¹) when grounded ice sheet was in the proximity at the shelf edge. Anderson & Andrews (1999) concluded that tills on the continental shelf and extending towards the shelf break in the Weddell Sea pre-dated MIS 2, and recorded a glaciation older than the range of radiocarbon dating. Bentley & Anderson (1998), Anderson & Andrews (1999) and Anderson *et al.* (2002) suggested a major deglaciation event predating the LGM, occurring some time before 26-25 ka BP, and implied that ice had not extended towards the continental shelf after that (Fig. 4).

Jonsson (1988) used observations of glacial striae to reconstruct ice thickness at LGM in northern Vestfjella (Fig. 1), and concluded that the ice here had been 360 m thicker than today, but that the highest nunataks had remained 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 (Fig. 1) it was less than, 200 m thicker.

Lützow-Holm Bay - Ice-free areas fringing Lützow-Holm Bay (Fig. 1) occur on a number of islands and headlands, the largest being 61 km². 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; Igarashi *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 ka 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 (42-33 ka BP) fossil shells in raised beach deposits. Similarly, Maemoku *et al.* (1997) also described old (46-32 ka BP) fossil shells from Lützow-Holm Bay in undisturbed raised beach deposits, and interpreted that to indicate that LGM glaciers had not overridden the area. The position of the ice-margin during the LGM is unknown (Igarashi *et al.*, 1995, Maemoku *et al.*, 1997). While offshore troughs are identified and interpreted to reflect erosion by expanded systems, no offshore marine geological or geophysical studies have been carried out (Anderson *et al.*, 2002). Raised beaches of Holocene age, 7-2 ka 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 EAIS might have covered the southern part of Lützow-Holm Bay during LGM.

Taken together, the evidence can be interpreted to suggest that the EAIS did not expand significantly during the LGM in Eastern Weddell Sea and Dronning Maud Land (Anderson *et al.*, 2002; Fig. 4).

Mac Robertson Land, Prydz Bay and Larseman Hills – The geomorphology of the shelf off Mac Robertson Land shows signs of glacial erosion during ice sheet advance, and glaciers retreating across the shelf (Harris & O'Brien (1996) could have formed several moraine-like features. Harris & O'Brien (1998) place the grounding line of the EAIS on the outer shelf, but due to lacking dating control it remains unresolved if it was there during LGM or earlier. An uncorrected bulk-organic radiocarbon date of marine mud from the mid-outer shelf, from sediments interpreted to have been deposited below a floating ice shelf (Harris & O'Brien, 1998), gave an age of 17,1 ka BP.

It has been suggested that the small (50 km²) 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, morainic ridges and glacial striae only sparsely occur (Burgess *et al.*, 1994). Domack *et al.* (1998) concluded that the most extensive glaciation of the Prydz Bay area pre-dated the Northern Hemisphere LGM. They proposed that the last maximum extension of ice there, during MIS 2, reflected an advance of outlet systems and subsystems, reaching the mid-shelf areas, and that it was not a shelf-wide glaciation or wholesale advance of the EAIS. They described uninterrupted sedimentation on the shelf for the past 30 ka.

Vestfold Hills - During LGM the whole 400 km² of the Vestfold Hills oasis (Fig. 1) 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). Zwartz *et al.* (1998) estimated that the LGM ice margin had stood on the continental shelf some 30-40 km outside present margins in the Vestfold Hills oasis, and that it had thinned by 600-700 m at the present coastline since the LGM.

Bunger Hills - The Bunger Hills (Fig. 1) form the most extensive oasis of deglaciated hills and marine inlets in East Antarctica, with a total size of 952 km² (Wisniewski, 1983). There are signs of extensive glaciation such as 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 was evidence that at some stage in the past the Antarctic ice sheet completely submerged most of the oasis, and that ice flow from southeast to northwest was independent of the local topography. Augustinus *et al.* (1997) proposed that this extensive glaciation may have pre-dated the LGM. Limited glacial erosion, thick glacial deposits in the north-western part of the hills and the relatively low altitude of



Figure 8. The Ross Sea Drift in Explorers Cove, in the mouth of Tylor Valley, Southern Victoria Land, containing kenyte erratics (Photograph: Ó. Ingólfsson, 1994).

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. However, Melles *et al.* (1997) found till in 18 sediment cores retrieved from different basins within the oasis. They concluded that probably all the oasis was inundated by glaciers during the LGM.

The Windmill Islands - The Windmill Islands (Fig. 1) at the Budd Coast in East Antarctica, show evidence of an extensive late Pleistocene ice cover including 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 of 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. Marine geological surveys of the shelf areas close to the Windmill Islands reveal that the inner shelf is deeply eroded, reflecting glacial scouring, which Harris *et al.* (1997) suggested supported Goodwin's reconstruction of an expanded LGM ice. Recent modelling



Figure 9. Glacial trimline marked by the upper limit of the Ross Sea Drift, at c. 350 m a.s.l., on Hjorth Hill, Taylor Valley, Southern Victoria Land (Photograph: Ó. Ingólfsson 1994).

by Goodwin & Zweck (2000), suggests that the LGM ice may have been up to 1000 m thick and extended 40 km out on the continental shelf.

Wilkes Land, Oates Land and the Pennell Shelf – Anderson *et al.* (1980) suggested that the LGM EAIS had grounded on the outer shelf off Wilkes Land (Fig. 1), and Goodwin & Zweck (2000) suggested that LGM thickness of ice along the margin between Wilkes Land and Oates Land had been 1000 m. The Pennell shelf (Fig. 1) has large glacial troughs extending across the shelf, and sediment cores retrieved from the area penetrate a blanketing diamicton interpreted by Anderson *et al.* (2002) as till. They concluded that LGM ice was grounded at least to the middle shelf (Fig. 4). A number of radiocarbon dates constrain the ice expansion onto the shelf to have occurred prior to *c.* 14.3 ka BP.

6.4 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 in space and time (*cf.* Stuiver *et al.*, 1981; Denton *et al.*, 1989, 1991; Denton & Hughes, 2000). Reconstructions of LGM ice flowlines for the drainage of the East- and WAISs to the Ross Sea are conflicting (Drewry, 1979; Denton *et al.*, 1989; Clapperton & Sugden, 1990; Kellogg *et al.*, 1996; Denton & Hughes, 2000), 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 (Fig. 8), show that during the LGM the WAIS thickened and grounded in the Ross Sea and McMurdo Sound. Lobes of ice were pushed onto coastal southern Victoria Land, leaving signs in form of trimlines (Fig. 9) and damming the Dry Valleys (Stuiver *et al.*, 1981; Denton *et al.*, 1989, 1991). Erratics on islands in south-western Ross Sea, at elevations up to 320 m a.s.l., and moraines of the Ross Sea Drift occurring at 350 m a.s.l. in the Transantarctic Mountains indicate ice sheet elevation in the Ross Sea during the LGM (Hall & Denton, 1999). Further north along the Victoria Land coast, major outlet glaciers, for example in Terra Nova Bay, drained the EAIS and coalesced with the marine-based ice in the Ross Sea (Denton *et al.*, 1989, Orombelli *et al.*, 1991). Although reconstructions of ice extent in the Ross Sea during the LGM agree that ice was considerably expanded compared to today (*cf.* 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; Conway *et al.*, 1999; Denton & Hughes, 2000), there remains uncertainty about the maximum position of the grounded ice. Licht *et al.* (1996, 1999) recognised tills in the western Ross Sea, but only in areas south of Coulman Island (Fig. 1). Kellogg *et al.* (1996) suggested that grounded LGM ice had extended to the shelf

break off Cape Adare, while Denton *et al.* (1989), Anderson *et al.* (1992), Baroni & Orombelli (1994a), Shipp & Anderson (1994), Licht *et al.* (1996), Conway *et al.* (1999) and Shipp *et al.* (1999) place the LGM-grounding line along the northern Victoria Land coast in the vicinity of Coulman Island.

