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 approximately 2000–5000 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 the 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 terrestrial and submarine evidence of LGM (and/or earlier) ice expansion; (Type V) Islands north of the Antarctic Polar Front with terrestrial evidence of LGM ice expansion; and (Type VI) islands with no data. Finally, we review the climatological and geomorphological settings that separate the glaciological history of the islands within this classification scheme.

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 modellers 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 even 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 has been that 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 IJ05, has been widely adopted by the satellite gravity and GPS communities as the ice sheet reconstruction with which to underpin their 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 is not fully comprehensive 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 sector. 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 glaciations (Clarke et al., 2005; Barnes et al., 2006; Convey et al., 2008). This knowledge will help explain current 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 2000–5000 year time slices, our review found surprisingly few high quality age constrains on changing glacier extents on these timescales in the maritime and sub-Antarctic sector. Thus we limited ourselves to an assessment of the terrestrial and offshore evidence for the maximum LGM ice extent, and establishment of 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 for 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, Elephant Island and Clarence Island which are in the maritime Antarctic region (Fig. 1), whilst 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 islands cover an area of approximately c. 26,000 km^2, 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 ancient 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 has been described by Pendlebury and Barnes-Keoghan (2007).. However, these are based on measurements for a relatively short instrumental period at often protected stations close to current sea level. Based on these datasets, mean temperatures of the coolest months range from -5°C in the South Sandwich Islands to +11°C at Amsterdam Island. Mean temperatures of the warmest month range from +1°C in the South Sandwich Islands to +18°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 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 ice 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 most recent radiocarbon 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 glacigenic 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

Falkland Islands

The landscape of the Falkland Islands (51°45'S, 59°00'W, 12,173 km^2) is dominated by periglacial features. There is little evidence of LGM glacial ice apart from the small cirques and short (max. 2.7 km) glacially eroded valleys described by Clapperton (1971a) and Clapperon 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, would provide this data.

The absence of widespread LGM glaciation at altitude is supported by cosmogenic isotope (^{10}Be and ^{26}Al) 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 –27,000 yr BP, and also confirms that parts of the stone runs may have been in existence from before 54,000 yr BP and so may substantially pre-date 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 there at c. 17000 cal yr BP, and 16,573–16,950 cal yr BP respectively, presumably at a time of increased moisture supply (Table 1). Elsewhere the base of peat deposits has been dated to the late glacial / early Holocene, for example at 12,500 cal yr BP on Beauchêne Island (Lewis Smith and Clymo, 1984) and 9765–11,000 cal yr BP at Port Howard (Barrow, 1978). Studies of Quaternary environments (e.g., Clark et al., 1998; Wilson et al., 2002) have also provided no evidence of LGM glaciation beyond the cirques and small valley glaciers, and there are no studies, or bathymetric data that show evidence for LGM glaciers extending offshore.

Elephant Island and Clarence Island (maritime Antarctic)

Elephant Island (61°08′S, 55°07′W, 558 km^2) is a 47 x 27 km mountainous island at the northern limit of the South Shetland Islands (Fig. 1). 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, 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 1300m 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.

Whilst 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).

South Orkney Islands (maritime Antarctic)

The South Orkney Islands (60°35′S, 45°30′W, 620 km^2), an archipelago located 600 km north-east of the tip of the Antarctic Peninsula comprises four main islands: Coronation Island which rises to 1266 m, 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 at the LGM (Herron and Anderson, 1990; Bentley and Anderson, 1998). To constrain the age of this unconformity, piston cores and bottoms grabs were recovered from 35 locations. Only a handful of these cores penetrated glacier proximal/subglacial till but nevertheless confirmed that grounded ice reached to at least the 220 m isobath. Radiocarbon analyses of articulated pelecypod shells found within diatomaceous glacial marine sediment at South Orkney Plateau Site 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 began accumulating at another site on the plateau from 8348–8660 cal yr BP (Lee et al., 2010). Analyses of the ice rafted debris (IRD) assemblage in slope cores, composed exclusively 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 South Orkney Ice cap with the Antarctic Peninsula Ice Sheet at the LGM has also been suggested by Johnson and Andrews (1986) and by ice sheet models (Pollard and DeConto, 2009; Golledge et al., 2012). However this hypothesis is based on limited geological data and forced by a regional climatic model respectively, so 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 lake sedimentation which began at Signy Island between 7292–7517 cal yr BP (Sombre Lake) and 6484–6791 cal yr BP (Heywood Lake) (Jones et al., 2000). Moss banks accumulated from 4799–6183 cal yr BP (Fenton, 1982; Fenton and Smith, 1983) and 2784–3006 cal yr BP (Royles et al., 2012).

South Georgia

South Georgia (54°17'S, 36°30'W, 3755 km^2) 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 being Mt Paget (2934 m). Glacial geomorphological research on South Georgia is more advanced than most areas 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. 3B) 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 most recent of these glacial stages that extended across the continental shelf occurred during the LGM, although there remains a lack of chronological control on these periods of extensive glaciation (Clapperton, 1990). However, more recent evidence based on the submarine geomorphology of the coastal fjords (Hodgson et al., 0000), 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 Late Glacial to Holocene moraines (Clapperton, 1971b; Sugden and Clapperton, 1977; Clapperton and Sugden, 1988; Clapperton et al., 1989; Bentley et al., 2007) which have been correlated with the submarine glacial geomorphology in the fjords (Hodgson et al., 0000). This evidence is supported by minimum deglaciation ages derived from the onset of lake sedimentation and peat formation (Clapperton et al., 1989; Wasell, 1993; Rosqvist et al., 1999; Rosqvist and Schuber, 2003; Van der Putten et al., 2004; Van der Putten, 2008). The oldest cosmogenic isotope dates on South Georgia range between 14,084–10,574 yr BP (Table 2). These mark the oldest mapped ice advance, estimated using an error-weighted mean to have occurred at 12,107 ± 1373 yr BP (Bentley et al., 2007). Evidence of this ice advance (which corresponds to Bentley et al's 'category 'a' moraines') is seen at Husvik and the Greene Peninsula and can be correlated on geomorphological grounds with the oldest moraine ridges at Antarctic Bay, Possession Bay and Zenker Ridge. The clear offshore expression of these moraines can also be seen in the submarine glacial geomorphology, for example in Moraine Fjord as a bouldery shoal at low tide, and in Cumberland East Bay, where a pronounced inner basin loop moraine occupies the entrance to the fjord (Fig. 3B).

Lake sedimentation in one inner fjord location on Tonsberg Point commenced as early as 18,621–19,329 cal yr BP (Rosqvist et al., 1999) but in other areas basal lake sediment dates are early Holocene in age, for example Lake 10 on Tonsberg Point was deglaciated before 10,116–10,249 cal yr BP) (Van der Putten and Verbruggen, 2005), Fan Lake on Annenkov Island, situated off the south coast, was deglaciated before 7656–7839 cal yr BP and a lake adjacent to Prince Olav Harbour before 7788–7969 cal yr BP (Hodgson D.A. unpublished data) (Table 1). Glaciofluvial sediments were deposited at Husdal in Stromness Bay prior to 10,113–10,570 cal yr BP (Van der Putten et al., 2012) followed by the onset of peat formation. Elsewhere the earliest onset of peat formation ranges from 12,150–9650 cal yr BP and 11,600–10,550 cal yr BP at Gun Hut Valley (Barrow, 1978; Van der Putten and Verbruggen, 2005), 10,624–10,869 cal yr BP on Dartmouth Point (Smith, 1981), 10,512–10,893 cal yr BP on Tønsberg Point, 9009–9270 cal yr BP on Kanin Point (Van der Putten et al., 2009), 9495–9680 cal yr BP at Maiviken (Smith, 1981), and 7571–7690 cal yr BP and 7174–7418 cal yr B.P at Husdal (Van der Putten et al., 2013) (Table 1). 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 north east 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 the result of fluvio-deltaic deposition at higher relative sea levels such as the c. 9 m a.s.l. 'Line M' in Stromness Bay which marks 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 islands, there still remains a paucity of chronological control to constrain ice cap positions through the last deglaciation, particularly at ice-marginal positions offshore.

