Terrestrial and submarine evidence for the extent and timing of the Last Glacial Maximum and the onset of deglaciation on the maritime-Antarctic and sub-Antarctic islands

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ABSTRACT

This paper is the maritime and sub—Antarctic contribution to the Scientific Committee for Antarctic Research (SCAR) Past Antarctic Ice Sheet Dynamics (PAIS) community Antarctic Ice Sheet reconstruction. The overarching aim for all sectors of Antarctica was to reconstruct the Last Glacial Maximum (LGM) ice sheet extent and thickness, and map the subsequent deglaciation in a series of 5000 year time slices. However, our review of the literature found surprisingly few high quality chronological constraints on changing glacier extents on these timescales in the maritime and sub—Antarctic sector. Therefore, in this paper we focus on an assessment of the terrestrial and offshore evidence for the LGM ice extent, establishing minimum ages for the onset of deglaciation, and separating evidence of deglaciation from LGM limits from those associated with later Holocene glacier fluctuations. Evidence included geomorphological descriptions of glacial landscapes, radiocarbon dated basal peat and lake sediment deposits, cosmogenic isotope ages of glacial features and molecular biological data. We propose a classification of the glacial history of the maritime and sub—Antarctic islands based on this assembled evidence. These include: (Type I) islands which accumulated little or no LGM ice; (Type II) islands with a limited LGM ice extent but evidence of extensive earlier continental shelf glaciations; (Type III) seamounts and volcanoes unlikely to have accumulated significant LGM ice cover; (Type IV) islands on shallow shelves with both
1. Introduction

Reconstructing the Antarctic Ice Sheet through its Last Glacial Maximum (LGM) and post LGM deglacial history is important for a number of reasons. Firstly, ice sheet models require field data against which to constrain and test their models of ice sheet change. The recent development of a practical approach to modelling grounding line dynamics (Schoof, 2007) has led to a new generation of models (e.g., Pollard and DeConto, 2009) that require field constraints. Secondly, the most recent millennia and centuries of ice sheet history provide data on the ‘trajectory’ of the ice sheet, which are valuable for the initialisation of models. Thirdly, the use of recent satellite gravity measurements (e.g. GRACE), and other geodetic data such as GPS, for ice sheet mass balance estimates requires an understanding of glacial–isostatic adjustment (GIA). In the case of GRACE, the satellite-pair cannot distinguish between recent changes in the mass balance of the ice sheet, and those from the transfer of mass in the mantle resulting from past ice sheet melting. This means that robust ice sheet reconstructions are required to generate GIA corrections and it is these corrections that are regarded as the greatest limiting factors for ice mass measurements from satellite gravity (King et al., 2012). It has been suggested that some mass estimates may be in error by as much as 100% (Chen et al., 2006).

Several decades of study have produced an impressive body of work on Antarctic Ice Sheet history. There have been a number of attempts to synthesise the data but many of these have just focussed on the LGM. A notable reconstruction was produced by Ivins and James (2005) which attempted to provide time-slices of the ice sheet from the LGM to the present-day to use as the basis of their GIA modelling. This ‘model’, termed JJ05, has been widely adopted by the satellite gravity and GPS communities as the ice sheet reconstruction to underpin GIA assessments. The model, although a benchmark at the time, is now becoming a little out-of-date, with the proliferation of data since the early 2000s, and no longer includes all of the glacial geological data available.

As a result, the Antarctic Climate Evolution (ACE) and subsequent Past Antarctic Ice Sheet Dynamics (PAIS) programmes of the Scientific Committee for Antarctic Research (SCAR) proposed a co-ordinated effort by the glacial geology community to develop a synthesis of Antarctic Ice Sheet history. This paper covers the maritime and sub-Antarctic sectors. Other sectors of the Antarctic Ice Sheet, including the maritime Antarctic islands west of the Antarctic Peninsula, are described elsewhere in this Special Issue.

Although the combined volume of the maritime and sub-Antarctic LGM glaciers has had a very limited effect on global sea level, understanding past extent and timing of past glaciations in the sub-Antarctic is important for a number of reasons. First, the maritime and sub-Antarctic glaciers have been amongst the earliest ice masses to respond to recent rapid regional warming (e.g. Gordon et al., 2008; Cook et al., 2010) and, therefore, provide a sensitive indicator of interactions between Southern Hemisphere climate and ice sheet stability. This interaction can, in turn, be used to provide boundary conditions for various physical parameters in glaciological models, including those associated with abrupt climate change and the terminal phases of ice sheet decay. Second, the timing, thickness and extent of glacial maxima and subsequent glacier fluctuations in the maritime and sub-Antarctic region can be used to address questions regarding the relative pacing of climate changes between the hemispheres. For example, it is still not known if many of the maritime and sub-Antarctic islands have synchronous glaciations, follow an Antarctic pattern of glaciation, a South American or New Zealand pattern, or a Northern Hemisphere one. This has clear relevance to research aiming to determine if Southern Hemisphere glaciations precede those in the north (or vice versa), whether polar climates are in or out of phase between the hemispheres (Blunier et al., 1998), and in identifying the significant climate drivers. Third, the extent of glacial maxima on the maritime and sub-Antarctic Islands has determined how much of their terrestrial habitats and surrounding marine shelves have been available and suitable as biological refugia for local and Antarctic continental biota during glacial periods (Clarke et al., 2005; Barnes et al., 2006; Convey et al., 2008). This knowledge helps explain evolutionary patterns in biodiversity and regional biogeography.

Whilst for some sectors of the Antarctic Ice Sheet it was possible to follow the original community aim of reconstructing the LGM and deglaciation in a series of 5000 year time slices, our review found surprisingly few high quality age constraints on changing glacier extents on these timescales in the maritime and sub-Antarctic sector. Thus, we limited ourselves to assessing the terrestrial and offshore evidence for the LGM ice extent, and establishing a minimum age for the onset of deglaciation. Specific aims for each of the maritime and sub-Antarctic islands were to:

1. Summarise evidence for LGM ice thickness and extent based on onshore geomorphological evidence, including evidence of glacial isostasy from relative sea level changes.
2. Summarise evidence for LGM ice extent and infer ice thickness using offshore geomorphological evidence from the continental shelf including regional bathymetric compilations.
3. Compile tables of minimum age constraints on glacial features relating to the local LGM (referred to hereon simply as ‘LGM’) and the onset of deglaciation.
4. Separate evidence of the LGM and onset of deglaciation from deglaciation associated with later Holocene glacier fluctuations.

In the discussion we propose a classification of the sub-Antarctic islands based on their glacial history and consider the different climatic and topographic factors controlling glaciation.

1.1. Study area

The sub-Antarctic islands considered in this review are located between 35 and 70°S, but are mainly found within 10–15° of the Antarctic Polar Front (Fig. 1). We also include the South Orkney Islands, and Elephant Island and Clarence Island (the northernmost South Shetland Islands) which are in the maritime Antarctic region (Fig. 1). The remaining South Shetland Islands are covered in the review of Antarctic Peninsula glacial history elsewhere in this Special Issue. Together with the Falkland Islands these sub-Antarctic and maritime Antarctic islands cover an area of approximately c. 26,000 km², just under half the area of Tasmania, or 1.3 times the area of Wales. This figure does not take into account the now-submerged offshore portions of the islands, which considerably increase the total area available for accommodating past glaciation.
We describe the sub-Antarctic and maritime Antarctic islands eastwards around the Southern Ocean, starting with the Atlantic sector then followed by the Indian Ocean and Pacific Ocean Sectors. Other approaches, such as latitudinal position relative to the Antarctic Polar Front, or mean altitude, would be equally valid from a glaciological perspective.

The geological origin of the sub-Antarctic islands has been described in detail by Quilty (2007). Their geological ages range from young volcanic islands such as Bouvet Island, Heard Island and the South Sandwich Islands, to islands composed of tectonically uplifted continental crust such as Macquarie Island or fragments of the continental crust of Gondwana, including islands on the Scotia Ridge such as South Georgia, the South Orkney Islands and Elephant and Clarence Islands.

The climates of the sub-Antarctic islands have been described by Pendlebury and Barnes-Keoghan (2007). However, these are based on measurements for a relatively short instrumental period, often at protected stations close to current sea level. Based on these datasets, mean temperatures of the coolest months range from \(-5^\circ C\) in the South Sandwich Islands to \(+11^\circ C\) at Amsterdam Island. Mean temperatures of the warmest month range from \(+1^\circ C\) in the South Sandwich Islands to \(+18^\circ C\) at Amsterdam Island. Mean annual precipitation ranges from 600 mm in the Falkland Islands to 3200 mm on Gough Island, although precipitation totals at high elevation (e.g., on South Georgia and Heard Island) are poorly constrained and could be considerably higher. The islands are influenced by a number of oceanic fronts including the Antarctic Polar Front, the sub-Antarctic Front and the South Subtropical Front (Fig. 1). All the islands are strongly influenced by the Southern Hemisphere Westerly Winds (mean wind speeds of 6–15 ms\(^{-1}\)), which mediate both the moisture supply required for snow accumulation and also the rate of evaporation and sublimation. Together, the temperature and moisture supply associated with the oceanic fronts, and the Southern Hemisphere Westerly Winds provide controls on the equilibrium line altitude and the thickness and extent of the region’s glaciers.