Although there are more than 450 radiocarbon dates reported in the literature that relate to the glacial history of the Ross Sea area (Denton & Hughes, 2000), the dating of the LGM-grounding line is not well constrained. Series of dates give maximum and minimum ages for the LGM position on the shelf: Orombelli *et al.* (1991), Licht *et al.* (1996) and Denton & Hughes (2000) constrained its age by maximum ages of >24 ka BP, where as a number of cores from sediments inside the LGM grounding line provide minimum ages in the range *c.* 20-11 ka BP (Anderson *et al.*, 1992; Licht *et al.*, 1996; Domack *et al.*, 1999a). Anderson *et al.* (1992) dated it off Northern Victoria Land to 17,3 ka BP and Licht *et al.* (1996) dated it to *c.* 20 ka BP. Penguin rookeries at Cape Adare and Cape Hallet were probably occupied between >35 ka BP and 17,3 ka BP (Baroni & Orombelli, 1994a). Since they can exist only where there is access to open water in summer, they show that the coastal northwestern Ross Sea was free of glacial ice during the LGM.

Marine geological studies in the central and eastern Ross Sea have revealed compelling evidences for LGM ice extending across the shelf towards the shelf edge: Alonso *et al.* (1992) and Shipp *et al.* (1999) reported the existence of a young unconformity in the stratigraphy, covered by till (Domack *et al.*, 1999a; Shipp *et al.*, 1999), indicating a major glacial expansion on the shelf. Swath bathymetry and side-scan sonar records show ice-contact features such as mega-scale glacial lineations and grounding zone sedimentary wedges, and the combined sedimentological and geophysical data suggests that LGM ice grounded at the shelf break in the Eastern Ross Sea (Shipp *et al.*, 1999; Anderson *et al.*, 2002). The dating control for the LGM expansion is problematic: Domack *et al.* (1999a) suggested that the LGM advance was marked by a large hiatus in radiocarbon ages between *c.* 25.5 and 19.5 ka BP (^{14}C dates on bulk-organic carbon, corrected by surface age of 3200 yrs), whereas unpublished dates from K. Licht (referred in Anderson *et al.* 2002) indicate that the ice may have been at its LGM position as late as 13.7 ka BP.

6.5 The Amundsen Sea

Large glacial troughs occur on the Amundsen Sea shelf, offshore drainage outlets of Marie Bird Land, and indicate evidence that ice sheets have expanded across the shelf (Anderson & Shipp, 2001). Drumlins and mega-scale glacial lineations also indicate a glacial overriding (Anderson *et al.*, 2002). The location of LGM grounding lines is not known, and there is no conclusive evidence of till on the outer shelf. The age control for ice retreat from the shelf areas and onset of glaciomarine sedimentation is



Figure 10. In situ shells (*Laternula elliptica*) at The Naze, northern James Ross Islands. They were ^{14}C -dated to 7.4 ka BP, and provide constraining minimum ages for the deglaciation as well as the age of the marine limit at c. 30 m a.s.l. (Photograph: Ó. Ingólfsson 1993).

in the form of constraining minimum ages giving ^{14}C ages of 12.9-11.5 ka BP (Anderson *et al.* 2002).

6.6 Summary of LGM ice-extent and its age around Antarctica

Most studies suggest that the LGM ice was considerably thicker and more extensive than present during the LGM, with ice extending off the present shore around most of Antarctica (Fig. 4). Some minor coastal areas may have remained wholly or partly ice-free. LGM-ice extended onto the mid-to-outer shelf areas off western Antarctica, Antarctic Peninsula and in the Ross Sea. In East Antarctica, the LGM ice reached its maximum extent at mid-shelf positions in some areas, whereas in other areas it only expanded less than that. The age of the last circum-Antarctic maximum extent of ice is poorly constrained by radiocarbon dates that broadly range between c. 24–14 ka BP, and there is data suggesting that there might have been an earlier (pre-MIS 2) glacial event of considerably more extensive glaciation onto the continental shelf areas.

7. Deglaciation and Holocene environmental development around Antarctica

7.1 The Antarctic Peninsula region

Deglaciation of the shelf and near-shore areas - There is a striking lack of radiocarbon dates constraining the deglaciation from the Antarctic Peninsula shelf area. This is probably partly a result of the lack of datable material in transitional glaciomarine sediments deposited below floating ice shelves in proximity to the grounding lines of the retreating glaciers. The oldest radiocarbon dates (Fig. 5), constraining ice retreat from the outer and middle shelf

areas west of the Antarctic Peninsula come from the Bransfield Strait (Banfield & Anderson, 1995) and the shelf areas west of Anvers Island (Pope & Anderson, 1992; Pudsey *et al.*, 1994). These indicate ice retreat prior to 14–13 ka BP and 12–11 ka BP, respectively. 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), glaciomarine sedimentation in the central part of the Gerlache Strait began after 8 ka BP, 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 c. 8 ka BP. The inner shelf areas around Anvers Island and in Marguerite Bay, as well as George VI Sound, were ice-free at 7–6 ka BP (Clapperton & Sugden, 1982; Kennedy & Anderson, 1989; Harden *et al.*, 1992; Pope & Anderson, 1992; Pudsey *et al.*, 1994; Hjort *et al.*, 2001). Domack *et al.* (2001a) recently published a Late Wisconsinan-Holocene chronology for environmental changes in the Palmer Deep (Fig. 5) that showed that the inner shelf areas there might have been deglaciated as early as c. 11 ka BP.

The low resolution in the deglaciation chronologies from most of the shelf areas makes it impossible to recognize more than the general pattern, that the outer-and middle shelf areas deglaciated between 14–8 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) recognised a rapid deglacial episode in the Palmer Deep record, characterised by high primary production and iceberg rafting, between c. 11–10 ka BP.



Figure 11. The large penguin rookery in Hope Bay. Penguins have occupied this site since *c.* 5.5 ka BP. The earliest Holocene penguin occupation in the Antarctic Peninsula-South Shetland Islands region has been dated to *c.* 6.5 ka BP in Marguerite Bay (Photograph: Ó. Ingólfsson 1987).

This deglacial event occurred concurrently with the Northern Hemisphere Younger Dryas Stadial. In the Palmer Deep record, it was followed by a climate cooling between *c.* 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 shelf areas were controlled by regional warming. Hence the ice retreat was probably controlled by the global sea-level rise that resulted from meltwater input from the Northern Hemisphere glaciers (*cf.* Hollin, 1962; Stuiver *et al.*, 1981)

Deglaciation of currently ice-free Antarctic Peninsula coastal areas - The deglaciation of coastal areas that are currently ice-free is constrained by minimum ages obtained from radiocarbon dating of mollusc shells from raised marine deposits, peat deposits in moss-banks on the islands off the Antarctic Peninsula, organics (bulk organics, microbial mats, aquatic mosses and algal flakes) from lake sediments and remains from abandoned penguin colonies. The oldest radiocarbon dates, on fossil mollusks from raised marine deposits and lake sediments, give a minimum age for initial deglaciation on King George Island as 9-8 ka BP (Mäusbacher, 1991; Sugden & John, 1973). A well-dated lithostratigraphical record from northern James Ross Island (Fig. 5), where glaciomarine and sub-littoral deposits overlie till in coastal sections (Fig. 10), constrains deglaciation there as to before *c.* 7.4 ka BP (Hjort *et al.*, 1997). According to Zale (1994a), Hope Bay was

deglaciated prior to *ca.* 6.3 ka BP. In the southern part of the peninsula, radiocarbon dated shell fragments from ice-shelf moraines at Two Step Cliffs on Alexander Island provide minimum ages for deglaciation and ice-shelf retreat in George VI Sound of 6.5-5.7 ka BP (Clapperton & Sugden, 1982; Hjort *et al.*, 2001).