South Sandwich Islands

The South Sandwich Islands (56°20'S, 26°00'W to 59°20'S, 28°00'W, 618 km^2), 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). The islands are all glaciated but vary greatly in ice cover depending on 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 onshore, and there are no age constraints.

Bouvet Island

Bouvet Island or Bouvetøya (54°26'S, 3°25'E, 50 km^2) is located south of the Antarctic Convergence (Fig. 1). It consists of a single dominant active cone volcano (Fig. 4a). It is a 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 frontal variations of the order of 10–100 m (Mercer, 1967; Orheim, 1981). These were attributed to differences in aspect with regard to wind direction, as well as to local tidewater effects. The island consists of young oceanic crust, 4–5 Ma in age (Mitchell, 2003). Thus, 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 that do exist show a 3–4 km-wide shelf of <200 m water depth (Fig. 5A). Hence, even with extensive ice grounding onto the submarine shelf, we can be sure that any former glacial ice cap on Bouvet Island probably had an aerial extent no larger than ~330 km^2. Even with complete glacial 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/).

Gough Island

Gough Island (40°21'S, 9°55'W, 65 km^2) is a young (1Ma) 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 cored in the south-east of the island. They recovered an infinite radiocarbon age of >43,000 ^{14}C yr BP from the basal sediments, and argued for a continuity of occupation in flora through the last glacial-interglacial cycle on that basis. The well developed terraces around the coast (-50 m to 75 m asl) are also considered to be the result of eustatic sea level changes on glacial-interglacial timescales rather than evidence of Holocene glacioisostatic processes (Quilty, 2007).

3.2. Indian Ocean sector

Marion Island and Prince Edward Island

Marion Island (46°55'S, 37°45'E, 293 km^2) and Prince Edward Island (46°39'S, 37°57'E, 46 km^2) 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 five glacial episodes, have been inferred from K-Ar dating of striated outcrops, till, fluvioglacial deposits and glaciogenic 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) (McDougall et al., 2001). Thus, based on recent geomorphological evidence (Boelhouwers et al., 2008), an initial 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 late Quaternary glacial and LGM glacial geomorphology of Marion Island are summarised by Boelhouwers et al. (2008) and Hall et al. (2011). These studies all 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 document an isostatic rebound following deglaciation (Hall, 1977), or be the result of 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 detailed bathymetry data for the coastal margins of the island, analysis of the present day coastline from aerial photographs and QuickBird satellite imagery (Fig. 6B) suggests that the position and orientation of some of the outer kelp beds, which indicate the presence of shallower water, may be revealing either the presence of offshore terminal moraines from which the former position of glaciers could be inferred (Fig. 5B); similar to those seen at the entrance to Moraine Fjord, South Georgia (Fig. 3B). Alternatively, these features could be the termination of submarine lava flows. This could be confirmed by a programme of direct sampling and nearshore bathymetric survey.

Although the collective evidence suggests that glaciers extended beyond the coastline in many areas, phylogenetic studies of invertebrate communities (Chown and Froneman, 2008) and well-developed periglacial landforms, such as solifluction terraces and sorted patterned ground (e.g. Nel, 2001) show at least some inland areas remained exposed as nunataks during the last glacial period. For example, differences in phylogenetic substructure among populations of springtails (Myburgh et al., 2007), mites (Mortimer and van Vuuren, 2007; Mortimer et al., 2012) and the cushion plant *Azorella selago* (Mortimer et al., 2008) on the island are considered consistent with a hypothesis of within-island disjunction of populations by advancing glaciers, followed by population expansion from these refuges following glacial retreat (Fraser et al., 2012).

At present, there are few age constraints for the glacial features on Marion Island. The base of one 3 m peat sequence from Albatross Ridge has been inferred at c. 17,320 years BP (Van der Putten et al., 2010) based on extrapolation from a date of 10,374–11,000 cal yr BP (9500 ± 140 ^{14}C yr BP, Table 1) reported at 175–185 cm within a 3 m long peat profile (Schalke and van Zinderen Bakker, 1971). This suggests the onset of deglaciation could be as early as c. 17,320 years BP in this area. However, this extrapolated date has been disputed as it assumes a uniform sedimentation rate which is questionable where tephra deposits are reported (Gribnitz et al., 1986), and because elsewhere on Albatross Ridge peat core basal ages of only 6601–6950 cal yr BP (depth: 353–363 cm) and 4426–4744 cal yr BP (depth: 165–180cm) have been reported (Scott, 1985). On nearby Skua Ridge the oldest peat basal age is 7574–7873 cal yr BP and at Kildakey 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 (3316–3403 cal yr BP; Junior's Kop), 4020+/- 65 (4225–4587 cal yr BP; near the Marion Base Station), 2685+/-130 (2351–3005 cal yr BP; Nellie Humps Valley) (Schalke and van Zinderen Bakker, 1971) and 4750 +/- 40 (5316–5485 cal yr BP; near the Marion Base Station) (Yeloff et al., 2007), but as 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 to Marion Island at 1240 m he considered it unlikely that the island had entirely escaped glaciations. However, satellite imagery may provide data to resolve possible glacial features similar to the moraines and other glacial features found on Marion Island, but further analysis and ground truthing is required.

Crozet Islands

The Crozet Islands (46°25'S, 51°38'E, 400 km^2) 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 producing 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 of likely glacial origin on Île de la Possession (Vallée des Branloires, 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; Bougère, 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 (~2500 km^2) 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 Aflred 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.

The Kerguelen Islands

The Kerguelen Islands (48°30’S, 68°27’E and 50°S, 70°35’E) consist of a main island (7200 km^2) 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 km^2 (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). However 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 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 15,396–16,624 cal yr BP (Van der Putten et al., 2010) and at the Golfe du Morbihan from 12,765–13,241 and 9141–9912 (Young and Schofield, 1973a; Young and Schofield, 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 – 5188 cal yr BP (Table 1, sample number 4, Fig. 7C), and 2208–716 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 ±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; Frenot et al., 1997a; Arnaud et al., 2009).

Recent glacier retreat has been documented from the first half of the 20th century (Aubert de la Rile, 1967; Vallon, 1977) and the total ice extent on Kerguelen Islands declined from 703 to 552 km^2 between 1963 and 2001, with the Cook Ice Cap retreating from 501 to 403 km^2 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; Frenot et al., 1997a; Jacka et al., 2004), and decreased precipitation since AD 1960 (e.g., Frenot et al., 1993; Frenot et al., 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).

Heard Island and McDonald Island

Heard and McDonald Islands (located at approximately 53°06'S, 73°31'E) are 380 km^2 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 km^2 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 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 along 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 continue to reach sea level today, and offshore on the continental shelf there is evidence in a bathymetric grid compilation (Beaman and O'Brien, 2011) of an 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. 8), but the age of these features remains unknown. The position and depth of 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 bathymetry data for the sea-floor plateau surrounding Heard Island now exist at a resolution that permits a closer analysis of these submerged glacial features (~100 m grid cell size; Fig. 8). 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 clear, 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. 8) and suggest that the ice cap was organised into several discrete faster-flowing outlets, in common with most examples of ice caps and ice sheets today.