While falling within the sub-polar belt, several New Zealand sub-Antarctic islands (Snares, Antipodes, Chatham, Bounty), were not considered in this review because they are of low mean altitude and no glacial deposits from the last glaciation have yet been

Fig. 1. Map and classification of the glacial history of the maritime and sub-Antarctic Islands included in this review, shown in relation to the position of the southern boundary of the Antarctic Circumpolar Current (red line), Antarctic Polar Front (yellow line), and sub-Antarctic Front (pink line).
Table 1
Selected radiocarbon ages of peat and lake sediment deposits on the sub-Antarctic islands used here as minimum age constraints for deglaciation. Calibration of radiocarbon dates was undertaken using the CALIB 6.01 and the SHcal04 Southern Hemisphere data set (McCormac et al., 2004). Where dates were beyond the SHcal04 calibration period then the intcal09.14c dataset was used (marked with *). Other superscript markers denote: 4 extrapolated age; ** see stratigraphic comment in Sellevik et al. (1998); 8 Age rejected by the original authors; 1 represents an unreliable minimum age for deglaciation as accumulation of sediments follows a volcanic event; R² calibrated using the Marine 09 data set with a Delta R of 948 (based on a local reservoir correction of 1348 (Herron and Anderson, 1990: p.268) minus the global marine reservoir of 400); the small size of these samples, taken over 5 and 2.5 cm slices, means that the ages from core PC 85–23 are likely to carry significant error; R² calibrated using the Marine 09 data set with a Delta R of 2509 (based on a local core top reservoir correction of 2909 minus the global marine reservoir of 400).

<table>
<thead>
<tr>
<th>Site name</th>
<th>Sample ID</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (m a.s.l.)</th>
<th>Material dated Stratigraphic depth</th>
<th>Reported¹⁴C age</th>
<th>Calibrated age range 2 sigma</th>
<th>Source publication</th>
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<td><strong>Falkland Islands</strong></td>
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<td>Plaza Creek</td>
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<td>51°23′18″S</td>
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<td>Peat</td>
<td>35,970 ± 280 40,521–41,705⁸</td>
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<td>Hooker Point</td>
<td>–</td>
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<td>57°46′49″W</td>
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<td>Peat</td>
<td>–</td>
<td>–</td>
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<td>60°11′27″W</td>
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<td>Peat</td>
<td>13,610 ± 45 16,573–16,950⁸</td>
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<td>Wilson et al., 2002</td>
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<td>52°54′00″S</td>
<td>59°11′00″W</td>
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<td>Peat</td>
<td>–</td>
<td>c. 12,500</td>
<td>Lewis Smith and Clymo, 1984</td>
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<td>Walker Point, Elephant Island</td>
<td>LU-2952</td>
<td>61°08′35″S</td>
<td>54°42′01″W</td>
<td>200–220</td>
<td>Moss peat</td>
<td>5350 ± 60 5927–6211</td>
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<td>Björck et al., 1991</td>
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<td><strong>South Orkney Islands</strong></td>
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<td>S. Orkney Plateau, Site PC85-23</td>
<td>60°49′10″S</td>
<td>45°44′40″W</td>
<td>304(−)</td>
<td>Marine pelecypods, 11,535 ± 900</td>
<td>9442–13,848¹¹</td>
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<td>Herron and Anderson, 1990</td>
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<td>S. Orkney Plateau, Site PC85-23</td>
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<td>45°44′40″W</td>
<td>304(−)</td>
<td>Marine pelecypods, 9570 ± 2180</td>
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<td>Herron and Anderson, 1990</td>
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<td>60°22′5″</td>
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<td>786(−)</td>
<td>Marine sediment, 502 cm</td>
<td>10,542 ± 70 8348–8660²⁵</td>
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<td>Sombre Lake, Signy Island</td>
<td>AA-10691</td>
<td>60°41′12″S</td>
<td>45°37′00″W</td>
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<td>Lake sediment, 250–252 cm</td>
<td>6570 ± 60 7292–7517</td>
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<td>Jones et al., 2000</td>
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<td>Heywood Lake, Signy Island</td>
<td>AA-10704</td>
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<td>45°36′11″W</td>
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<td>Fenton and Smith, 1983; Fenton, 1982</td>
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<td>Site ‘D, Moss Braes Signy Island</td>
<td>BETA-281618</td>
<td>60°68′S</td>
<td>45°62′W</td>
<td>112</td>
<td>Moss bank, 178 cm</td>
<td>2860 ± 40 2784–3006</td>
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<td>Rosqvist et al., 1999</td>
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<td><strong>South Georgia</strong></td>
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<td>Tansberg Point, Lake 1</td>
<td>UA-2991</td>
<td>54°10′02″S</td>
<td>36°41′30″W</td>
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<td>Lake sediment, 499 cm</td>
<td>15,715 ± 150 18,621–19,329²⁹</td>
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<td>Barrow, 1978</td>
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<tr>
<td>Gun Hut Valley, Site 4</td>
<td>SRR-736</td>
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<td>Gun Hut Valley</td>
<td>SRR-1979</td>
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<tr>
<td>Tansberg Point, Dartmouth Point</td>
<td>UTC-4179</td>
<td>54°10′S</td>
<td>36°39′W</td>
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<td>Peat, 308 cm</td>
<td>9520 ± 80 10,512–10,893</td>
<td></td>
<td>Van der Putten et al., 2004</td>
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<td>Husdal, Sink Hole sequence</td>
<td>SRR-1165</td>
<td>54°19′S</td>
<td>36°26′W</td>
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<td>Peat</td>
<td>9433 ± 120 10,264–10,869</td>
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<td>Van der Putten et al., 2009</td>
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<td>UTC-6232</td>
<td>54°10′09″S</td>
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<td>Lake sediment, 447 cm</td>
<td>9060 ± 50 10,116–10,249</td>
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<td>Smith, 1981</td>
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<td>Maiviken</td>
<td>SRR-1162</td>
<td>54°15′S</td>
<td>36°29′W</td>
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<td>Peat, 180 cm</td>
<td>8657 ± 45 9495–9680</td>
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<td>Smith, 1981</td>
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<td>Gun Hut Valley, Site 3</td>
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<td>54°11′09″S</td>
<td>36°41′44″W</td>
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<td>8225 ± 45 9009–9270</td>
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<td>Barrow, 1978</td>
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<td>Husvik Harbour, Kanin Point</td>
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<td>36°41′44″W</td>
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<td>Peat, 312 cm</td>
<td>8225 ± 45 9009–9270</td>
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<td>Van der Putten et al., 2009</td>
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<td>Black Head bog</td>
<td>Beta-271,303</td>
<td>54°04′07″S</td>
<td>37°08′41″W</td>
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<td>Peat, 373 cm</td>
<td>8110 ± 50 8723–9123</td>
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<td>Hodgson, D.A. (unpublished data)</td>
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<td>Lake sediment, 197 cm</td>
<td>7110 ± 40 7788–7969</td>
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<td>Hodgson, D.A. (unpublished data)</td>
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<td>54°29′55″S</td>
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<td>Lake sediment, 584 cm</td>
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<td>Husdal</td>
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<td>63°42′92″W</td>
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<td>Peat, 290 cm</td>
<td>6415 ± 40 7174–7418</td>
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<td>Skua Ridge, First boring</td>
<td>SRR-3214</td>
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reported (McGlone, 2002). The sub-Antarctic islands of the Cape Horn archipelago are also excluded, but readers are referred to Sugden et al. (2005) for a recent review.

2. Methods

This review synthesises the existing literature on maritime and sub-Antarctic island glaciation incorporating earlier brief reviews of the regional glacial history by Hall (2004), Hall (2009) and Hall and Meiklejohn (2011), together with new and unpublished data from the contributing authors. We summarise evidence for late Quaternary (particularly post-LGM) glaciation on each of the islands, and where possible differentiate age constraints derived from robustly defined glacial features with age constraints from features whose provenance and age are less well established. Where age constraints for glacial features are unavailable we identify minimum ages for deglaciation based on, for example, the onset of peat formation and lake sediment deposition.

Where possible the standardised approach for the reporting of age constraints developed by the ACE/PAIS community (see sheet reconstruction team was applied (Tables 1 and 2). For example, radiocarbon dates are reported as conventional ages (with errors) and as calibrated age ranges (2-sigma) and, where required, corrected for marine reservoir effects. Radiocarbon dates were...
Recalibrated with the calibration curves in CALIB 6.01. Where the data are available the type of organic material dated, its location and stratigraphic context are also reported. Evidence of glaciation described in the paper includes: (1) geomorphological and geological evidence for ice presence such as glacial troughs and subglacial till; (2) ice marginal landforms including moraines, till deposits, polished rock and striae, proximal glacialic deposits, and minimum ages for deglaciation from basal peat deposits and lake sediments; (3) ice thickness constraints taken from trimlines, drift limits and exposure age dates, along with indirect constraints from raised marine features and; (4) constraints based on molecular biological data that provide limits on the maximum extent of glaciers (Convey et al., 2008). Further details of data sources are provided within the individual case studies.