A number of studies, exploring primarily lake-sediment archives and moss-bank peats, suggest that once the glaciers were inside the present coastline, glacial retreat and ice-disintegration was slow (Barsch & Mäusbacher, 1986a; Mäusbacher *et al.*, 1989; Zale & Karlén, 1989; Mäusbacher, 1991; Ingólfsson *et al.*, 1992; Björck *et al.*, 1993, 1996a, 1996b; Lopez-Martinez *et al.*, 1996; Hjort *et al.*, 1997): On King George Island, glaciers were at or within their present limits by *c.* 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 were deglaciated as late as 5-3 ka BP (Björck *et al.*, 1996a). The 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 & Orombelli, 1994). The timing of penguin occupation at coastal sites is therefore a good proxy for minimum age of



Figure. 12. Ice-pushed ridges where the George VI Sound ice shelf transgresses Alexander Island. Before c. 6.5 ka BP the George VI Sound was free of shelf ice (Photograph: Ó. Ingólfsson 2000).

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 (Fig. 11). 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 & Mäusbacher, 1986b).

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

Mid-Holocene glacial re-advances - There are indications from a number of sites for mid-Holocene glacial expansions: Mäusbacher (1991) found evidence of increased glacial activity between 5-4 ka BP on King George Island, supporting Sugden & John (1973) who had suggested glacial expansion here after 6 ka BP. Hansom & Flint (1989) recorded a glacial readvance on Brabant Island some time after 5.3 ka BP, and the post 5.7 ka BP expansion of the George VI Sound ice shelf is well documented (Fig. 12) (Sugden & Clapperton, 1980, 1981; Clapperton & 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 evidence to have culminated in an at least 7 km ice-advance around 4.6 ka BP (Hjort *et al.*, 1997). Zale (1994) 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 *c.* 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 Shevenell *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 - Palaeoclimatic records, with high-resolution and reliable chronologies, are scarce from the Antarctic Peninsula region (Fenton, 1980, 1982; Barsch & Mäusbacher, 1986; Zale & Karlén, 1989; Mäusbacher *et al.*, 1989; Mäusbacher, 1991; Schmidt *et al.*, 1990; Björck *et al.*, 1991b, 1991c, 1993, 1996a, 1996b; Wasell & Håkansson, 1992; Wasell, 1993; Zale, 1993, 1994; Håkansson *et al.*, 1995; Yang & Hartwood, 1997; Jones *et al.*, 2000; Domack *et al.*, 2001a). Björck *et al.* (1996a) presented a palaeoclimatic synthesis for the middle-late

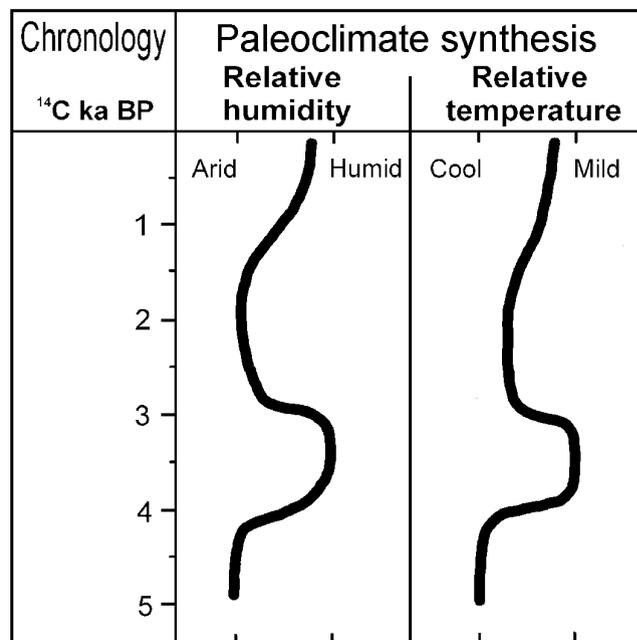


Figure 13. The synthesis of Björck *et al.* (1996a) for mid-late Holocene climate development in the Antarctic Peninsula region.

Holocene development in the Antarctic Peninsula region, based multivariate analysis on a large and complex data set containing numerous stratigraphical variables in lake sediments and moss-bank peats (Fig. 13). Their reconstruction indicates that the climate fluctuated from relatively mild and humid conditions prior to *c.* 5 ka BP to more cold and arid conditions later. Around 4.2 ka BP a gradual warming occurred, coupled with increasing humidity. These mild and humid conditions reached an optimum around 3 ka BP, where after a distinct and rapid climatic deterioration occurred.

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 this Björck *et al.* (1996a) suggested might indicate that the primary factor controlling the climatic variations is the strength of the high-pressure cell over the Antarctic ice sheet. A paleoenvironmental record from lake sediment cores from the maritime Signy Island, in the South Orkney Islands archipelago, is interpreted to indicate a Holocene climate optimum between *c.* 3.8-1.3 ka BP (Jones *et al.*, 2000). This demonstrates that the climate optimum persisted for a longer period at Signy Island than at sites more proximal to the Antarctic continent.

The marine record from Lallemand Fjord (Fig. 5) recognises 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.* 8-3 ka BP,