Sketches of more extensive glaciers by visiting sealers in the 1850s to 1870s and photographic evidence documents glacial retreat over the latter half of the 20th Century (Kiernan and McConnell, 1999; Kiernan and McConnell, 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 ±113 ^{14}C 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 (McIvor, 2007). There is no published information on its glacial history.

Amsterdam and St Paul Islands

Situated between South Africa and Australia, Amsterdam Island (37°50'S, 77°30'E, 55 km^2) and Saint-Paul Island (38°43'S, 77°31'E, 6 km^2) are volcanic islands dating from about 400–200 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

Macquarie Island

Macquarie Island (54°37'S; 158°54'E, 200 km^2) 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. It is composed of ocean-floor rocks belonging 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 palaeobeaches, terraces and cobbles seen around the island are not interpreted as being the result of glacioisostatic uplift, but of marine erosion during the geological uplift of the island over the last the 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, 103 m asl) attributed to Marine Isotope Stage 9 (340–330 ka) and Stage 5e (130–125 ka), respectively, which although the TL errors are very large, 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 4 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 overlie 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, after which radiocarbon evidence shows that the 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 palaeo Lake Skua (Selkirk et al., 1991) and 11,284–12,581 cal yr BP at the Finch Creek Ridge peat deposit (Selkirk et al., 1988; Keenan, 1995) (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 form perglacial conditions as basal ages have yet to be determined for some of the lakes, such as Palaeolake Toutcher (Selkirk et al., 1988). Younger palaeolake deposits of are found at 8185–8639 cal yr BP, at Palaeolake Sandell and peat deposits at 7682–8203, 5986–7476 and 6206–7272 cal yr BP at Green Gorge Ridge, Wireless Hill and Finch Creek ridge (Selkirk et al., 1982) (Table 1).

Since the tectonic uplift of the island the geomorphological evidence therefore suggests extensive periglacial rather than glacial activity has occurred on the cold uplands of Macquarie Island. This has resulted in the formation 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 Selkirik, 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.

Campbell Island

Campbell Island (52°33'S, 166°35'E, 120 km^2) 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. The late Quaternary glacial history of the island has been summarised by McGlone (2002).

There are no detailed studies on the glacial history of Campbell Island. However, a 'corrie and moraine' was described as early as 1896, most likely on Mt Honey (Marshall, 1909) and there is evidence of many geomorphological features associated with glacial U-shaped valleys (Marshall, 1909; Quilty, 2007). Early soundings from Perseverance Harbour suggest that it has been over deepened by ice derived from glaciers at Mount Honey and Mount Lyall and a valley east of Mount Honey has been 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). The absence of soils between the bedrock and overlying peat deposits has been interpreted as a result of 'severe climatic conditions, which gave rise to relatively large glaciers' (Marshall, 1909). Studies by McGlone (1997; 2002) 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 2–3m 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. These glaciers retreated and the earliest peat soils began forming 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). Deglaciation may have been rapid as one coastal site at Hooker Cliffs has a minimum age for deglaciation of 14,845–16,629 cal yr BP, whilst Rocky Bay was deglaciated later, between 13,352–13,767 cal yr BP (McGlone, 2002).

Auckland Island

The Auckland Island archipelago (50°50'S, 166°05'E), with a combined area of 625 km^2, 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 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). 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. 9), and submarine 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) which was deposited during the last glaciation by an extended glacier flowing from the uplands (400 to 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 18,009–18,672 cal yr BP from a sandy layer with fine organics from the base of a c. 4m 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).

Balleny, Scott, and Peter I Islands

Balleny Island (66°55'S, 163°20'E, 400 km^2) and Scott Island (67°24'S, 179°55'W, > 1 km^2) are the subaerial 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 km^2) 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.

Diego Ramirez

The Diego Ramirez Islands (56°30'S, 68°42'W, c. 2 km^2) 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 determinations of 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:

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, 1999). 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).

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 the 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 this group 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). Patagonian climate, east of the Andes was also more arid at this time (Recasens et al., 2011) due, in part, to the same factors, combined with the rain shadow effect of the mountains. These islands may therefore have followed a glacial history more similar to that of central Patagonia (46°S), the closest continental landmass at these latitudes, where a series of Pleistocene glaciations (of Marine Isotope Stage 20 and younger) extended beyond LGM limits (Singer et al., 2004) with the most extensive glacial advance occurring at c. 1.1 Ma (Rabassa, 2000), although the pattern of South American glaciation may be rooted in other drivers, such as glacial erosion (Kaplan et al., 2009), in addition to climate processes. An alternative hypothesis is that over many glacial cycles, the glacial erosion of the alpine valleys and fjords has been sufficient to reduce the length of glaciers in the most recent cycle because theoretically glacier length can scale linearly with erosion depth (Anderson et al. 2012). In such cases there are often earlier moraines deposited well beyond the LGM limit, referred to by Anderson et al. (2012) as a 'far-flung' moraine. This suggests that the glacially modified landscape, rather than a different climate, may be capable of explaining the earlier more extensive glacier extents.

In either case, this glacial history contrasts with much of the Antarctic continent, including the Antarctic Peninsula, where the LGM glaciation was amongst the most extensive in the Quaternary.

Type III) Seamounts and volcanoes unlikely to have accumulated significant LGM ice cover

These islands can be divided into two sub-groups. First those which are situated south of the Antarctic Polar Front including the South Sandwich Islands, Clarence Island and Peter I Island which are unlikely to have accumulated significant expansion of ice due to steep flanks which fall away rapidly into the deep sea. Second, islands to the north of the Antarctic Polar Front, including Amsterdam and St Paul Islands and Gough Island. These have no evidence of glaciation, low mean altitudes and also have steep flanks which fall away rapidly into the deep sea.

Type IV) Islands on shallow shelves with both terrestrial and submarine evidence of LGM (and/or earlier) ice expansion.

These include volcanic islands such as Heard Island, Bouvet Island, Marion Island, Prince Edward Island and Crozet Island which are located on top of extensive submarine plateaux, and non-volcanic islands including the South Orkney Islands and Elephant Island which are located on the South Scotia Ridge and surrounded by shallow shelves. On some of the volcanic islands, such as Heard Island and possibly Marion Island, there is geomorphological evidence that the glaciers extended onto the adjacent shelf; and on Heard Island, perhaps as far as the shelf break in some areas. This expansion would have been facilitated by the glacial eustatic sea level fall. Glaciation of these volcanic islands may have been initiated by a northward shift of the Antarctic Polar Front during the last glacial resulting in cooler temperatures and increased precipitation as snow. Loss of ice by calving of tidewater glaciers may have also been diminished as a result of the expansion of Antarctic sea ice which would have acted to reduce wave energy (Balco, 2007). At the South Orkney Islands there is very good evidence that grounded ice reached to at least the 220 m isobath, whilst on the Elephant Island archipelago the presence of a large shallow continental shelf also shows that a large area for ice accumulation was exposed during glacial low stands.

Type V) Islands north of the Antarctic Polar Front with terrestrial evidence of LGM ice expansion

These islands include Campbell and Auckland Islands both of which have terrestrial geomorphological evidence of extensive glaciations through the LGM and minimum ages for post-LGM ice retreat based on the onset of peat accumulation.