3. Results

3.1. Atlantic sector

3.1.1. Falkland Islands

The landscape of the Falkland Islands (51°45’S, 59°00’W, 12,173 km²) is dominated by periglacial features. There is little evidence of LGM ice apart from the small cirques and short (max. 2.7 km) glacially eroded valleys described by Clapperton (1971a) and Clapperton and Sugden (1976). These occur on East Falkland at Mount Usborne and on West Falkland at Mount Adam and the Hornby Mountains. The minimum age of deglaciation of these cirques has not yet been determined, but chronological analyses of basal lake sediments in those occupied by tarns, or cosmogenic isotope analyses of moraines reported in some cirques, could provide this data.

The absence of widespread LGM glaciation at altitude is supported by cosmogenic isotope (10Be and 26Al) surface exposure dates on valley-axis and hillslope stone runs (relict periglacial block streams) which range from 827,366 to 46,275 yr BP (Wilson et al., 2008, Table 2). These old ages suggest not only an absence of large scale glaciation at the LGM, but also the persistence of periglacial weathering and erosion features, through multiple glacial-interglacial cycles. These features include coarse rock debris, silt and clay regoliths, and sand (Wilson et al., 2008). OSL dating of the sediments that underlie some stone runs suggest a period of enhanced periglacial activity between about 32,000 and 27,000 yr BP, and also confirm that parts of the stone runs may have been in existence from before 54,000 yr BP substantially pre-dating the LGM (Hansom et al., 2008).

Peat deposits as old as 40,521–41,705 cal yr BP have been found at Plaza Creek (Clark et al., 1998). Other peat sections, for example at Hookers Point (Long et al., 2005) and Lake Sullivan (Wilson et al., 2002) show peat accumulation commenced at c. 17,000 yr BP (Wilson et al., 2008, Table 2). These old ages suggest not only an absence of large scale glaciation at the LGM, but also the persistence of periglacial weathering and erosion features, through multiple glacial-interglacial cycles. These features include coarse rock debris, silt and clay regoliths, and sand (Wilson et al., 2008). OSL dating of the sediments that underlie some stone runs suggest a period of enhanced periglacial activity between about 32,000 and 27,000 yr BP, and also confirm that parts of the stone runs may have been in existence from before 54,000 yr BP substantially pre-dating the LGM (Hansom et al., 2008).

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3.1.2. Elephant Island and Clarence Island (maritime Antarctic)

Elephant Island (61°08’S, 55°07’W, 5,582 km²) is a 47 × 27 km mountainous island at the northern limit of the South Shetland Islands (Figs. 1 and 2). It has a maximum elevation of 853 m at
Pardo Ridge. Twenty km to the east, Clarence Island (61°12’S 054°05’W) is a 19.3 km long island that rises steeply to 2300 m at Mt Irving (Fig. 2). The islands are part of the Mesozoic Scotia metamorphic complex on the Scotia Ridge (Marsh and Thomson, 1985). Both are heavily glaciated today, with numerous tidewater glaciers. Offshore, bathymetry data show that Elephant Island shares a shallow continental shelf of ~200–600 m water depth with the two smaller outlying Gibbs and Aspland Islands 30–40 km to the south west (Fig. 2A). A significant proportion of this shelf is shallow (<200 m) suggesting the presence of a large area available for ice accumulation during glacial low stands, consistent with the majority of South Shetland Islands and the western Antarctic Peninsula.

In contrast, the bathymetry surrounding Clarence Island falls away steeply on all sides to ocean depths of at least 600 m. There are no clear glacial troughs radiating from Elephant Island in existing bathymetric datasets, but there appears to be an over deepening (a trough in excess of 1300 m water depth) in the breach between Elephant Island and Clarence Island to the east. Within this trough, there is no evidence of former ice grounding, for example in the form of streamlined bed forms as observed in troughs elsewhere along the west Antarctic Peninsula shelf (Fig. 2B). Instead, sets of sinuous ridges and channels are observed which are partially covered by a substantial sediment infill, forming flat and featureless bathymetric zones in the base of the trough. While we cannot rule out a glacial origin for these ridge/channel features (e.g., as subglacial eskers or meltwater channels), there is no indication in the surrounding valley sides for substantial glacial moulding of the landscape and thus former ice overriding. At the shelf break around Elephant Island, multibeam data are similarly inconclusive over the presence or absence of geomorphic features that might have formed at grounding line positions if local ice had extended towards the shelf break in the past.

No marine geochronological data constraining offshore ice extent or deglaciation have been reported. At Elephant Island a basal age from the deepest known moss bank in Antarctica at Walker Point provides a minimum age for local deglaciation onshore of 5927–6211 cal yr BP (Björck et al., 1991).

### 3.1.3. South Orkney Islands (maritime Antarctic)

The South Orkney Islands (60°35’S, 45°30’W, 620 km²), an archipelago located 600 km north-east of the tip of the Antarctic Peninsula, comprises four main islands: Coronation Island, Laurie Island, Powel Island and Signy Island. Their geology consists of folded metamorphic sediments (Matthews and Malling, 1967) forming part of the Scotia Ridge. Geomorphological mapping by Sugden and Clapperton (1977), together with seismic data and piston cores obtained from the South Orkney Islands plateau during DF-85 (USCGC Glacier) by Herron and Anderson (1990), provide the only published data constraining the offshore extent of grounded ice at the LGM. These studies described several offshore glacial troughs fed by glaciers draining an expanded ice cap. A seismic profile across the western plateau showed a prominent glacial unconformity between the 250–300 m isobaths, interpreted as marking the limit of grounded ice during the most recent phase of extensive glaciation (Herron and Anderson, 1990; Bentley and Anderson, 1998). To constrain the age of this unconformity, piston cores and sediment grabs were recovered from 35 locations. Only a handful of the cores penetrated glacier proximal/subglacial till but nevertheless confirmed that grounded ice reached to at least 246 m water depth (core 85–21) but possibly as deep as 311 m (core 85–30). Radiocarbon analyses of articulated pelecypod shells found within diatomaceous glacial marine sediment at South Orkney Plateau Site PC 85–23 indicated that the ice cap had retreated from the inner portion of the plateau and to within 15 km of Signy Island prior to 9442–13,848 cal yr BP (11,535 ±14°C yr BP, Table 1) (Herron and Anderson, 1990); although this had previously been reported as c. 6000–7000 years BP based on calculated accumulation rates (Herron and Anderson, 1990; Bentley and Anderson, 1998). Consistent with this deglaciation age, diatom ooze layers were accumulating at another site on the plateau by at least 8348 to 8660 cal yr BP (Lee et al., 2010). Analyses of the ice rafted debris (IRD) assemblage in slope cores, which were composed predominantly of material derived from the South Orkney Islands, led Herron and Anderson (1990) to speculate that the outer shelf was covered by a large ice shelf at the LGM. The presence of a much more extensive regional ice shelf, connecting the

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**Fig. 2.** (A) Regional bathymetric plot showing the large shallow continental shelf (<200 m depth) connecting Elephant Island, Gibbs Island and the Apsland Islands. In contrast, Clarence Island falls away steeply on all sides to ocean depths of at least 600 m. (B) Elephant Island and Clarence Island are separated by an over deepened trough in excess of 1300 m water depth with sinuous ridges and channels partially covered by a substantial sediment in fill. Note the different orientations of the figures.
South Orkney Ice cap with the Antarctic Peninsula Ice Sheet at the LGM has also been suggested by Johnson and Andrews (1986). However, this hypothesis is based on limited geological data and, whilst ice sheet models imply a more extensive ice shelf (Colledge et al., 2012; Pollard & DeConto, 2009), they indicate it was not contiguous with the Antarctic Peninsula. Thus, the alternative interpretation that the Antarctic Peninsula Ice Sheet and South Orkney Ice Cap behaved as independent ice centres must still be considered. New marine geological and geophysical data acquired from the South Orkney shelf by RRS James Clark Ross in 2011 (JR244) will hopefully resolve this issue (W. Dickens, personal communication).

On-shore, a minimum age for deglaciation can be inferred from the onset of lake sediment accumulation which began at Signy Island between 7292 and 7517 cal yr BP (Sombre Lake) and 6484–6791 cal yr BP (Heywood Lake) (Jones et al., 2000). Moss banks accumulated on Signy Island from 4799 to 6183 cal yr BP (Fenton, 1982; Fenton and Smith, 1983) and 2784–3006 cal yr BP (Royles et al., 2012).