Environmental changes in the Antarctic Peninsula region since LGM

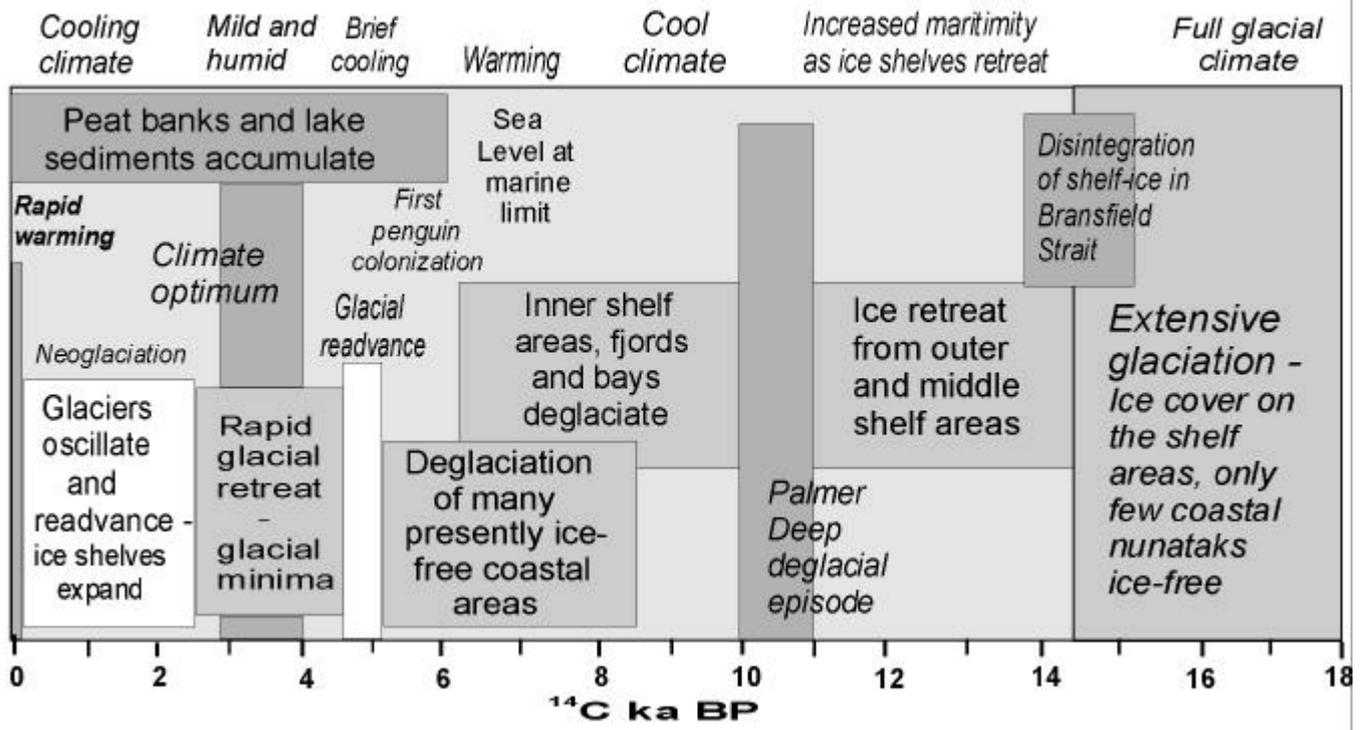


Figure 14. Tentative synthesis for environmental changes in the Antarctic Peninsula since LGM. Modified after Hjort *et al.* (in press).

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 palaeoclimatic 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: climate cooling and increased ice volumes - The palaeoclimatic synthesis of Björck *et al.* (1996a) recognised that after *c.* 3 ka BP the climate became characterised by cold and dry conditions, which persisted until *c.* 1.5 ka BP. Thereafter the climate was somewhat warmer and more humid, but still cold compared to the situation during the climatic optimum. Several investigations suggest that glaciers have oscillated and expanded in the Antarctic Peninsula region during the past *c.* 2.5 ka (John & Sugden, 1971; John, 1972; Sugden & John, 1973; Curl, 1980; Birkenmajer, 1981; Clapperton & Sugden, 1988; Zale & Karlén, 1989; Clapperton, 1990; López-Martínez *et al.*, 1996). For example, in the South Shetlands, Curl (1980), Birkenmajer (1981), Clapperton & 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 advances in the Northern Hemisphere. Lichenometric dating, using *Rhizocarpon geographicum* thalli, date the advances to *c.* A.D. 1240, 1720 and 1780-

1822 (Curl, 1980; Birkenmajer, 1981). Whalebone found on a moraine at the front of Rotch Dome on Livingston Island dates the advance here to after *c.* 0.3 ka BP, after *c.* AD 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 recognise a distinct neoglaciation cooling as well as a LIA cold pulse. The marine record from Lallemand Fjord (Shevenell *et al.*, 1996) suggests decreased productivity in the fjord after 2.7 ka BP due to cooling. Diatom abundances reflect more extensive and seasonally persistent sea ice after *c.* 2.7 ka BP (Shevenell *et al.*, 1996) and after *c.* 0.4 ka BP an ice shelf advance into the fjord is recorded. 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.* 2.5 ka BP marked the termination of the Holocene climate optimum as reflected in the sediment core record. Domack *et al.* (2001a) define a Neoglaciation cooling in the Palmer Deep record, starting at *c.* 3 ka BP and lasting until about 100 years ago. The Neoglaciation conditions are characterised by greater concentrations of IRD, decreasing sediment accumulation and decreased

biological productivity. The timing of the Neoglacial onset, as reflected in the Palmer Deep record, agrees well with other marine (Smith *et al.*, 1999) and terrestrial (Björck *et al.*, 1996a; Ingólfsson *et al.*, 1998) evidence.

The last 100 years: a distinct warming - A number of palaeoclimatic 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 years (*cf.* review in Smith *et al.*, 1999). Air-temperature records from the Antarctic Peninsula demonstrate a distinct and dramatic warming trend for the past 50 years (King, 1994; Stark, 1994; Smith *et al.*, 1996). Polar ecosystem research and palaeoecological records indicate ecological transitions that have occurred in response to this climate change (Fraser *et al.*, 1992; Emslie, 1995; Trivelpiece & Frasier, 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, 1980's (Rott *et al.*, 1996; Vaughan & Doake, 1996; Cooper, 1997; Skvarca *et al.*, 1999; Scambos, 2000). The retreat appears to have been caused by 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 *c.* 10 ka (Hjort *et al.*, 2001; Pudsey & Evand, 2001). Domack *et al.* (2000b) and Pudsey & Evans (2001) suggested that the expansion of the Larsen Ice Shelf and the Prince Gustav Channel Ice Shelf, respectively, was a neoglacial phenomena, related to enhanced cooling and more persistent sea ice in the last 2,5 ka. The environmental development in the Antarctic Peninsula region since LGM is summarized in Fig. 14.

7.2 Deglaciation and Holocene development in East Antarctica

Queen Maud Land - There are only very few constraints on the ice-volume decrease in Queen Mud Land. Lintinen & Nenonen (1997) reported radiocarbon dates of stomach oil deposits from nesting sites of snow petrels (*Pagodroma nivea*) giving 7.4 BP as the minimum age for deglaciation of the southern Vestfjella nunataks and the lower altitudes of Heimefrontfjella (Fig. 1). There has been continuous avian occupation since then, but details of any glacial and climate fluctuations are lacking.

Schirmacher and Untersee Oases - The Schirmacher Oasis (Fig. 1) is a small (34 km²) ice-free area located *c.* 100 km south of the Lazarev Sea. The last glacial submergence of the Schirmacher Oasis is reflected in a sparse coverage of morainic material, fractured rocks, striated surfaces and *roches moutonnées* in many places in the oasis (Richter & Bormann, 1995). Radiocarbon dating of lake sediments suggests that deglaciation of the Schirmacher Oases began

before 3,5 ka BP (M. Schwab, personal communication, 1998; cited in Ingólfsson *et al.*, 1998). Hence, deglaciation here may have occurred significantly later than in Untersee Oasis to the south (see below), and also later than on the Lazarev Sea shelf to the north, where it was under way by 9,5 ka BP (Gingele *et al.*, 1997).

The Untersee Oasis (Fig. 1) forms the eastern rim of the ice free 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 pre-dates the LGM (Hiller *et al.*, 1988). This is indicated by Late Wisconsinan ¹⁴C-ages on snow petrel stomach oil deposits from high-altitude sites in the Untersee Oasis.

No indications are known from Untersee nor Schirmacher Oases for significant Holocene re-advances 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 re-advanced some time between 82 ka BP (Gingele *et al.*, 1997).

Mac Robertson Land - Evidence from marine cores from the Mac Robertson Land shelf suggest that deglaciation of the outer shelf occurred some time prior to 10,8-10 ka BP, and that afterwards that open-marine deposition prevailed (Harris & O'Brien, 1998; Taylor & McMinn, 2001). Harris *et al.* (1996) suggest open marine conditions on the mid-shelf by *c.* 7 ka BP and within the inner shelf basins by *c.* 5 ka BP. The diatom record suggests a succession of mid-late Holocene high-productivity events, associated with a climate optimum occurring broadly within the interval 5,8-0,4 ka BP. Sedwick *et al.* (2001) suggested that grounded ice on the outer shelf started to retreat before 9 ka BP, and that the front of a floating ice shelf had retreated to the inner shelf by *c.* 5 ka BP. They recognised six episodes of increased accumulation of biogenic material since 10,8 ka BP, which they interpreted as the result of enhanced diatom production over the outer shelf, possibly related to climatic warm periods.