Type VI) Islands with no data

Balleny Island, Scott Island and Diego Ramirez have no published glacial history that we are aware of.

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; Wagstaff et al., 2011; Fraser et al., 2012), birds (McCracken et al. 2013) and in limpets on the continental shelf (González-Wevar et al., in press), 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 (Wagstaff 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 Pleistoccene 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 that they were deprived of moisture by 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; reducing 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 an important influence on 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 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 submarine 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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Tables

Table 1.

Selected radiocarbon ages of peat and lake sediment deposits on the sub-Antarctic islands that are considered to provide reliable minimum age constraints for deglaciation. Calibration of radiocarbon dates were 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: [a] extrapolated age; ** see stratigraphic comment in Selkirk et al 1998; [R] Age rejected by the original authors; [V] represents an unreliable minimum age for deglaciation as accumulation of sediments follows a volcanic event; [R1] calibrated using the Marine 09 data set with a Delta R of 948 (based on a local reservoir correction of 1348 minus the global marine reservoir of 400), the small size of this sample, taken over 5cm, means that the age from core 85–23 is 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 2.

Selected cosmogenic isotope exposure ages that can be used to provide constraints on glaciation the Falkland Islands (Prince's Street stone runs) and South Georgia (Husvik and Greene Peninsula). Cosmogenic ages were recalculated from Wilson et al (2008) and Bentley et al (2007) using the latest version of the CRONUS online calculator (Balco et al, 2008) (Wrapper script: 2.2; Main calculator: 2.1; Constants: 2.2.1; Muons: 1.1). We used a standard atmosphere flag for all samples, and South Georgia samples have an assumed density of 2.5 g.cm-3. We calculate mean and weighted mean ages of the samples along single moraines at Husvik and Greene Peninsula.. [a]Model exposure age assuming no inheritance, zero erosion, density 2.65 g/cm3, and standard atmosphere using a constant production rate model and scaling scheme for spallation of Lal (1991) / Stone (2000). This version of the CRONUS calculator uses a reference spallogenic ^{10}Be production rate of 4.49 ± atoms g-1 yr^{-1} (± 1σ, SLHL) and muonogenic production after Heisinger et al. (2002a; 2002b). The quoted uncertainty is the 1σ internal error, of which 0.39 reflects measurement uncertainty only.

Caption text you need as follows.

Figures

Figure 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).

Figure 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 infill.

Figure 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 in Moraine Fjord (from Bentley et al., (2007), together with shipborne swath bathymetry data presented in Hodgson et al (0000), illustrating fjord-mouth (‘inner basin’) moraines of presumed similar age. Bathymetry is shown at 5-m grid cell size.

Figure 4. Zavodovski Island, one of the South Sandwich Islands showing that the steep submerged slopes that flank these volcanic islands limit the potential for ice expansion offshore. The submarine geomorphology is dominated by features related to slope instability and volcanism and no distinct glacial features have been identified (Leat et al., 2010).

Figure 5. Regional bathymetry around selected volcanic islands: Bouvet Island (A), South Atlantic, and the Crozet Islands (C), southern Indian Ocean, both drawn from the Global Multi-Resolution Topography (GMRT) synthesis (Ryan et al., 2009). Contours at -100 m and -200 m water depths illustrate shallow plateaus around Bouvet, as well as several of the islands of the Crozet archipelago. Aerial photograph in Figure 4B shows modern glacial cover on Bouvet Island, looking West from an altitude of 361 km; taken from the Image Science and Analysis Laboratory, NASA-Johnson Space Center. "The Gateway to Astronaut Photography of Earth." http://eol.jsc.nasa.gov/scripts/sseop/photo.pl?mission=ISS017&roll=E&frame=16161 last accessed 12/04/2012 12:48:51.

Figure 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 latero-frontal moraines from which former glacier positions can be inferred, or the termination of submarine lava flows.

Figure 7. (A) Location of the Kerguelen Islands. (B) Location of glaciological investigations at the Ampère Glacier and the Gentil Glacier. (C) The Baie d'Ampère showing the location of the 9 radiocarbon dated peat deposits listed in Table 1, and more recent moraines post AD 1700. (D) The Gentil Glacier frontal and lateral moraines at the base of Mont Ross that predate AD 934 ±46 (1016 cal yr BP) based on the absence of a diagnostic ash layer from the Allouarn Volcano (Arnaud et al., 2009)

Figure 8. Regional bathymetric grid of Heard Island showing well-developed cross-shelf troughs and moraines extending as much as 50–80 km from the present shoreline. Data drawn from compilation by Beaman and O'Brien (2011).

Figure 9. Satellite image of Auckland Island, highlighting well-developed glacial troughs and hanging valleys. From the Image Science and Analysis Laboratory, NASA-Johnson Space Center. "The Gateway to Astronaut Photography of Earth." http://eol.jsc.nasa.gov/scripts/sseop/photo.pl?mission=STS089&roll=743&frame=5; last accessed 12/19/2012 16:25:42.