3.1.4. South Georgia

South Georgia (54°17′S, 36°30′W, 3755 km²) is a large heavily glaciated island 170 km long and 39 km wide dominated by the continental rock of the Allardyce and Salvesen Ranges, with the highest peak at Mt Paget (2934 m). Glacial geomorphological research on South Georgia is more advanced than most islands of the sub-Antarctic and includes studies on both the terrestrial and submarine glacial geomorphology, together with age constraints from lake sediments, peat deposits and moraines. Compilations of bathymetric soundings from the continental shelf have revealed large cross shelf glacial troughs, moraines and trough mouth fans on the shelf and adjacent slope (Graham et al., 2008). These observations suggest that one or more glaciations have extended to the continental shelf break (Fig. 3A) with their isostatic signature recorded by the raised beaches found at onshore altitudes of 6–10 m, 52 and 124 m a.s.l. (Clapperton et al., 1978). Early work assumed that the LGM glaciers extended across the continental shelf, although there was no chronological control on these periods of extensive shelf glaciation (Clapperton, 1990). More recent evidence based on the submarine geomorphology of the coastal fjords (Hodgson et al., 2014), combined with age constraints on land (Bentley et al., 2007) suggest that these continental shelf glaciations probably pre-date the LGM and that the LGM glaciers were most likely restricted to the inner fjords. The possibility that cold-based, generally non-erosive glaciers, were present at the LGM has not yet been considered in the literature.

Further evidence that the LGM was restricted to the inner fjords includes geomorphological mapping and cosmogenic isotope and radiocarbon dating of the onshore Lateglacial to Holocene moraines (Clapperton, 1971b; Sugden and Clapperton, 1977; Clapperton and

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Fig. 3. (A) Map of the South Georgia continental block illustrating well-developed glacial cross-shelf troughs (bathymetric data from Fretwell et al. (2009). (B) Cumberland East Bay, South Georgia showing an example of the oldest dated terrestrial category ‘a’ moraines at the northern end of the Greene Peninsula east of Moraine Fjord (from Bentley et al. (2007)). These are shown, together with shipborne swath bathymetry data presented in Hodgson et al. (2014), illustrating the fjord-mouth (‘inner basin’) moraines in Moraine Fjord and Cumberland East Bay which are of presumed similar age. Bathymetry is shown at 5-m grid cell size.
10,570 cal yr BP (Van der Putten et al., 2012) followed by the onset were deposited at Husdal in Stromness Bay prior to 10,113 e

The submarine geomorphology is dominated by features related to slope instability and volcanism and no distinct glacial features have been identified and include lichen growth rate evidence from a series of ice-free moraine ridges down slope of two small mountain cirques in Prince Olav Harbour. These dates are considered reliable as minimum age constraints for deglaciation as they are either based on plant macrofossils at the base of peat sequences or lake sediments, or on bulk basal lake sediments in which radiocarbon reservoirs are absent or well constrained. Raised marine features, interpreted as raised beaches, are also found at a relatively low level around the north east coast of South Georgia (2–3 m a.s.l. in Clapperton et al., 1978; <10 m a.s.l. in Bentley et al., 2007). Some of these features have been reinterpreted as being the result of fluvo-deltaic deposition at higher relative sea levels, such as the c. 9 m a.s.l. ‘Line M’ in Stromness Bay which may mark the inland position of a former coast line in Husdal (Van der Putten et al., 2013). Both interpretations imply a maximum of <10 m of post-glacial rebound since exposure of these areas by Holocene ice retreat, and in most cases just 2–3 m, although these features remain undated. The implication of these data taken together is that large parts of the South Georgia coastline, particularly the peninsulas along the north coast, were free of grounded ice very early on in the post-glacial interval—and possibly during the LGM—and that, contrary to previous suggestions (Clapperton et al., 1989), the LGM extent of the South Georgia ice cap was restricted to the inner fjords.

Late Holocene glacier fluctuations on South Georgia have also been identified and include lichen growth rate evidence from a series of ice-free moraine ridges down slope of two small mountain cirques in Prince Olav Harbour. These suggest ice retreat from the outermost moraines occurred between the end of the ‘Little Ice Age’ (post c. 1870) and the early 20th century, and from the innermost moraines during the second half of the 20th century (Roberts et al., 2010). The latter retreat has been linked to the well-documented warming trend since c. 1950 and can also be seen in the extensive photographic record the retreat of glacier fronts around South Georgia (Gordon et al., 2008; Cook et al., 2010).

Although our understanding of glaciation is relatively advanced for South Georgia, at least compared with other sub-Antarctic

in Prince Olav Harbour at 8723–9123 cal yr BP (Hodgson D.A. unpublished data), and 7571–7690 cal yr BP and 7174–

7418 cal yr BP at Husdal (Van der Putten et al., 2013) (Table 1).
islands, there still remains a paucity of chronological control to constrain ice cap positions through the last deglaciation, particularly at ice-marginal positions offshore.

3.1.5. South Sandwich Islands

The South Sandwich Islands (56°20’S, 26°00’W to 59°20’S, 28°00’W, 618 km²), comprise a 390 km long chain of submarine volcanic edifices that emerge as small volcanic islands at the eastern periphery of the Scotia Sea. The ten islands are strongly influenced by cold ocean currents from the Weddell Sea. They are up to 90% permanently ice covered (e.g., Montague Island) with variations in ice cover being the result of differences in altitude, and heat flow from the eruption of volcanoes. Areas of shallow shelf surrounding each edifice are limited, preventing widespread glaciation. The submerged slopes that flank the islands are mostly steep and fall away sharply into water depths >500 m depth (Leat et al., 2010) (Fig. 4) and many of the islands exhibit dynamic erosional coastlines (Allen and Smellie, 2008; Leat et al., 2010). Thus, any potential thicker ice cover at the LGM would have likely remained localised to the island summits and would have been restricted to extents very similar, if not identical, to those today. A close inspection of available multibeam bathymetric data for the South Sandwich arc confirms that no distinct glacial features are preserved in the sea-floor record, instead being dominated by features related to slope instability and volcanism (Leat et al., 2010) (Fig. 4). No studies have been carried out on the late Quaternary glacial history offshore, and there are no age constraints.

3.1.6. Bouvet Island

Bouvet Island or Bouvetøya (54°26’S, 3°25’E, 50 km²) is located south of the Antarctic Convergence (Fig. 1). It consists of a single dominant active cone volcano (Fig. 5A). It is heavily ice-covered (~92%, Hall, 2004, Fig 5B) with many hanging glaciers discharging at the present coastline. A recent review (Hall, 2009) found that information on Quaternary glaciation is limited to observational data on glacier extent through the 20th century, with ice front variations in the order of 10−100 m (Mercer, 1967; Orheim, 1981). These were attributed to differences in aspect relative to the prevailing winds, as well as to local tidewater effects. The island consists of young oceanic crust, 4−5 Ma in age (Mitchell, 2003). On land, any record of Quaternary glaciations may have been obscured by continuing volcanism and tectonic activity, or remains covered today by extensive snow and ice. Offshore, the limited bathymetry data show a 3−4 km-wide shelf of <200 m water depth (Fig. 5A). Hence, even with extensive ice grounding onto the submarine shelf, any former ice cap on Bouvet Island probably had an aerial extent no larger than ~330 km². Even with complete ice cover, this would be comparable in size to some of the smaller glacier systems in Svalbard and the Southern Patagonian Ice Field today (World Glacier Monitoring Service, 1999, updated 2012; www.geo.uzh.ch/microsite/wgms/).

3.1.7. Gough Island

Gough Island (40°21’S, 9°55’E, 65 km²) is a young (1 Ma) volcanic island. The island is not glaciated today, and appears to have no evidence of former glaciation. Bennett et al. (1989) dated a bedded, polleniferous peat sequence from the south-east of the island. They recovered an infinite radiocarbon age of ~43,000 14C yr BP from the basal sediments, and argued for a continuous occupation by the local flora through the last glacial-interglacial cycle. The well developed terraces around the coast (from −50 m below sea level to 75 m above sea level) are also considered to be the result of...
Fig. 6. (A) Reconstruction of palaeo-glaciers with limited offshore extent on sub-Antarctic Marion Island, based on glacial bedform evidence and landscape interpretations presented in Hall and Meiklejohn (2011). (B) Satellite image showing the position and orientation of some of the outer kelp beds, which may reveal the presence of offshore lateral-frontal moraines from which former glacier positions can be inferred. An alternative hypothesis is that the kelp beds mark the termination of submarine lava flows.
progressive elevation during the evolution of the volcano, with those above sea level being related to coastal erosion at the last interglacial (3–4 m), Middle Pleistocene (12–15 m and 30 m) and earlier Pleistocene (55 m) and older (75 m). Those below sea level are thought to be related to erosion during intervals of lower sea level during glaciation(s) (Quilty, 2007).

3.2. Indian Ocean Sector

3.2.1. Marion Island and Prince Edward Island

Marion Island (46°55′S, 37°45′E, 293 km²) and Prince Edward Island (46°39′S, 37°57′E, 46 km²) are young (0.45 Ma) active volcanic islands (McDougall et al., 2001; Boelhouwers et al., 2008) located on top of a small submarine plateau with a rapidly disappearing ice cap (Sumner et al., 2004). Up to eight volcanic and dacitic islands (McDougall et al., 2001; Boelhouwers et al., 2008) have been inferred (Fig. 6B); similar to the kelp beds seen at the entrance to shallower water) may be revealing the presence of offshore terrigenous deposits intercalated with lavas (McDougall et al., 2001). Some of the earlier volcanic episodes were correlated with glacial stages (Marine Isotope Stages 2, 4, 10 and 12) and the four most recent episodes correlate or overlap with interglacials (Marine Isotope Stages 1, 3, 5, 7). This data, together with recent geomorphological evidence (Boelhouwers et al., 2008), meant that the original hypothesis, that faulting and volcanic activity on Marion Island were periodically triggered by deglaciation (Hall, 1982), had to be reassessed (Hall et al., 2011).