Larsemann Hills and Prydz Bay - A number of sediment cores have been retrieved from lakes in the Larsemann Hills. There are serious problems with radiocarbon dates from the sediments, expressed as reversed stratigraphical age successions (Burgess *et al.*, 1994). While most dated samples give mid-late Holocene ages for the deglaciation, a single date gave 9,4 ka BP and another 25 ka BP. The Amery Ice Shelf (Fig. 1), the largest in East Antarctica, has been present throughout the Holocene since retreat from the mid-shelf LGM terminal moraines in Prydz Bay (O'Brien & Harris, 1996; Domack *et al.*, 1998; Harris *et al.*, 1998; O'Brien *et al.*, 1999). The Holocene grounding line retreat is constrained by a number of radiocarbon dates to having

occurred successively since *c.* 11 ka BP (O'Brien & Harris, 1996; Domack *et al.*, 1999b).

Vestfold Hills - The last deglaciation of Vestfold Hills has been determined by radiocarbon 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; Bird *et al.*, 1991; Bronge, 1992). A prominent feature of the Vestfold Hills is its numerous lakes. There are about 300 lakes, with fresh to saline and hypersaline water. Many were formerly marine inlets and became isolated by isostatic uplift following the glacial retreat. The oldest radiocarbon 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 8,6-8,4 ka BP (Pickard & Seppelt, 1984; Fitzsimons & Domack, 1993; Bird *et al.*, 1991; Fulford-Smith & Sikes, 1996). According to Adamson & Pickard (1986) and Pickard *et al.* (1986), ice retreat thereafter was slow or stepwise, averaging 1-2 m/yr, with 20% of the land area exposed by 8 ka BP, 50% by 5 ka BP, and the ice margin reaching its present position in the last 1000 years. Relative sea level stood at the marine limit in Vestfold Hills, at *c.* 9 m a.s.l., at *c.* 6 ka BP (Zwartz *et al.*, 1998). Fitzsimons & Domack (1993) and Fitzsimons & Colhoun (1995) maintain that the southern part of the Vestfold Hills deglaciated prior to 8,6 ka BP.

Domack *et al.* (1991b) found evidence on the shelf for a middle Holocene readvance of floating ice tongues some time within the interval 7,3-3,8 ka BP. There is no dated evidence for this from the Vestfold Hills, but Adamson & Pickard (1986) suggested that a mid-Holocene glacial advance may be recognised in the form of moraine ridges in the area. A late Holocene ice advance, termed the Chelnok Glaciation, is poorly dated but probably occurred after 2.5 ka BP and before 1 ka BP (Pickard *et al.*, 1984; Adamson and Pickard, 1986; Zhang, 1992; McMinn *et al.*, 2001). Fitzsimons and Colhoun (1995) found evidence for minor (\leq 500 m) late Holocene (post 0.7 ka BP) ice-marginal fluctuations in the form of a discontinuous series of ice-cored moraine ridges; they may correlate with the Little Ice Age. Pickard *et al.* (1984, 1986) concluded that the 'post-glacial' climate of the Vestfold Hills had been very stable and similar to that of today. Zhang (1992), however, on the basis of marine fossil assemblages and geomorphological criteria, proposed that a climatic optimum had occurred here between 6,2 and 3,7 ka BP. In the geomorphological record, Pickard (1982) found evidence that the predominant wind direction had been stable for the past 4 ka.

Björck *et al.* (1996a) re-interpreted the lake sediment evidence of Pickard *et al.* (1986) in terms of palaeoclimate development for the Vestfold Hills. They suggested that the deglaciation prior to 4.7 ka BP occurred under an arid and cold environment (low lake levels, high salinity), followed by a relatively warm and humid climate between 4,7-3,0 ka

BP which caused intense melting of dead ice in the vicinity of the lakes. Massive input of fresh water into the basins caused salinity to fall and lake levels to rise. After 3 ka BP the climate again turned arid and cold, with decreased fresh water input and lowered lake levels. Björck *et al.* (1996a) thus conclude that a late Holocene (4.7 - 3.0 BP) climate optimum occurred in the Vestfold Hills. Roberts & McMinn (1996, 1999) used transfer functions for the reconstruction of past lakewater salinity from fossil diatom assemblages. They found that since 5,2 BP, one lake studied had undergone cycles of varying salinity. The data indicate a long period with relatively low salinity, possibly indicating warmer and wetter conditions in the time interval *c.* 4.2 - 2.2 ka BP. McMinn *et al.* (2001) studied marine sediment cores from the Vestfold Hills area. They recognised a mid-late Holocene warm period ($>$ 3.6 - 2.5 ka BP) and a significant cooling and increase in sea-ice cover after 2.5 ka BP. They correlate the cooling with the Chelnok Glaciation event, and suggest it can also be identified in lake cores (Roberts & McMinn, 1997, 1999) and marine cores (McMinn, 2000; McMinn *et al.*, 2001), as well as in the geomorphological record (Pickard *et al.*, 1984; Adamson & Pickard, 1986).

Bunger Hills - Numerous radiocarbon 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 - 8 ka BP (Bolshiyarov *et al.*, 1990, 1991; Melles *et al.*, 1994; Melles *et al.*, 1997; Kulbe *et al.*, 2001), and radiocarbon dates of stomach oil deposits at nest sites of snow petrels show occupation as early as 9.5 ka BP (Bolshiyarov *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. Kulbe *et al.* (2001) recognised the period after initial deglaciation, between *c.* 9 - 7 ka BP, to have been relatively warm. The southern part of the Bunger Hill Oasis deglaciated at *c.* 9 ka BP, and synchronous accumulation of subaquatic sediments in basins and stomach oil deposits on ice-free hills precludes the proximity of any large ice masses since *c.* 8.5 ka BP (Kulbe *et al.*, 2001). Between 9 - 7.7 ka BP the sea flooded all the major inlets in Bunger Hills (Colhoun & Adamson, 1991, Melles *et al.*, 1997; Kulbe *et al.*, 2001). The breeding colonies of snow petrels expanded continuously, following ice retreat and the down-wasting of dead ice (Verkulich & Hiller, 1994), with the most intense phases of colonisation between 6.7-4.7 ka BP. Before 5.6 ka BP when the sea stood at the marine limit at between 9-7 m, glaciers were at or behind their present margins (Colhoun & Adamson, 1992a, Colhoun & Adamson, 1992b).

Figure 15. Summary of glacial and environmental changes since LGM in Eastern Antarctica. Modified after Ingólfsson *et al.* (1998).

¹⁴ C ka BP	Glacial and environmental changes		
	Vestfold Hills	Bunger Hills	Windmill Islands
1	Glacial readvance	Glacial readvance (Younger Eddisto Moraines)	More extensive sea ice
2	Chelnok glacial readvance		Glacial readvance
3		Climate optimum - open marine conditions -	Warmer than present?
4	Climate Optimum		
5	Glacial readvance	Glacial readvance (Older Eddisto Moraines)	
6	Down-wasting of dead ice		Northern Islands deglaciates
7	Slow deglaciation	Melting of dead ice; expanding avian habitats	
8	Onset of lake sedimentation		
9	Ice retreat on land	Deglaciation of southern part of the Oasis	Southern Islands deglaciates Shelf areas deglaciates
10	Deglaciation of inner shelf areas	Shelf areas deglaciates	
11	Complete glacial coverage at LGM	Complete glacial coverage at LGM(?)	Complete glacial coverage at LGM

Evidence on the mid-late Holocene glacial and climatic evolution of Bunger Hills is, however, somewhat controversial. Bolshiyarov *et al.* (1991) suggested that the area was re-glaciated several times during the Holocene, causing damming of tributary valleys with periodical 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 shifting 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 on 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 so-called Older Edisto Moraines, post-dating 6.2 ka BP (Colhoun & Adamson, 1992a).