Table 1

Site name	Sample ID	Latitude	Longitude	Elevation (m a.s.l.)	Material dated / Stratigrpahic depth	Reported ^{14}C age	Calibrated age range 2 sigma	Source publication
Falkland Islands								
Plaza Creek	SRR-3906	51°23'18''S	58°29'20''W	<5	Peat	35970 ± 280	40521 - 41705*	Clark et al., 1998
Hooker Point	-	51°42'00''S	57°46'49''W	0	Peat	-	c. 17000	Long et al., 2005
Lake Sullivan	SRR-3898	51°49'57''S	60°11'27''W	-	Peat	13610 ± 45	16573 - 16950*	Wilson et al., 2002
Beauchene Island	-	52°54'00''S	59°11'00''W	-	Peat	-	c. 12500	Lewis-Smith and Clymo, 1984
Port Howard, Site 9						9280 ± 260	9765 - [11000*]	Barrow, 1978
Elephant and Clarence Islands								
Walker Point, Elephant Island	LU-2952	61°08’35’’S	54°42’ 01’W	200-220	Moss peat	5350±60	5927 - 6211	Bjorck et al 1991
South Orkney Islands								
S. Orkney Plateau, Site PC85-23		60°49.10''S	45°44.70''W	304(-)	Marine pelecypods, 264-269 cm	11,535±900	9442 - 13848[R1]	Herron and Anderson, 1990
S. Orkney Plateau, Site PC85-23		60°49.10''S	45°44.70''W	304(-)	Marine pelecypods, 83.5-86 cm	9570±2180	4177 - 15099[R1]	Herron and Anderson, 1990
S. Orkney Plateau, GC02-SOI03	NZA18576	60°22'S	47°00'W	786(-)	Marine sediment, 502 cm	10542±70	8348 - 8660[R2]	Lee et al., 2010
Sombre Lake, Signy Island	AA-10691	60°41'12''S	45°37'00''W	5	Lake sediment, 250-252 cm	6570±60	7292 - 7517	Jones et al., 2000
Heywood Lake, Signy Island	AA-10704	60°41'24''S	45°36'31''W	4	Moss fragment, 238-240 cm	5890±60	6484 - 6791	Jones et al., 2000
Site 'C', Signy Island	SRR-1089				Moss bank, 125 cm	4801±300	4799 - 6183	Fenton and Smth, 1982; Fenton, 1982
Site 'D', Moss BraseSigny Island	BETA-281618	60°68'S	45°62'W	112	Moss bank, 178 cm	2860±40	2784 - 3006	Royles et al., 2012
South Georgia								
Tønsberg Point, Lake 1	UA-2991	54°10'02''S	36°41'30''W	-	Lake sediment, 499 cm	15715 ± 150	18621 - 19329*	Rosvqist et al., 1999
Gun Hut Valley', Site 4	SRR-736				Peat, 258 cm	9493 ± 370	9650-12150	Barrow, 1978
Gun Hut Valley'	SRR-1979				Peat, 350 cm	9700 ± 50	10550-11600	in Van der Putten and Verbruggen, 2005
Tønsberg Point, Tønsberg sequence	UtC-4179	54°10'S	36°39'W	-	Peat, 308 cm	9520 ± 80	10512-10893	Van der Putten et al., 2004
Dartmouth Point	SRR-1165	54°19''S	36°26'W	-	Peat	9433 ± 120	10264 - 10869	Smith, 1981
Husdal, Sink Hole sequence	UtC-3307	54°11'24" S	36° 42' 12" W	-	Peat, 460 cm	9160 ± 110	10113 - 10570	Van der Putten et al., 2012
Tønsberg Point Lake 10	UtC-6232	54°10'09''S	36°39'54''W	-	Lake sediment, 447 cm	9060 ± 50	10116-10249	Van der Putten and Verbruggen, 2005
Grytrviken	SRR-1168			-	Peat, 460 cm	8737 ± 50	9536 - 9795	Smith, 1981
Maiviken	SRR-1162	54°15''S	36°29'W	-	Peat, 180 cm	8657 ± 45	9495-9680	Smith, 1981
Gun Hut Valley', Site 3					Peat, 160 cm	8537 ± 65	9396 - 9553	Barrow, 1978
Husvik Harbour, Kanin Point	UtC-6866	54°11'09''S	36°41'44''W	-	Peat, 312 cm	8225 ± 45	9009 - 9270	Van der Putten et al., 2009
Black Head bog	Beta-271303	54°04'07''S	37°08'41''W	43	Peat, 373 cm	8110±50	8723 - 9123	Hodgson, D.A. (unpublished data)
Prince Olav Harbour Lake 1	Beta-271300	54°04'24''S	37°08'08''W	335	Lake sediment, 197 cm	7110 ± 40	7788 - 7969	Hodgson, D.A. (unpublished data)
Fan Lake, Annenkov Island	SUERC-12584	54°29'55''S	37°03'03''W	90	Lake sediment, 584 cm	6953 ± 37	7656-7839	Hodgson, D.A. (unpublished data)
Husdal River site	UtC-6869	54°11'51'' S	36°42'12'' W	-	Peat, 300 cm	6840 ± 40	7571–7690	Van der Putten et al., 2013
Husdal	UtC-6867	54°11'63'' S	63°42'92'' W	-	Peat, 290 cm	6415 ± 40	7174–7418	Van der Putten et al., 2013
Gough Island								
					Peat	>43,000		Bennett et al., (1989)
Marion Island								
Macaroni Bay - extrapolated age	-			50	Peat, 300 cm	-	c. 17320[a]	Van der Putten et al., 2010
Macaroni Bay	K-1064			50	Peat, 175-185 cm	9500 ± 140	10374-[11000*]	Schalke & van Zinderen Bakker, 1971
Macaroni Bay	I-2278			50	Peat, 275-295 cm	10600 ± 700	10371-13841*	Schalke & van Zinderen Bakker, 1971[R]
Kildakey Bay peat section	Pta-3208				Peat, 600 cm	7300 ± 70	7934-8198	Scott, 1985
Skua Ridge, First boring	Pta-3214				Peat, 130-140 cm	6930 ± 90	7574-7873	Scott, 1985
Albatross Lakes, Third boring	Pta-3232				Peat, 353-363 cm	5990 ± 70	6601-6950	Scott, 1985
Albatross Lakes, Fourth boring	Pta-3231				Peat, 165-180 cm	4140 ± 70	4426-4744	Scott, 1985
Crozet - Ile de la Possession								
Base A. Faure, Baie du Marin	KIA-19231	46°25'49''S	51°51'31''E	110	Peat, 402 cm	9655 ± 60	10750 - [11000*][V]	Van der Putten et al., 2010
Morne Rouge Volcano flank	KIA-31355	46°23'35.45''S	51°48'28.85''E	12	Peat, 197 cm	6110 ± 40	6779 - 7020[V]	Ooms et al., 2011
Morne Rouge lake core	NZA-11510	46°23'26''S	51°48'45''E	50	Lake sediment, 405 cm	5750 ± 60	6389 - 6640[V]	Van der Putten et al., 2008
Morne Rouge peat sequence	NZA-11509	46°23'26''S	51°48'45''E	50	Peat, 532 cm	5480 ± 60	6000 - 6316[V]	Van der Putten et al., 2008
Kerguelen Islands								
Estacade	SacA 7753	49°16' 03''S	70°32'29''E	7	Peat, 468 cm	13190 ± 50	15396 - 16624*	Van der Putten et al., 2010
Golfe du Moribihan, Core 2					Peat 525 cm	11010 ± 160	12765 - 13241*	Young and Schofield, 1972a,b
Ampère Glacier	2	49°23'50'S	69°10'14''E	265	Peat, sample 2	10120 ± 90	11336 - 12054*	Frenot et al., 1997a
Ampère Glacier	1	49°23'50'S	69°10'14''E	260	Peat, sample 1	10140 ± 120	11264 - 12151*	Frenot et al., 1997a
Ampère Glacier	3	49°23'42''S	69°10'55''E	280	Peat, sample 3	9930 ± 70	11212 - 11629*	Frenot et al., 1997a
Golfe du Moribihan, Core 1					Peat, 260 cm	8595 ± 125	9141-9912	Young and Schofield, 1972a,b
Ampère Glacier	4	49°23' 47''S	69°09'55''E	240	Peat, sample 4	4590± 60	5054 - 5188	Frenot et al., 1997a
Ampère Glacier	5	49°24' 15''S	69°10'23''E	30	Peat, sample 5	2220±80	2098 - 2208	Frenot et al., 1997a
Ampère Glacier	6	49°24' 15''S	69°10'23''E	30	Peat, sample 6	1960±80	1732 - 1928	Frenot et al., 1997a
Ampère Glacier	7	49°23' 47''S	69°09'55''E	240	Peat, sample 7	1670±50	1384 - 1621	Frenot et al., 1997a
Ampère Glacier	8	49°24' 15''S	69°10'23''E	30	Peat, , sample 8	1320±70	1166 - 1282	Frenot et al., 1997a
Ampère Glacier	9	49°24' 17''S	69°10'23''E	160	Peat, sample 9	900±70	716 - 804	Frenot et al., 1997a
Heard Island								
Deacock Glacier moraine Long beach	Wk 9485			4.2	Subfossil sedge, 250 cm	220 ±113	modern - 340	Kiernan and McConnell, 2008
Macquarie Island								
West Mt Eitel	SUA 3045	54°35'S	158°51'E	-	Freshwater diatom peat	carbon dead	> 40 000	Adamson et al., 1996
West Mt Eitel	Beta-57317	54°35'S	158°51'E	-	Freshwater diatom peat	carbon dead	> 40 000	Adamson et al., 1996
Emerald Lake	NZA 50632	54°40'22''S	158°52'14''E	170	Lake sediment, 90 cm	13659 ± 56	16620 - 16987*	Saunders, K (unpublished data)
Palaeo Lake Skua	SUA 2736	54°37'S	158°50'E	180	Lake sediment, 1360 cm	13570 ± 150	15975 - 17034*	Selkirk et al., 1991
Palaeo Lake Skua	Beta-20165	54°37'S	158°50'E	180	Lake sediment, 900 cm	12470 ± 140	14063 - 15119*	Selkirk et al., 1988
Palaeolake Toutcher	Beta-20162			200	Lake sediment	11010 ± 200	12579 - 13276*	Selkirk et al., 1988
Finch Creek Ridge	Beta-1386	54°34'S	158°54'E	100	Peat	10275 ± 230	11284 - 12581*	Selkirk et al. 1988**
Palaeolake Nuggets	SUA-1894			30	Lake sediment, 450 cm	9400 ± 220	10146 - [11000*]	Selkirk et al., 1988
Palaeolake Sandell	Beta-20163			210	Lake sediment, >420 cm	7960 ± 110	8185 - 8639	Selkirk et al., 1988
Green Gorge Ridge	SUA-1461	54° 38'S,	158°54'E	100	Peat, 130 cm	7200 ± 130	7682 - 8203	Selkirk et al., 1988
Wireless Hill	Beta-1387			100	Sandy peat 360 cm	5960 ± 360	5986 - 7476	Selkirk et al., 1982
Finch Creek	SUA-1845X	54°34'S,	158° 55'E	100	Peat, 190 cm	5930 ± 240	6206 - 7272	Selkirk et al., 1988
Campbell Island								
Homestead Scarp	Wk-19746	52°33'S	169°08'E	30	Peat	13648 ± 73	16577 - 16997*	McGlone et al., 2010
Hooker Cliffs	NZ 6898	52°28'S	169°11'E	60	Peat	12950 ± 200	14845 - 16629*	McGlone, 2002
Mt Honey	Wk-13466	52°34'S	169°08'E	120	Peat	12445 ± 76	14132 - 15024*	McGlone et al., 2010
Rocky Bay	NZ 6984	52°33'S	169°04'E	130	Peat	11700 ± 90	13352 - 13767*	McGlone, 2002
Auckland Island								
McCormick Peninsula	NZA 4509	50°32'S	166°13'E	25	Peat	15170 ± 140	18009 - 18672*	McGlone, 2002
Deas Head	NZA 4607	50°32'S	166°13'E	20	Peat	11951 ± 95	13496 - 14031*	McGlone et al., 2000
Hooker Hills	NZA 9293	50°33'S	166°10'E	275	Peat	10859 ± 77	12590 - 12926*	McGlone et al., 2000