The most recent advances in understanding the late Quaternary glacial and LGM glacial geomorphology of Marion Island are summarised by Boelhouwers et al. (2008) and Hall et al. (2011). These studies both suggest that the island was covered by a large LGM ice mass that separated into individual glaciers near their terminal margins (Fig. 6A). Raised beaches are also present which may be the result of isostatic rebound following deglaciation (Hall, 1977), or tectonic uplift. Thick tills at the present coastline, and the location and orientation of lateral moraines (e.g. flanking Long Ridge), suggest the likelihood of extensive seaward expansion of glaciers during times of lower glacial sea levels. Therefore, offshore evidence of the maximum extent of glaciers should be preserved on the continental shelf. Even though there are no high resolution bathymetric data for the coastal margins of the island, analysis of satellite images it may be possible to resolve glacial features similar to the moraines and other glacial features found on Marion Island, although elsewhere on Albatross Ridge peak core basal ages of only 6601–6950 cal yr BP (depth: 353–363 cm) and 4426–4744 cal yr BP (depth: 165–180 cm) have been reported (Scott, 1985). On nearby Skua Ridge the oldest peak basal age is 7574–7873 cal yr BP and at Kildale Bay it is 7934–8198 cal yr BP (Scott, 1985). As all these sites overlie old grey lavas they are considered reliable minimum ages for deglaciation. Other peat cores that have been taken on the island were dated at 3180 ± 20 14C yr BP (3316–3403 cal yr BP; Junior’s Kop), 4020 ± 65 14C yr BP (4225–4587 cal yr BP; near the Marion Base Station), 2685 ± 130 14C yr BP (2351–3005 cal yr BP; Nellie Humps Valley) (Schalke and van Zinderen Bakker, 1971) and 4750 ± 40 14C yr BP (5316–5485 cal yr BP; near the Marion Base Station) (Yeloff et al., 2007), but these overlie Holocene black lava flows they provide minimum age constraints on these volcanic episodes rather than deglaciation.

Some late Holocene (possibly Little Ice Age) ice advances have been inferred from striated basalt surfaces (Hall et al., 2011) and geomorphological evidence of Holocene ice is present in small cirque basins at Snok and the summit of the island (Boelhouwers et al., 2008). Similarly, perennial high altitude late Holocene snow cover and volcanic activity have been suggested from the absence of the large-scale relic periglacial landforms above 750 m a.s.l (Boelhouwers et al., 2008; Hedding, 2008). The last remnants of the Holocene ice cap had largely disappeared by the late 1990s (Sumner et al., 2004), presumably as a result of regional climate changes and/or geothermal activity (c. 1980 AD).

On nearby Prince Edward Island, Verwoerd (1971) found no geomorphological evidence of glacial activity. Whilst he attributed this to the lower altitude of the island, which rises to 672 m compared with Marion Island at 1240 m, he considered it unlikely that the island had entirely escaped Quaternary glacialiation. From satellite images it may be possible to resolve glacial features similar to the moraines and other glacial features found on Marion Island, but further analysis and ground-truthing are required.

3.2.2. Crozet Islands

The Crozet Islands (46°25′S, 51°38′E, 400 km²) consist of five main oceanic islands situated in the southern part of the Indian Ocean (Fig. 1). They are volcanic, built by several magmatic events which started about 8.1 Ma (Lebouvier and Frenot, 2007; Quilty, 2007). The islands are currently free of ice, but there is evidence of strong glacial erosion resulting in a series of radially arranged glacial valleys, a major cirque complex and related moraines on Île de l’Est, and three steep sided U-shaped valleys, likely glacial origin on Île de la Possession (Vallée des Branlaires, Baie de la Hébé, Baie du Petit Caporal) (Lebouvier and Frenot, 2007; Quilty, 2007), together with mapped moraines and lakes formed by glacial activity (Chevallier, 1981). This suggests the presence of Quaternary glaciers (Camps et al., 2001; Giret et al., 2003), although earlier papers have suggested these may pre-date the LGM (Chevallier, 1981; Giret, 1987; Bougere, 1992; Hall, 2009) or were not glacial features (Bellair, 1965). Offshore, examination of bathymetric compilations shows no clear indication for past glaciations, although a significant portion of the surrounding sea-floor
lies at shallow depths, indicating the potential for
more extensive ice accumulation during glacial lowstands (Fig. 5C).

There is no chronology on glacial extents since the LGM, but
palaeoenvironmental records suggest that Baie du Marin (close to
the base Alfred Faure) must have been free of ice at 10,750–
11,000 cal yr BP based on organic sediment layers in peat cores (Van
der Putten et al., 2010) (Table 1). Additional dates from the Mourne
Rouge flank in the Vallée des Branloires of 6779–7020 cal yr BP
(Ooms et al., 2011) and basal dates from Mourne Rouge Lake of
6389–6640 cal yr BP and a peat sequence of 6000–6316 cal yr BP
(Van der Putten et al., 2008) have also been published, but because
these are from within the Morne Rouge volcano they are indicative
of a minimum age for the eruption rather than a minimum age for
deglaciation.

3.2.3. The Kerguelen Islands

The Kerguelen Islands (48°30'S, 68°27'E and 50°S, 70°35'E)
consist of a main island (7200 km²) surrounded by numerous
smaller islands of mostly ancient (39–17 Ma) volcanic origin.
The main island is characterised by mountains up to 1850 m (Mt Ross),
the large 403 km² (in 2001) Cook Ice Cap on Le Dome (1049 m), and
several glaciers on the western part of the island (Fig. 7). The
eastern part of the island is generally of lower relief, but includes
widespread evidence of glacial striations, glacial outwash and
glacial moraines (Quilty, 2007).

Despite being one of the sub-Antarctic islands that remain
partially glaciated, there is remarkably little information on the
Quaternary glacial history of the Kerguelen Islands. Some studies
have suggested that the main island may have been completely
covered at the LGM (Hall, 1984); an interpretation at least partly
supported by the presence of numerous ice-scoured lake basins
(Heirman, 2011), U-shaped valleys radiating from the Cook Ice Cap,
deeply-incised fjords and the lack of terminal moraines, which
implies that ice may have extended offshore (Bellair, 1965). How-
ever, other studies have suggested that the LGM glaciation was
limited (Nougier, 1972), and this is supported by the absence of
present day isostatic rebound (Testut et al., 2005). This latter theory
suggests that glaciers were restricted to the central plateau and to
the east and south west where there are glacial erratics, aeolian
sands, depressions filled with peat, gelifraction soils and moraine
complexes, as well as residual valley glaciers and cirques.

Conversely, in the north the highly degraded morphology of the
moraines in the Loranchet Peninsula and the near absence of glacial
erratics has been interpreted as evidence of more ancient glaciation
(Nougier, 1972).

There are no chronological constraints on maximum glacier
extent at the LGM. However, there are reliable minimum bulk

radiocarbon ages for deglaciation from peat deposits at Estacade, the Golf du Morbihan (Young and Schofield, 1973a), and the Baie d’Ampère (Fig. 7B), and geomorphological observations on the Gentil glacial moraines at the base of Mont Ross (Fig 7D). The oldest peat deposit at Estacade dates from 15,396\(\pm\)6,624 cal yr BP (Van der Putten et al., 2010) and at the Golfe du Morbihan from 12,765\(\pm\)13,241 and 9141\(\pm\)9912 cal yr BP (Young and Schofield, 1973a, 1973b). In the Baie d’Ampère the recent (post 1990 AD) retreat of the front of Ampère glacier has re-exposed a series of early Holocene peat deposits (Frenot et al., 1997b). One group provides minimum ages for deglaciation between 13,241 and 11,212 cal yr BP (Table 1, sample numbers 1–3, Fig 7C). These can be clearly separated from later periods of Holocene glacial retreat from 5054\(\pm\)5188 cal yr BP (Table 1, sample number 4, Fig. 7C), and 2208 to 716 cal yr BP (Table 1, sample numbers 5–9, Fig. 7C) that may correspond to warm periods inferred from peat deposits (e.g., Young and Schofield, 1973a). Other older frontal and lateral moraines associated with the Gentil Glacier have been identified at the base of Mont Ross (Fig 7D). It is not known if these date from the LGM, but they must predate AD 934\(\pm\)46 (1016 cal yr BP) based on the absence of a diagnostic ash layer from the Allouarn Volcano (Arnaud et al., 2009). In terms of maximum ice thickness, erosional evidence produced by the ice flow on rock cliffs on both sides of the valley above Lac d’Ampère reveal that the surface of the glacier was about 150 m higher than today during the maximum Holocene extent. Whether this is equivalent to the LGM ice thickness is not known. The lack of remains of lateral or frontal moraines on the slopes of both sides of the valley may indicate that previous Holocene glacial extents were smaller than those of the last millennium or that at its maximum the glacier reached positions in the fjord that are submerged offshore today. The possibility that cold-based, generally non-erosive glaciers, were present at the LGM has not yet been considered in the literature.