Verkulich & Melles (1992), Melles *et al.* (1997) and Kulbe *et al.* (2001) have studied sediment cores from freshwater lakes and from marine basins in the Bunger Hills Oasis. Melles *et al.* (1997) report high and stepwise increasing biogenic production between 4.7 to 2.0 ka BP,

taken to indicate increased temperatures and correlated with the Antarctic Peninsula climate optimum at about the same time. Melles *et al.* (1997) also conclude that the glacial advance forming the Older Edisto Moraines pre-dated the 4.7 ka BP warming. This constrains the Edisto advance to between 6.2-4.7 ka BP, which is coincident with the cooling and glacial readvance in the Antarctic Peninsula area (see above). Roberts *et al.* (2000) recognised a period between 4-2 ka BP of decreased salinity in lake sediments cores from a Bunger Hill lake, which they thought signified more humid and warmer conditions. Rozycki (1961) and Colhoun & Adamson (1992a, 1992b) interpreted beach morphology to suggest 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. A recent study by Kulbe *et al.* (2001) suggest that there was a significant mid-late Holocene climatic variability in the Bunger Hills Oasis, with a climate optimum between c. 3.5-2.5 ka BP. An abrupt and dramatic cooling followed it within less than 200 years.

Finally, Colhoun & Adamson (1992a) described a glacial advance during the last few centuries, leading to the formation of the so-called 'Younger Edisto Moraines'. Melles *et al.* (1997) concluded that the 'Younger Edisto

Moraines' post-date 1,1 ka BP. Marine shell fragments collected from the moraines dated to 0,2 ka BP (Colhoun & Adamson, 1992a), suggesting a Little Ice Age glacial event.

Windmill Islands - Radiocarbon dating of bulk samples from basal lake sediments provide minimum estimates for the deglaciation of the Windmill Islands (Goodwin, 1993). These indicate that the southern part of the islands were deglaciated before 8 ka BP, while the northern islands were only deglaciated some time before 5,5 ka 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 2-1 ka 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 (Fig. 1) onto part of the Windmill Islands between 4-1 ka 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 evidence - The oldest deglaciation dates show open marine and lacustrine environments, as well as avian habitats, developing within inner shelf basins and coastal areas between 10 – 8 ka BP (Fig. 15). The deglaciation history of Vestfold Hills is controversial. One reconstruction specifies it as successive and slow and mainly post-dating 8 ka BP. In the other reconstruction, deglaciation is advanced by 8.6 ka BP. There are indications from the 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 ka 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 down-wasting of the dead ice may have occurred successively after 8.8 ka BP. There are indications of a mid-Holocene glacial re-advance; expansion of the Edisto Glacier occurred after 6.2 ka BP and before 4.7 ka BP. The Windmill Islands were successively deglaciated between 8 - 5 ka BP. There are indications from both Vestfold Hills and Bunger Hills of a climate warmer and wetter than the present in the interval c. 4.7 - 2.5 ka BP, and the Windmill Islands record may tentatively be interpreted as indicating the same for the 4-1 ka BP interval. The records from both Vestfold Hills and Bunger Hills strongly suggest Neoglacial cooling after c. 2.5 ka BP. The East Antarctic data also show minor fluctuations of the ice margins over the past few hundred

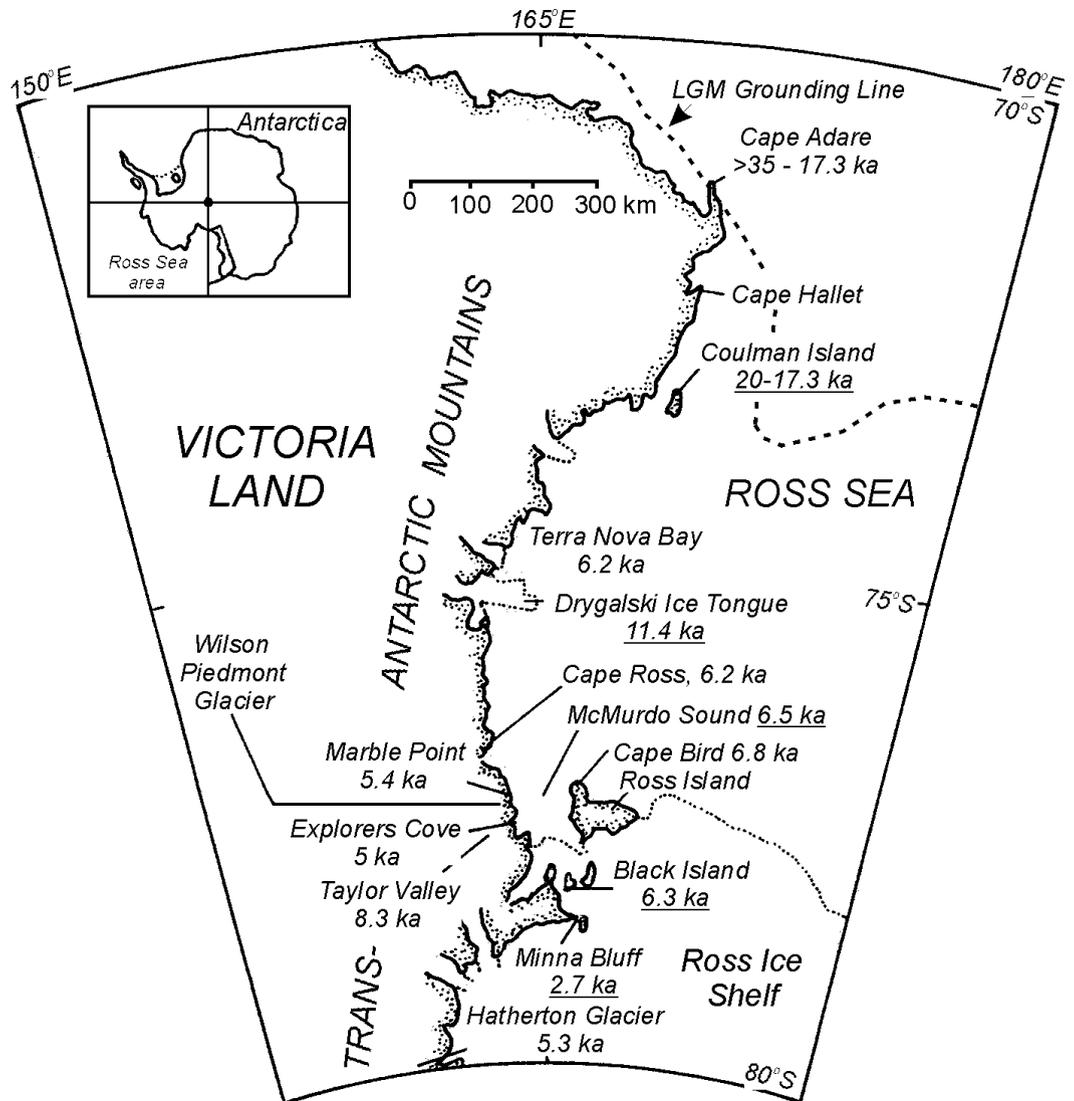
years (Goodwin, 1998), possibly correlative with the Little Ice Age advances in the Northern Hemisphere.