Table 2

Site name	Sample ID	Lat. (°S)	Long. (°W)	Elevation (m a.s.l.)	Elevation flag	Thickness of sample (cm)	Density (g cm^{-1}) Measured (italics), or assumed (normal)	Topographic shielding	Erosion rate	Isotope concentration ^{10}Be (at/g^{-1})	±^{10}Be (at/g^{-1})	^{10}Be Standard	^{26}Al (at/g^{-1})	±^{26}Al (at/g^{-1})	^{26}Al Standard	Exposure age^{a} (yr)	internal uncertainty (yr)	external uncertainty (yr)	Mean age (yr)	error (yr)	Weighted mean age (yr)	error (yr)
Falkland Islands																						
Prince's Street - stone runs (max)	PS/VAB-03	-51.61	-58.09	101	std	4.4	2.71	0.997	0	3.72E+06	1.48E+05	NIST_27900	1.87E+07	8.40E+05	Z92-0222	**827366**	40765	97892				
Prince's Street - stone runs (min)	PS/HSB-04	-51.62	-58.09	152	std	4.3	2.56	0.995	0	2.64E+05	1.30E+04	NIST_27900	1.81E+06	8.00E+04	Z92-0222	**46275**	2305	4674				
South Georgia																						
Husvik	HUS1	-54.1814	-36.7199	65	std	5	2.50	0.98	0	6.81E+04	1.00E+04	S555	0.00E+00	0.00E+00		**11506**	1695	1969				
Husvik	HUS2	-54.1814	-36.7191	65	std	5	2.50	0.98	0	8.33E+04	1.58E+04	S555	0.00E+00	0.00E+00		**14084**	2681	2948	12055	5769	12107	1373
Husvik	HUS4	-54.1814	-36.7175	49	std	5	2.50	0.98	0	6.16E+04	2.80E+04	S555	0.00E+00	0.00E+00		**10574**	4819	4906				
Greene Peninsula	GRE5	-54.3205	-36.4235	15	std	5	2.50	1	0	7.36E+04	4.45E+04	S555	0.00E+00	0.00E+00		**12811**	7771	7851				
Greene Peninsula	GRE1	-54.337	-36.6019	162	std	5	2.50	0.99	0	2.59E+04	1.41E+04	S555	0.00E+00	0.00E+00		**3925**	2139	2166				
Greene Peninsula	GRE2	-54.3349	-36.4525	150	std	5	2.50	0.99	0	2.23E+04	1.62E+04	S555	0.00E+00	0.00E+00		**3419**	2486	2503				
Greene Peninsula	GRE3	-54.3368	-36.4519	155	std	5	2.50	0.99	0	2.23E+04	1.67E+04	S555	0.00E+00	0.00E+00		**3402**	2550	2567	3521	4512	3515	1080
Greene Peninsula	GRE4	-54.3335	-36.453	160	std	5	2.50	1	0	2.22E+04	1.17E+04	S555	0.00E+00	0.00E+00		**3338**	1761	1784				