Collectively, the evidence from the moraines suggests that the Kerguelen glaciers are highly sensitive to climate changes and that various Holocene ice advances may have approached LGM ice maxima. For example, various studies have shown that the Ampère Glacier has advanced and retreated up to 3.8 km from its 2010 front position on multiple occasions in the late Holocene (Frenot et al., 1993, 1997a; Arnaud et al., 2009).

Recent glacier retreat has been documented from the first half of the 20th century (Aubert de la Rue, 1967; Vallon, 1977) and the total ice extent on Kerguelen Islands declined from 703 to 552 km² between 1963 and 2001, with the Cook Ice Cap retreating from 501 to 403 km² in the same period (Berthier et al., 2009). Current rapid deglaciation at the Kerguelen Islands is exceptional (Cogley et al., 2010) and possibly linked to increased temperature (Frenot et al., 1993, 1997a; Jacka et al., 2004), and decreased precipitation since AD 1960 (e.g., Frenot et al., 1993, 1997a; Berthier et al., 2009). An alternative hypothesis is that the retreat is related to migration of the sub-Antarctic convergence from the north to the south of the Kerguelen Islands around AD 1950 (Vallon, 1977).

3.2.4. Heard Island and McDonald Island

Heard Island and McDonald Island (located at approximately 53°06’S, 73°31’E) are 380 km² in area. Heard Island consists of an active strato-volcano, Big Ben (2745 m), situated just south of the present day Polar Front. It is heavily glaciated with ice covering 70% or 257 km² of the island, with 12 major glaciers radiating towards the sea from the summit of Big Ben or the peaks of Laurens Peninsula (McIvor, 2007). The island is one of the few exposures of

![Fig. 8.](image-url)
the Kerguelen Plateau, the second largest submarine plateau on Earth. It comprises young volcanic material that has built on top of the Late Miocene – Early Pliocene Drygalski Formation, which today forms a flat 300 m high plateau off the northern coast of Heard Island (Kiernan and McConnell, 1999).

There are no published data on Heard Island’s glacial history since the LGM, with the exception of descriptions of till and moraine formation (Lundqvist, 1988), and the Dovers Moraines; a series of lateral moraines and extensive hummocky moraines (Kiernan and McConnell, 1999), which are undated, but most likely of Holocene age (Hall, 2002).

Some of the glaciers reach sea level today. Offshore on the continental shelf a bathymetric grid compilation (Beamond and O’Brien, 2011) shows evidence of extensive glaciation with at least four, and possibly more, large cross-shelf troughs and moraines extending as much as 50–80 km from the present shoreline (Balco, 2007) (Fig. 8A), but the age of these features remains unknown. Based on their position and depth, these features would require grounded ice to a depth of at least 180 m and a palaeo-grounding line at 120 m below the LGM sea level (Hall et al., 2011). This observation suggests the ice was a minimum of 135 m thick at its margin and, hence, several hundred metres thick at its centre (Balco, 2007). New bathymetric data for the sea-floor plateau surrounding Heard Island now exist at a resolution that permits a closer analysis of these submersed glacial features (~100 m grid cell size; Fig. 8B). The moraine belt is well-resolved over a distance of ~80 km on the new bathymetric grids but is not resolved to the west, east and south of the plateau. Where it is well-resolved, the moraine belt is broadly symmetric in profile, 50–80 m high and up to 4 km in width. The size of the feature suggests it is a terminal moraine of a larger ice cap that covered significant portions of the island and its marine plateau in the past. Balco (2007) also observed over-deepened troughs, likely of glacial origin, that cut across the shelf inshore of the moraine. These are clearly represented in the new bathymetry (Fig. 8B) and suggest that the ice cap was organised into several discrete fast-flowing outliers, in common with most examples of ice caps and ice sheets today.

Sketches of more extensive glaciers by visiting sealers between 1850s–1870s AD, and photographic evidence, documents glacial retreat over the latter half of the 20th Century (Kiernan and McConnell, 1999, 2002; Ruddell, 2005; Thost and Truffer, 2008). This may be linked to a shift in the position of the Polar Front which now regularly migrates to the south of Heard Island. A radiocarbon date of modern to 340 cal yr BP (220 ± 13 14C yr BP; Wk 9485) from plant material buried beneath beach gravels at Long Beach provides a local minimum age for deglaciation at that site (Kiernan and McConnell, 2008), but is not related to the retreat of an LGM ice cap. Nevertheless, the relatively small area of the island that has periglacial features does suggest that onshore deglaciation has been relatively recent and this may also explain why glacioisostatic features such as raised beaches have not been described. Well-formed vegetation banked terraces occur at Mt Andree possibly marking one of the longest exposed areas (Kiernan and McConnell, 2008), but these have not been dated. As Heard Island contains abundant volcanic deposits such as lava flows and tephra, there is potential to use these in future to help constrain the glacial history.

Nearby McDonald Island, approximately 40 km to the west, has undergone recent volcanic activity, notably in the AD 1990s when the main island was observed to have doubled in size (Melvor, 2007). There is no published information on its glacial history.

3.2.5. Amsterdam and St Paul Islands

Situated between South Africa and Australia, Amsterdam Island (37°50’S, 77°30’E, 55 km²) and Saint-Paul Island (38°43’S, 77°31’E, 6 km²) are volcanic islands dating from about 400 to 200 ka (Lebouvier and Frenot, 2007; Amsterdam Island) which have evidence of continued and recent volcanic activity. No glacial geomorphological data are published.

3.3. Pacific sector

3.3.1. Macquarie Island

Macquarie Island (54°37’S; 158°54’E, 200 km²) is situated north of the Polar Front (Fig. 1). It is nearly 34 km long and up to 5 km wide and largely consists of a high plateau of between 150 and 300 m a.s.l with the highest point being Mt Hamilton (433 m). There are no permanent snowfields or glaciers. The island consists entirely of oceanic crust together with remnants of submarine volcanoes. These ocean-floor rocks belong to the Miocene Macquarie Ridge, which stretches from around 61°S to New Zealand (Carmichael, 2007). The island emerged 4000 m above the ocean floor about 600,000–700,000 years ago and the current tectonic uplift rate is somewhere between about 0.8 mm yr⁻¹ (Adamson et al., 1996) and 1.5 mm yr⁻¹ (Colhoun and Goede, 1973). Therefore the palaeo-beaches, terraces and cobbles seen around the island are not interpreted as being the result of glacioisostatic uplift, but of marine erosion and deposition during the geological uplift of the island over the last six Quaternary glacial-interglacial cycles.

The scientific debate concerning the glacial history of Macquarie Island is summarised in Selkirk et al. (1990). In brief, early interpretations of glacial features such as erratics, polished and striated cobbles, moraines, kame terraces, over-deepened lakes, meltwater channels, glacial valleys and cirques on the island (Mawson, 1943; Colhoun and Goede, 1974; Löffler and Sullivan, 1980; Crohn, 1986) have now been explained as topographic expressions of faulting and non-glacial erosion associated with the tectonic uplift of the island, and as periglacial features (Ledingham and Peterson, 1984; Adamson et al., 1996). This is supported by the presence of multiple raised beaches with thermoluminescence ages of 340 ± 80 ka at Hasselborough Bay, 263 m asl and 172 ± 40 ka (at Wireless Hill, 100 m asl) attributed to Marine Isotope Stage 9 (340–330 ka) and Stage 5e (130–125 ka), respectively. Although the TL errors are very large, these dates imply that the island has not been subject to extensive glacial erosion (Adamson et al., 1996). A thermoluminescence date of 92 ± 120 ka from a lacustrine deposit exposed in a bank of North Bauer Creek suggests that lake sediments accumulated in the early half of the last glacial cycle between Oxygen Isotope Stage 9 and the middle of Stage 5 (Adamson et al., 1996). This deposit was subsequently overlain by periglacial mass flows that accumulated during the last glacial. Peats with infinite radiocarbon ages of >40,000 yr BP have also been found at West Mount Eitel (Adamson et al., 1996). These peats overlain rounded beach cobbles, and in turn are overlain by a thick deposit of sub-angular matrix-supported cobbles (the likely product of periglacial conditions), capped by a thick sandy peat with present-day vegetation.