7.3 The deglaciation of Ross Sea and Victoria Land

Conway *et al.* (1999), Hall & Denton (2000a, 2002b) and Denton & Hughes (2000) have recently reviewed the retreat of ice in Ross Sea and along the Victoria Land coast. Several radiocarbon dates constrain the marine-based ice sheet retreat in the Ross Sea from the LGM-grounding line after 17 ka 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. 11,4 ka BP (Licht *et al.*, 1996). At the LGM, the Terra Nova Bay region was occupied by coalescing outlet glaciers, draining the EAIS and joining the marine based Ross Ice Sheet (Baroni & Orombelli, 1991; Orombelli *et al.*, 1991). Glacial drift deposits relating 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 11.4 ka 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 6.2 ka 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 to 5.8 ka BP.

Denton *et al.* (1991), Hall (1997) and Hall & Denton (2000a) date the presence of Ross Sea Ice Sheet in Explorers Cove, southern Victoria Land, damming lakes along the valley threshold, at 8.3 ka BP. Dates from penguin rookeries on Ross Island and in the McMurdo Sound area provide minimum ages for the deglaciation in the southern Ross Sea. Rookeries at Cape Bird were occupied by 6.8-5.7 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 6.3 ka BP (Kellogg *et al.*, 1990), which is in line with the penguin data. Radiocarbon 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. 6.3-6.1 ka 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 6.3 ka BP and provides one minimum estimate for the deglaciation of coastal southern Victoria Land (Conway *et al.*, 1999; Hall & Denton, 1999). It coincides well with the oldest radiocarbon date on a bivalve from a core in McMurdo Sound, which gives 6.5 ka BP as the minimum age of grounding line retreat from the Sound (Licht *et al.*, 1996). The marine limit becomes gradually younger southwards,

Figure 16. The Ross Sea-Victoria Land region, with minimum ages for deglaciation. Underlined ^{14}C ages relate to dated grounding line recession. Modified after Ingólfsson *et al.* (1998). Location of LGM grounding line after Conway *et al.* (1999) and Anderson *et al.* (2002).



which might indicate lagging of shelf ice recession behind that of the grounding line. The age of the marine limit is 6.2 ka BP in Terra Nova Bay, 5.4 ka BP at Marble Point/South Stream and 5 ka BP in the Explorer's 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.* 5.3 ka BP. Numerous radiocarbon 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 2.7 ka BP (Kellogg *et al.*, 1990). Conway *et al.* (1999) estimated the rate of grounding line retreat to have been *c.* 120 m y^{-1} for the past 7.5 ka. At present the Ross Shelf grounding line is located *c.* 900 km south of McMurdo Sound, still retreating. Conway *et al.* (1999) suggested that continued recession might lead to the

complete disintegration of the WAIS within the present interglacial.

Mid-late Holocene glacier oscillations - There is some information available as to mid- to late-Holocene glacier variations in Victoria Land. Raised beaches in the area between Explorers Cove and Marble Point (Fig. 16), ^{14}C dated to *c.* 5 ka BP, terminate against the Wilson Piedmont Glacier and presumably extend beneath the ice (Hall & Denton 2000b). This suggests that the Wilson Piedmont Glacier has expanded since 5 ka BP. 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 6.2 and 5.3 ka BP. A readvance across raised beaches took place some time after 5.3 ka BP (Baroni & Orombelli, 1994b). There was then a renewed withdrawal phase between *c.* 1 - 0.5 ka BP, correlated by Baroni (1994) and Baroni & Orombelli (1994b) with the Northern Hemisphere Medieval Warm

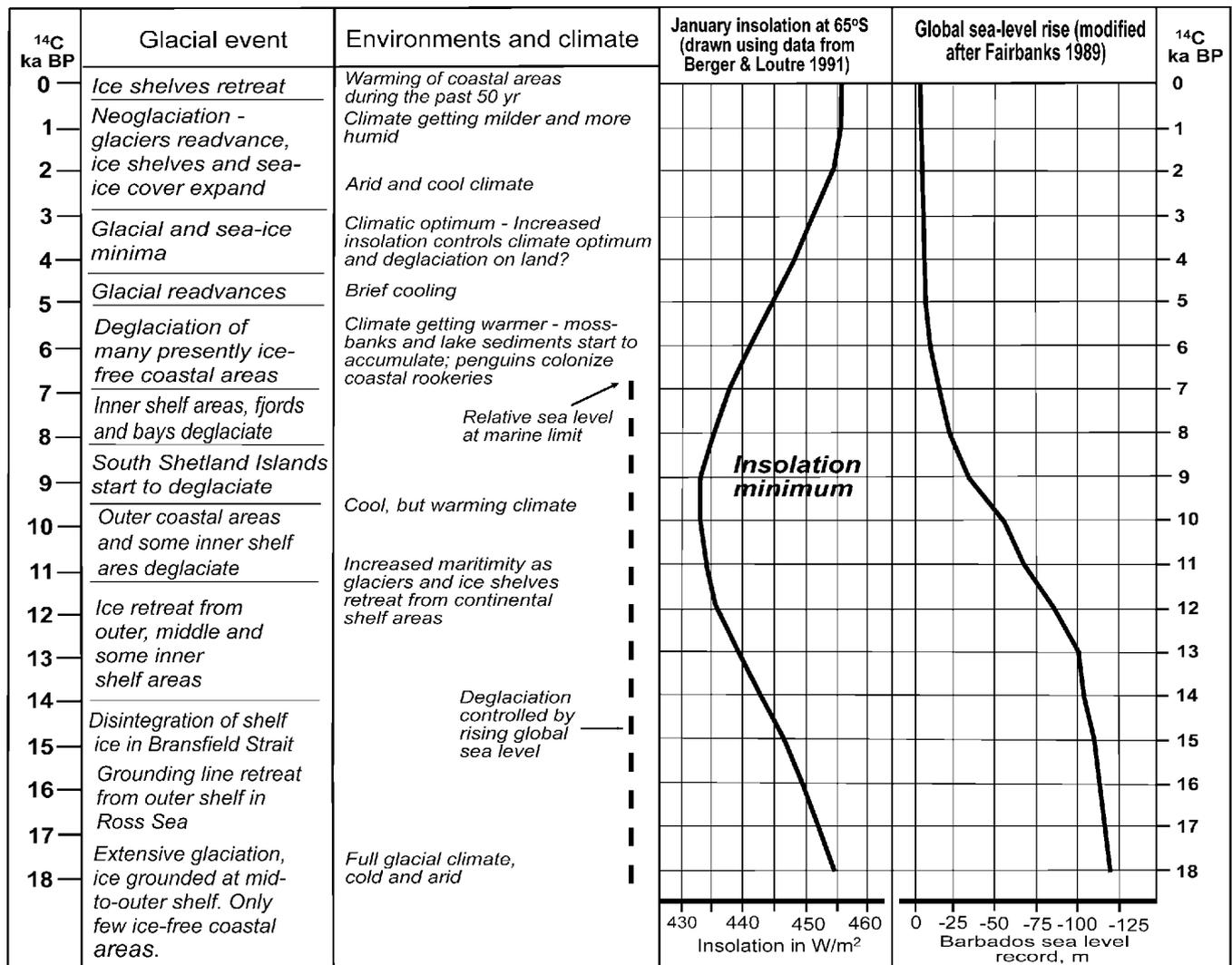


Figure 17. A broad synthesis for glacial and environmental changes in Antarctica since LGM.