Figure 1

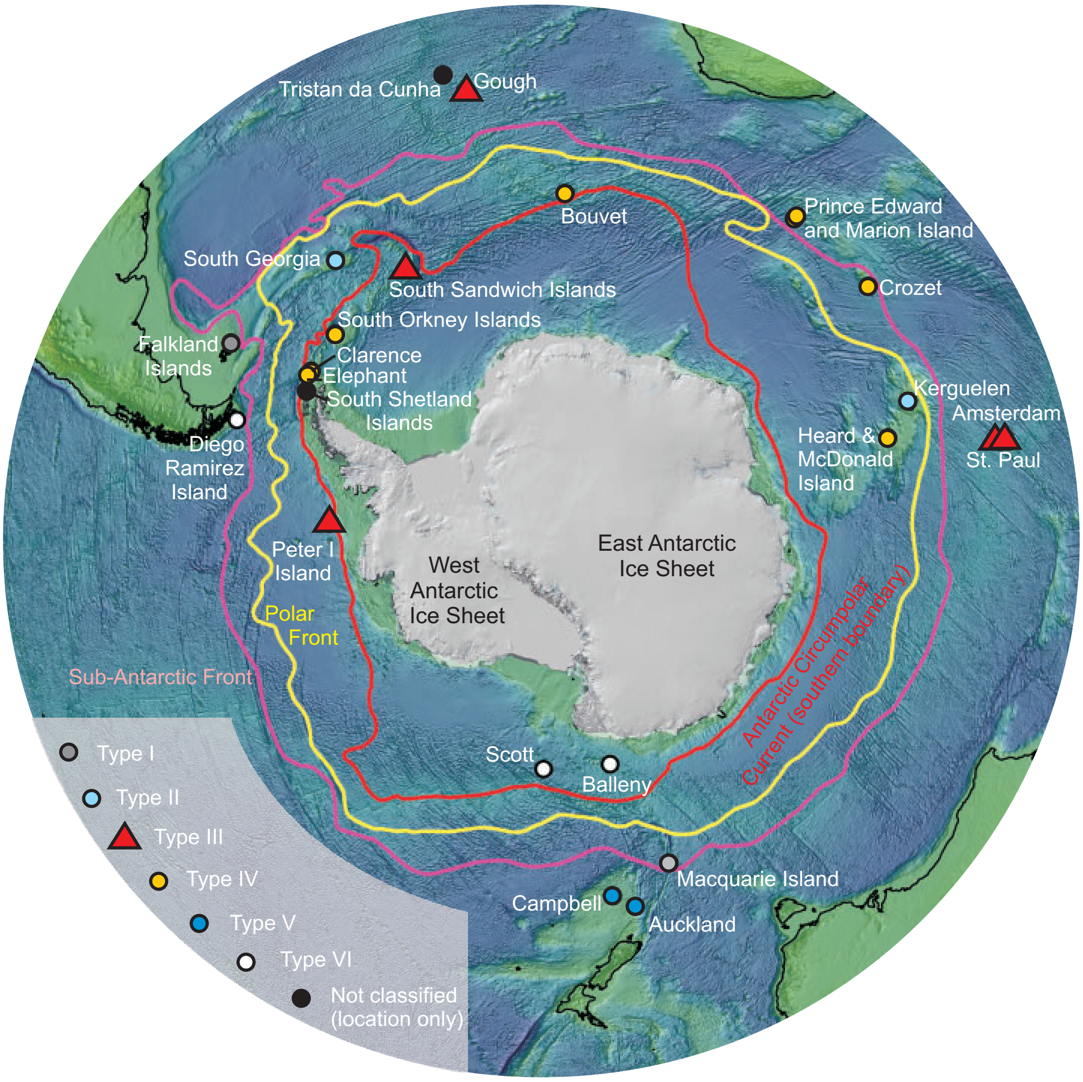

Figure 1, Hodgson et al.

Figure 2

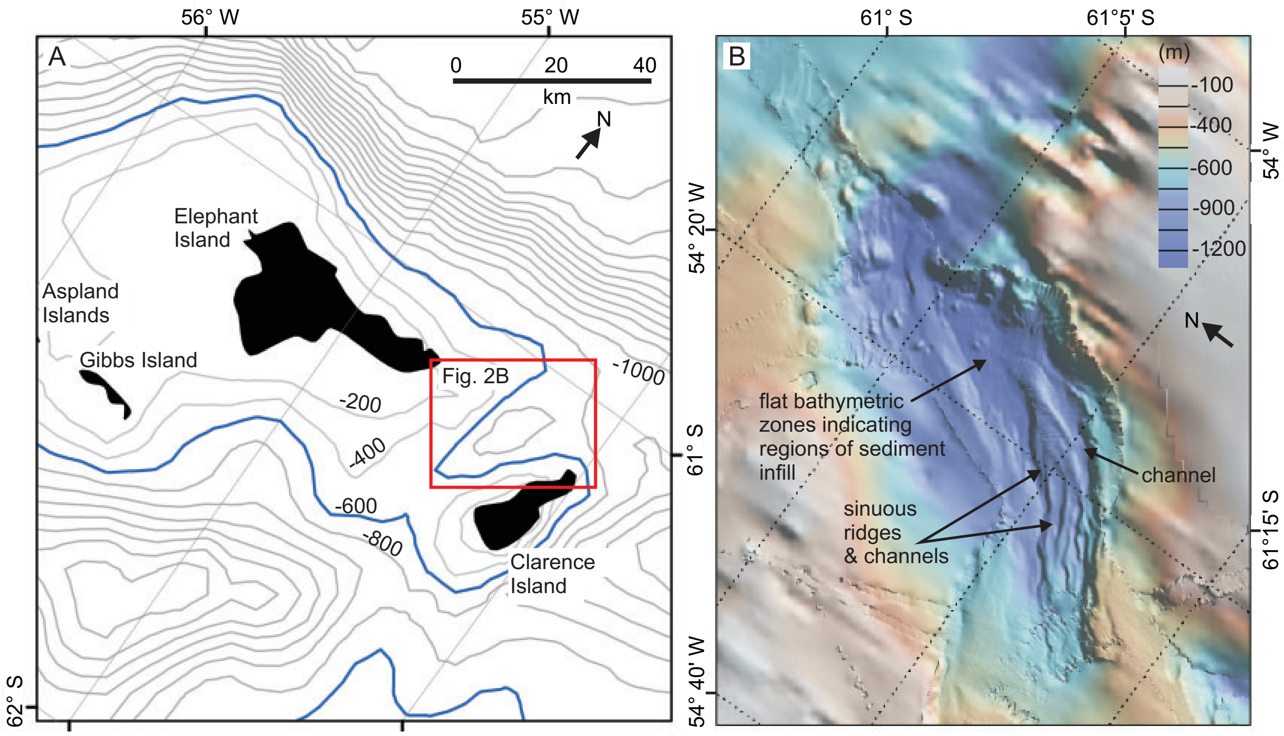

Figure 2, Hodgson et al.