A near island-wide periglacial environment most likely persisted until just after the peak of the LGM, following which radiocarbon evidence shows that periglacial conditions moderated sufficiently to permit the continuous deposition of lake sediment and terrestrial peat deposits. These date from 15,975–17,034 cal yr BP to 14,063–15,119 cal yr BP at Palaeolake Skua (Selkirk et al., 1991) and 11,281–12,581 cal yr BP at the Finch Creek Ridge peat deposit (Selkirk et al., 1988) (Table 1). Sediments in extant lakes date from 16,620–16,987 cal yr BP (Saunders, K. Unpublished data). These can only be considered minimum ages for the transition from periglacial conditions as basal ages have yet to be determined for some of the lakes, such as Palaeolake Touche (Selkirk et al., 1988). Younger deposits dated to 8185–8639 cal yr BP are found at Palaeolake Sandell. In addition, peat deposits dating to...
7682–8203, 5586–6200 and 6206–7272 cal yr BP occur at Green Gorge, Wireless Hill and Finch Creek ridge (Selkirk et al., 1982; Selkirk et al., 1988) (Table 1).

Combined with evidence for tectonic uplift, geomorphological observations suggest extensive periglacial rather than glacial processes were the most important in fashioning the cold uplands of Macquarie Island. This has resulted in the formation of turf banked and stone banked terraces in several locations, mainly on the leeward eastern parts of the island (Selkirk et al., 1990; Selkirk, 1998; Selkirk-Bell and Selkirk, 2013). Whist there may have been small nivation cirques on some areas of the plateau during glaciations (Hall, 2004) there is no evidence for any former ice caps or glaciers (Ledingham and Peterson, 1984; Adamson et al., 1988). Similarly, early suggestions that the island’s present biota arrived by long-distance dispersal following retreat of an overriding ice sheet (Taylor, 1955) have also subsequently been disproven (Van der Putten et al., 2010). On the basis of this evidence we concur with Selkirk et al. (1990) and Adamson et al. (1996) in concluding that there is no compelling evidence of LGM glaciation of Macquarie Island.

3.3.2. Campbell Island

Campbell Island (52°33'S, 166°35'E, 120 km²) is the southernmost of the New Zealand sub-Antarctic Islands. It is of ancient volcanic origin (6±11 Ma), being a remnant of a shield volcano centered approximately on Dent Island (Fig. 9A). It is not currently glaciated.

There are no detailed studies on the glacial history of Campbell Island. However, there are several prominent U-shaped valleys including Perseverance Harbour and Northeast Harbour. These appear to be arranged in a more or less radial pattern, around the former volcanic centre at Dent Island (Fig. 9A). This suggests that the radial drainage pattern may have been established in the mid to late Miocene before the western part of the former shield volcano was eroded away by the sea, or experienced a flank collapse, leaving only the eastern part. This hypothesis would explain why the main U-shaped valleys, particularly Perseverance Harbour, now appear to lack a sufficient ice catchment area to explain their large scale relative to the remaining landmass of island (Fig. 9A).

Unequivocal evidence of LGM glaciation is not very obvious on the island (Quilty, 2007). Nevertheless, glacial features such as ‘corrie and moraine’ (most likely on Mt Honey) have been reported since at least 1896 (Marshall, 1909). Geomorphological features associated with glacial U-shaped valleys have also been described (Marshall, 1909; Quilty, 2007). These include early soundings from Perseverance Harbour that suggested that it was over deepened by ice derived from glaciers at Mount Honey and Mount Lyall, with a valley east of Mount Honey being interpreted as a hanging valley occupied by an ice tributary to a larger glacier (Marshall, 1909). A sill at the entrance of Perseverance Harbour has been variously interpreted as a glacial till, or debris associated with longshore currents (Quilty, 2007). Studies by McGlone et al. (1997) and McGlone (2002) also described cirque-like features at around 150 m a.s.l. on the higher mountains, diamictons interpreted as tills, and a possible lateral moraine composed of a bouldery sandy gravel, 2–3 m thick, exposed at the top of the 90 m high Hooker sea cliff in the north of the island. Possible kame terraces, terminal moraines, and erratic blocks have also been considered as evidence of extensive ice cover at the LGM.

Most, if not all, of the glacial features on the island remain undated and even those that are present are either covered by a deep blanket of overlying peat (McGlone, 2002), are heavily eroded, or unproven. In Perseverance Harbour for example the fjord walls are not steep sided with well-preserved glacial features compared with those described at some other sub-Antarctic Islands such as the

Fig. 9. (A) Satellite image of Campbell Island showing the approximately radial arrangement of the drainage system (black arrows) from the former centre of the shield volcano at Dent Island (triangle). This drainage system likely determined the orientation of the over-deepened glacial troughs including Perseverance Harbour and Northeast Harbour. (B) Satellite image of the Auckland Islands showing the approximately radial arrangement of the drainage system (white arrows) from the centres of the shield volcanoes at Carnley Harbour and Disappointment Island (triangles). The image clearly shows the deeply incised glacial troughs and hanging valleys on the east coast. The dashed white lines marking the former extent of the shield volcanoes are derived from Quilty (2007). Satellite images compiled by Google Earth (2013) from SIO, NOAA, U.S. Navy, NGA and GEBCO sources.
Auckland Islands and South Georgia. This raises the possibility that many of the glacial features described pre-date the LGM and that Quaternary glaciation has been both limited, and within the footprint of earlier glacial episodes. This interpretation is supported by the high number of endemic species on the island which excludes full LGM ice cover. Some researchers have even argued that the valley systems were formed by fluvial processes (Campbell, 1981), although most evidence supports at least partial glaciations during the Quaternary.

Following the last glaciation the earliest dates for peat accumulation range between 16,577–16,997 cal yr BP at Homestead Scarp and 14,132–15,024 cal yr BP at Mt Honey (McGlone et al., 2010). One coastal site at Hooker Cliffs has a minimum age for peat deposition of 14,845–16,629 cal yr BP, whilst peat accumulated at Rocky Bay from 13,352–13,767 cal yr BP (McGlone, 2002).

3.3.3. Auckland Island

The Auckland Island archipelago (50°50’S, 166°05’E), with a combined area of 625 km², is the largest of the New Zealand sub-Antarctic islands, situated to the northwest of Campbell Island, 465 km south-southeast of the South Island of New Zealand. The islands are not currently glaciated. They are entirely volcanic in origin, the emergent parts of the Campbell Plateau basement continental crust, and are composed of basaltic volcanics of Oligocene-Miocene age (Wright, 1967).

Similar to Campbell Island, former shield volcanoes, formed between 25 and 10 Ma, appear to have had a strong influence on the drainage pattern established on the Auckland Islands. Many of the fjords are radially arranged around inferred former ice domes centred on Carney Harbour in the South and Disappointment Island in the north (Fig. 9B); locations where Quilty (2007) identifies two former shield volcanoes (Fig. 9B). As at Campbell Island, evidence of glaciation is best preserved on the east coast with that on the western part of the island having been removed through marine erosion, or flank collapse (to the north) and marine erosion or explosive loss of the interior of the caldera (in the south). The presence of endemic plants and animals likely rules out complete glaciation during the LGM. However, there is extensive evidence of Quaternary glaciation, for example in the many deeply incised fjords, such as McLennan Inlet. This suggests that although the Auckland Islands are at a latitude 200 km north of Campbell Island the higher maximum altitude (705 m versus 569 m at Campbell Island) and larger land area has enabled extensive Quaternary ice accumulation.

Glacial features in the Auckland Islands were first described by Speight (1909). The eastern flank of the main Auckland Island has an impressive abundance of evidence of past glacial activity in the form of deeply cut wide U-shaped valleys with long coastal inlets and lateral moraines, hanging valleys, moraine-dammed lakes and cirques (Fig. 9B), and sub-terminal moraines (Speight, 1909); but there are currently no age constraints for these features. McGlone’s (2002) interpretation is that at the LGM all the major inlets in the east were glacier-filled, with cirques forming between 250 and 300 m in altitude (Wright, 1967). Fleming et al. (1976) and McGlone (2002) described till on Enderby Island (a small low-lying island close to the northeastern extremity of the mainland, Fig. 9B) which was deposited during the last glaciation by an extended glacier flowing from the uplands (400–460 m high) north-eastwards, filling Laurie Harbour. The till is separated into two members by laminated lake silts suggesting that two distinct glacial advances, possibly within the LGM, are recorded. The oldest Auckland Island radiocarbon date is 18,008–18,672 cal yr BP from a sandy layer with fine organics from the base of a c. 4 m thick blanket peat from the northern lowland slopes of the Hooker Hills. As this area was overrun by the Laurie Harbour palaeoglacier, it provides a minimum age for deglaciation (McGlone, 2002). Peat deposits have also been dated at Deas Head (13,496–14,031 cal yr BP) and Hooker Hills (12,590–12,926 cal yr BP) (McGlone et al., 2000).

3.3.4. Balleny, Scott, and Peter I islands

Balleny Island (66°55’S, 163°20’E, 400 km²) and Scott Island (67°24’S, 179°55’W, > 1 km²) are the sub-aerial expressions of a series of submarine ridges formed by volcanic activity on a timescale of ~10 Ma. No glacial geomorphological data are published, although Scott Island is largely glaciated today.

Peter I Island (68°50’S, 90°35’W, 154 km²) is the remnant of a former shield volcano formed 0.3–0.1 Ma and is heavily glaciated. No glacial geomorphological data are published. The well mapped bathymetry data around the island reveal that significant ice expansion is not possible due to steep flanks which fall away rapidly into the deep sea.