Period (ca. 1000-1300 A.D.). Moraine ridges containing fossil marine shells then show a glacial readvance in Terra Nova Bay after 0.5 ka 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 A.D., when the English 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 at present some of these are 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 LGM (Hall & Denton, 2000a) and merged with it north of Explorers Cove. The Wilson Piedmont glacier

retreated inside the coast between 5.7 - 5 ka BP, but readvanced in late Holocene times, and at a number of places it advanced across raised beaches which date back to 5,4-5 ka BP (Nichols, 1968, Stuiver *et al.*, 1981, Denton *et al.*, 1989, Hall, 1997).

Climatic optimum in Victoria Land - 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). They document the continuous presence of Adélie penguins after *c.* 7 ka BP, but the greatest diffusion of rookeries occurred between 3.6 - 2.6 ka 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 2.6 ka 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 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 5.4 ka 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 5.4 ka BP (Stuiver *et al.*, 1981, Berkman, 1997). Cunningham *et al.* (1999) studied the palaeoceanographic changes in the western and west-central Ross Sea from >12 ka BP to the present. They recognised a distinct warm period during the mid-Holocene, *c.* 6-3 ka BP, and increasingly cooler late Holocene after 3 ka BP.

Studies on lake histories from the McMurdo Dry Valleys of Southern Victoria Land (Doran *et al.*, 1994; Lyons *et al.*, 1998) give different reconstructions of the late Holocene climate in southern Victoria Land. They recognise a relatively mild period in the interval *c.* 3 - 2 ka BP, followed by a cool, dry period between 2 - 1.2 ka BP and succeeded by a mild period from 1.2 - 0 ka BP. This discrepancy between the records is difficult to understand, but it might be explained by either a combination of microclimatic and topographical variables controlling hydrological, ecological and chemical balances in the lakes, or increased distality of the area to the retreating Ross Ice Shelf. Records of thinning of perennial ice thickness on lakes in southern Victoria Land have been used to substantiate warming and increased summer melting during the past few decades (Webster *et al.*, 1996; Wharton *et al.*, 1992). Meteorological data for the past *ca.* 50 years show that Antarctic surface air temperature trends towards cooling in the interior and warming in the coastal regions, and Doran *et al.* (2002) have described ecosystem response in areas affected by this cooling.

8. Summary and conclusions of the glacial history since the LGM

Detailed reconstruction of the glacial and climatic history of Antarctica since the LGM is hampered by the scarcity of available archives, low resolution in many datasets and chronological problems. Therefore, any synthesis (Fig. 17) must be regarded as a tentative description of the environmental development. However, the following broad pattern can be recognized:

- a) There is evidence of much more extensive ice cover in Antarctica prior to *c.* 20-14 ka BP. LGM glaciers from the WAIS and the Antarctic Peninsula generally grounded on the middle-outer shelf, at water depths of ≥ 400 m, and were probably fringed by ice shelves extending over the outer shelf and the shelf break. East Antarctic glaciers expanded to mid-shelf positions, and in

some areas there was little or no LGM glacial expansion.

- b) Ice retreat from the LGM positions was under way by 17-14 ka BP. The initial ice retreat was probably eustatically-controlled.
- c) Deglaciation occurred mainly during the time period >14-6 ka BP. Outer- and middle shelf areas deglaciated between 14-8 ka BP, while most inner shelf areas, fjords, bays and most currently ice-free coastal land areas deglaciated prior to 8-6 ka BP. This suggests that Antarctica could have contributed to global sea-level rise at least until mid-Holocene times (*cf.* Flemming *et al.*, 1998; Ingólfsson & Hjort, 1999). There are signs in the East Antarctic record (Bunger Hills) that early Holocene glaciation may have been at least partly driven by rising temperatures.
- d) There are no signs in Antarctica of a Younger Dryas glacial or cold climatic event.
- e) The transition from glacial to interglacial conditions in Antarctica, indicated 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.
- f) Mid-Holocene glacial- and ice-shelf readvances have been described from a number of sites.
- g) Terrestrial palaeoclimatic records, based on stratigraphical variables in lake- and moss-bank archives, as well as proliferation of penguin rookeries, suggest a circum-Antarctic climate optimum occurring broadly in the period 4.5-2.5 ka BP. It is best constrained in the stratigraphical records from the Antarctic Peninsula (4-3 ka BP), Bunger Hills in East Antarctica (3.5-2.5 ka BP) and Victoria Land (3.6-2.6 ka BP).
- h) Available data suggest that after *c.* 2.5 ka BP a distinct Neoglacial cooling occurred, and that many glaciers and ice-shelves expanded during the late Holocene. The Ross Ice Shelf probably continued retreating even after 2.5 ka BP and is still in a retreat mode.
- i) A rapid and significant warming has occurred in the Antarctic Peninsula region for the past 100 years, which might be causing ice-shelf instability, and coastal areas in the Ross Sea also show signs of warming. The interior East Antarctica is slightly cooling.

The available data suggest that Antarctic deglaciation and Holocene climate development lagged that of the Northern Hemisphere (Hjort *et al.*, 1998; Ingólfsson *et al.*, 1998; Bentley, 1999; Hall & Denton, 1999, 2000a). The late Holocene climate optimum, *c.* 4.5 - 2.5 ka BP is recognised as broadly synchronous environmental changes in the Antarctic Peninsula, East Antarctica and Victoria Land,

which supports the notion of a circum-Antarctic climate optimum. This optimum is not recognised 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 by data from 6 ice-cores, indicated only subtle temperature variations ($\leq \pm 1^\circ\text{C}$) for the past 10 ka BP. Although that study recognised somewhat warmer-than-present conditions between c. 4 - 2 ka BP, it places the Antarctic Holocene climate optimum between c. 10 - 8 ka BP. The Palmer Deep record (Domack *et al.*, 2001a), on the contrary, marks this as a period of cool climate. A study on 11 ice-core isotope records for the reconstruction of Holocene climate variability in Antarctica (Masson *et al.*, 2000) likewise demonstrated a widespread early Holocene climate optimum (11.5 - 9 ka calibrated ice-core years: c. 10 - 8.5 ka radiocarbon years), in addition to a second period of climate amelioration in the Ross Sea area between 7 - 5 ka cal. yr BP (c. 6 - 4.5 ka radiocarbon years BP). These periods of relatively warm climate are not apparent in the geological stratigraphical record. The ice-core data further suggests an East Antarctic climate optimum between 6 - 3 ka cal. yr BP (Masson *et al.*, 2000), whereas the geological record from Bunger Hill Oasis indicates the period 7.6 - 4.5 ka cal. yr BP as being relatively cold, and the climate optimum occurring between 3.5 - 2.5 ka cal. yr BP (Kulbe *et al.*, 2001). It is interesting to note that the early Holocene climate optimum proposed by the ice core data coincides with Antarctic summer insolation minimum (Fig. 17; Berger & Loutre, 1991), whereas the late Holocene climate optimum reconstructed on the basis of the geological data is more compatible with rising temperatures caused by increased insolation (Fig. 17).

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 respond to environmental changes in the peripheral maritime areas of the system. White & Steig (1998) suggested that the inland-plateau ice cores might not be entirely representative of Antarctic climate development, and that records from far more sites are needed to complete the story. Perhaps the poor fit between the ice-core and geological records is an illustration of just that problem.

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