3.3.5. Diego Ramirez

The Diego Ramirez Islands (56°30’S, 68°42’W, c. 2 km²) are a group of small islands at the southernmost tip of Chile, formed during subduction of the continental crust. No glacial geomorphological data are published.

4. Discussion

Although many of the sub-Antarctic and maritime Antarctic Islands have been visited for several decades, this review demonstrates that few systematic studies of their glacial geomorphology and geochronology have been undertaken. As a result, the position of the LGM ice limits are not well defined, and in most cases there are no LGM age constraints, or constraints on the onset of deglaciation. Nevertheless, existing cosmogenic isotope dating studies on moraines and the basal ages of peat and lake deposits permit minimum ages for deglaciation to be inferred for some islands.

In terms of maximum ice volumes at the LGM, the sub-Antarctic islands can be divided into the following groups:

4.1. Type I) Islands which accumulated little or no LGM ice

These include the Falkland Islands and Macquarie Island. Situated north of the Antarctic Polar Front (Fig. 1) they are characterised by periglacial features with little evidence of extensive glaciations except for upland tarns and nivation hollows (Falkland Islands). This suggests either an insufficient moisture supply during glacial periods, insufficient altitude and relief to develop significant glaciers, or stronger Westerly Winds and more wind-driven ablation preventing glacier initiation. In these environments, glaciation was very limited and periglacial landscapes prevailed, for example the stone runs in the Falkland Islands (Wilson et al., 2008), and stone stripes and polygons on Macquarie Island (Selkirk et al., 1990). Where glaciers accumulated on the Falkland Islands they appear to have been restricted to eastern slopes, suggesting an important role for preferential snow accumulation on the lee side of ridges sheltered from the prevailing Westerly Winds. Elsewhere, there is evidence of wind erosion through the LGM where wind-blown sand grains carried up to heights of a metre above ground level have eroded the lower faces of exposed rock, forming distinct rock pillars in some parts of West Falkland such as the Port Stephens Formation (Aldiss and Edwards, 1989). On Macquarie Island, the moderating effect of the maritime climate and the relatively low altitude of the plateau (c. 150–300 m) would have also played a role in limiting snow accumulation (Selkirk et al., 1990).
4.2. Type II) Islands with a limited LGM ice extent but evidence of extensive earlier continental shelf glaciations

These islands include South Georgia and possibly Kerguelen, although for the latter data are still limited. Current chronological data suggests that the LGM ice extent at these locations was limited to the fjords despite there being glacial geomorphological evidence of earlier glaciations that extended across their continental shelves. This is of interest because both islands retain permanent ice caps today on account of their high altitude (up to 2934 m on South Georgia, and 1049 m on Kerguelen) and would have had substantially lower equilibrium lines during the last glacial. One hypothesis is that glacier extent was limited at the LGM because they were deprived of moisture by the more extensive sea ice (Bentley et al., 2007; Allen et al., 2011; Collins et al., 2012), and stronger Westerly Winds. This is a common feature of sub-Antarctic islands where the combination of more northerly sea ice and strong winds increased aridity; hence most peat and lake sequences only start to accumulate in the early to mid-Holocene (Van der Putten and Verbruggen, 2005; Van der Putten, 2008), with occasional exceptions dating from at or before the LGM (Rosqvist et al., 1999).

In addition to the geomorphological evidence, biological and molecular biological data confirm that the majority of the sub-Antarctic islands were not completely ice-covered at the LGM. This is because various elements of the flora and fauna have survived on the islands intact throughout the LGM and possibly earlier glaciations, resulting in the development of distinct floral provinces in the South Atlantic Ocean, South Pacific Ocean, and South Indian Ocean (Van der Putten et al., 2010). The evolution of endemic species also points to the long term persistence of glacial refugia. For example, highly divergent mitochondrial DNA lineages within the endemic weevil group Ectemnorhinus have been found within and among sub-Antarctic islands, most of them estimated to have existed since long before the LGM (Grobler et al., 2011). Similarly, evidence of biotic persistence on sub-Antarctic islands is found in mites (Mortimer et al., 2011) and flowering plants (Van der Putten et al., 2010; Wagstaft et al., 2011; Fraser et al., 2012), birds (McCracken et al., 2013) and in limpets on the continental shelf (González-Wevar et al., 2013), from at least the beginning of the Quaternary, with some genera such as Pleurophyllum possibly being the last remnants of a once-diverse Antarctic flora that dispersed northward in response to Neogene glacial advance (Wagstaft et al., 2011).

The differences in glacial history in the sub-Antarctic region appear to be a result of both latitudinal changes in climate and topographic control on the glacial equilibrium line altitude. For example, islands south of the Polar Front are generally colder,
accumulate glaciers and typically retain ice cover today because the glacial equilibrium line altitude is low. On these islands, the eustatic sea level fall during the LGM would have been sufficient to enable glaciers to expand, particularly where this opened up new exposures of shallow sea-floor to accumulation. On other islands such as Macquarie Island and the Falkland Islands topographic control appears to be more important. In these cases, their low mean altitudes meant that they have never accumulated significant ice masses. In contrast, the high mean altitudes of both South Georgia and Kerguelen have resulted in ice caps that have persisted to the present but experienced limited expansion at the LGM relative to earlier Pleistocene glaciations. This may be the result of the impact of the earlier glacially modified landscape on maximum LGM ice extent (see Anderson et al., 2012), or moisture deprivation resulting from more extensive sea ice (as described above); a feature seen along the Antarctic coast where relatively low winter precipitation and cloudiness occurs when the sea ice extent is greater (King and Turner, 1997). In the case of South Georgia, Bentley et al. (2007) note that the extent of sea ice in the northern Weddell Sea and central Scotia Sea is critical in determining the moisture content of depressions reaching the island. In addition to changes in sea ice extent, reduced moisture delivery is a product of a northward shift of the Southern Hemisphere Westerly Winds during the glacial; including the moisture supply from subtropical air masses ( Björck et al., 2012; Stager et al., 2012) and enhancing evaporation and sublimation rates. One simplified study with a general circulation model (Toggweiler et al., 2006) also suggests that the belt of the Southern Hemisphere Westerly Winds may move northward towards the Equator during cold periods (and vice versa). Other general circulation models have suggested no change in the latitudinal position of the Westerlies, but a general drying out at these latitudes (Rojas et al., 2009). Nevertheless, it seems likely that changing moisture supply was important in determining the mass balance of glaciers in the maritime and the sub-Antarctic regions (see discussion in Bentley et al., 2007), with altitude, temperature, insolation and aspect also being influential.

Although the sub-Antarctic islands' glaciers responded to different forcing at the LGM, and in particular have a regionally heterogeneous glaciation history that in some cases mirrors a South American pattern (see comments on Type II glacial histories) and others an Antarctic one (see comments on Elephant Island and the South Orkney Islands in the discussion of Type IV glacial histories), there is good evidence that those which have remaining ice cover are responding in the same way to the current warming trend. The majority of glaciers on these islands are showing evidence of recent retreat, which seems to have accelerated over the past three to five decades (e.g., Thost and Truffer, 2008; Berthier et al., 2009; Cook et al., 2010; Hall et al., 2011).

5. Conclusions

In the context of the ACE/PAIS community Antarctic Ice Sheet reconstruction (this Special Issue) the ice volume changes associated with the post-LGM deglaciation of the sub-Antarctic Islands are unlikely to have made a significant contribution to global sea level. However, being peripheral to the main Antarctic Ice Sheet, they are, and have been, very responsive to past climate changes and provide examples of later stages of deglaciation and the processes involved. For example, the deglaciation of the fjords of South Georgia in the early Holocene is remarkably similar to that occurring in the fjords of the western Antarctic Peninsula today. This early Holocene analogue serves as a useful gauge for determining the predictive accuracy of ice and climate models. Elsewhere the rapid recent deglaciation, and in some areas total loss of ice (e.g. Marion Island), provide examples of the final stages of deglaciation.

The lack of information on sub-Antarctic glaciation in this review highlights a need for future focus on the glacial history of the islands. Research priorities and future work should encompass:

- A greater emphasis on delimiting onshore and offshore limits of past glaciation, using glacial geomorphic, geophysical and sedimentary investigations and imaging and dating of subma-
  rine glacial features such as moraines and trough mouth fans.
- Targeted dating of glacial and postglacial sequences to increase understanding of the timing and pattern of post-LGM deglaciation.
- The use of volcanic markers to help constrain glacial history, given that many sub-Antarctic islands contain abundant lavas and tephras.
- Closer integration of ice-sheet modelling with climate and topographic forcing to reconstruct likely patterns of former glacial activity, especially where glacial geologic evidence is sparse or lacking.
- Glacier mass balance modelling, including sensitivity tests, to ascertain the key drivers of glacial change in the sub-polar belt.
- Examining patterns of Holocene glacier and ice-cap change in more detail to provide context to the widespread deglaciation occurring throughout the sub-Antarctic today.

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