Figure 3

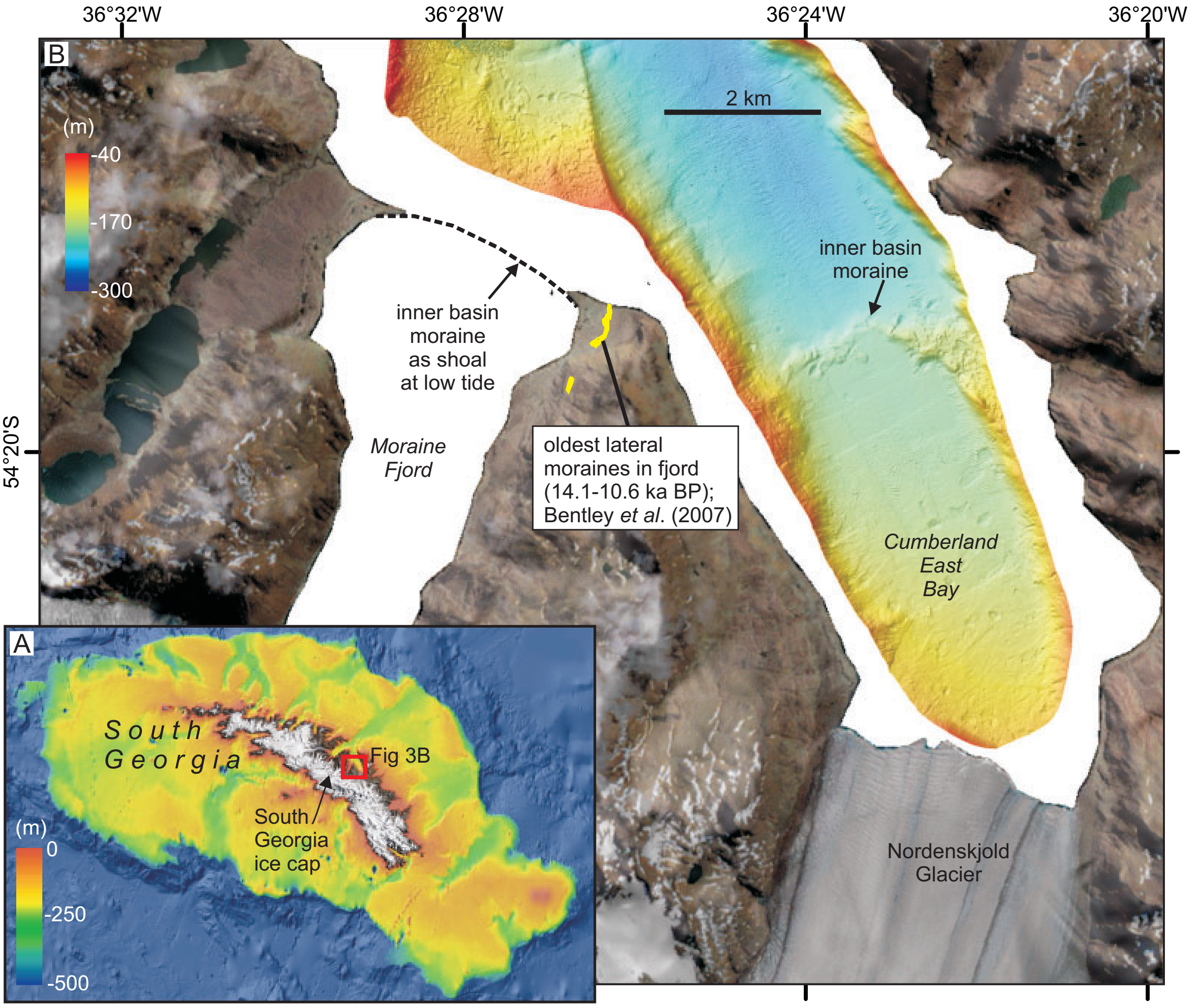

Figure 3, Hodgson et al.

Figure 4

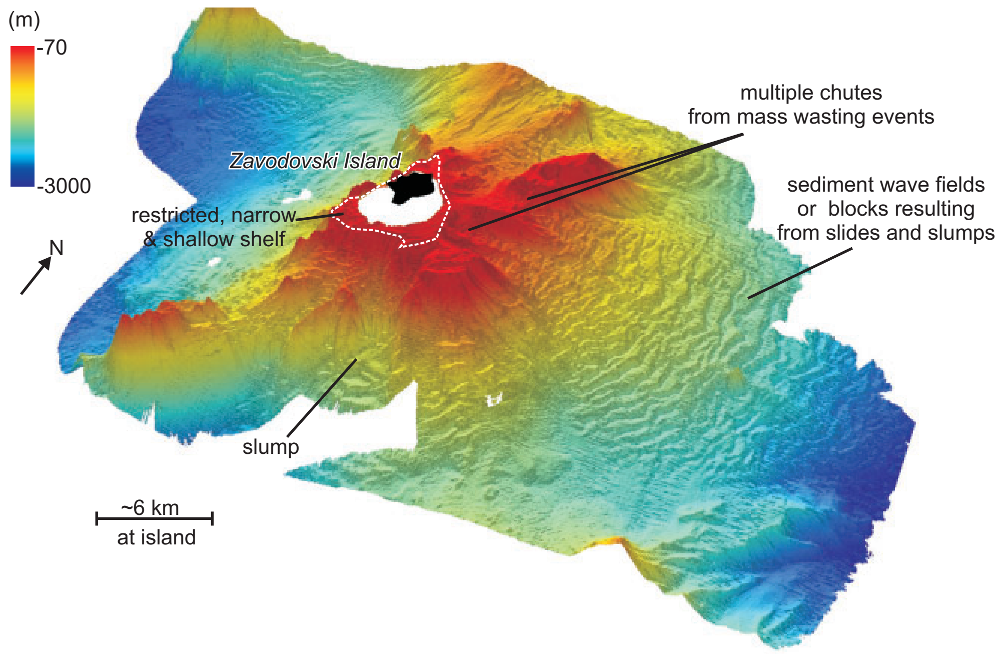

Figure 4, Hodgson et al.

Figure 5

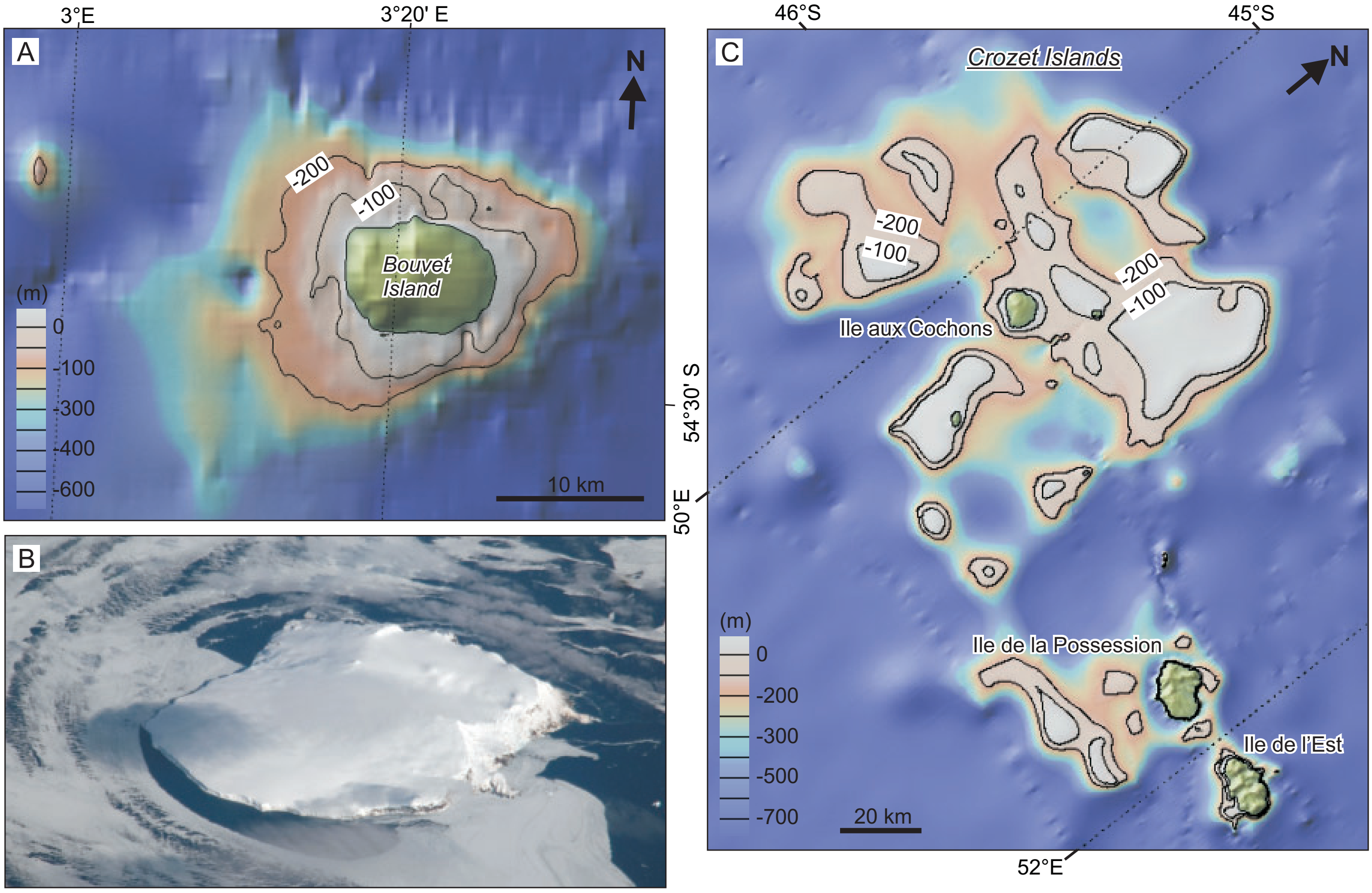

Figure 5, Hodgson et al.

Figure 6

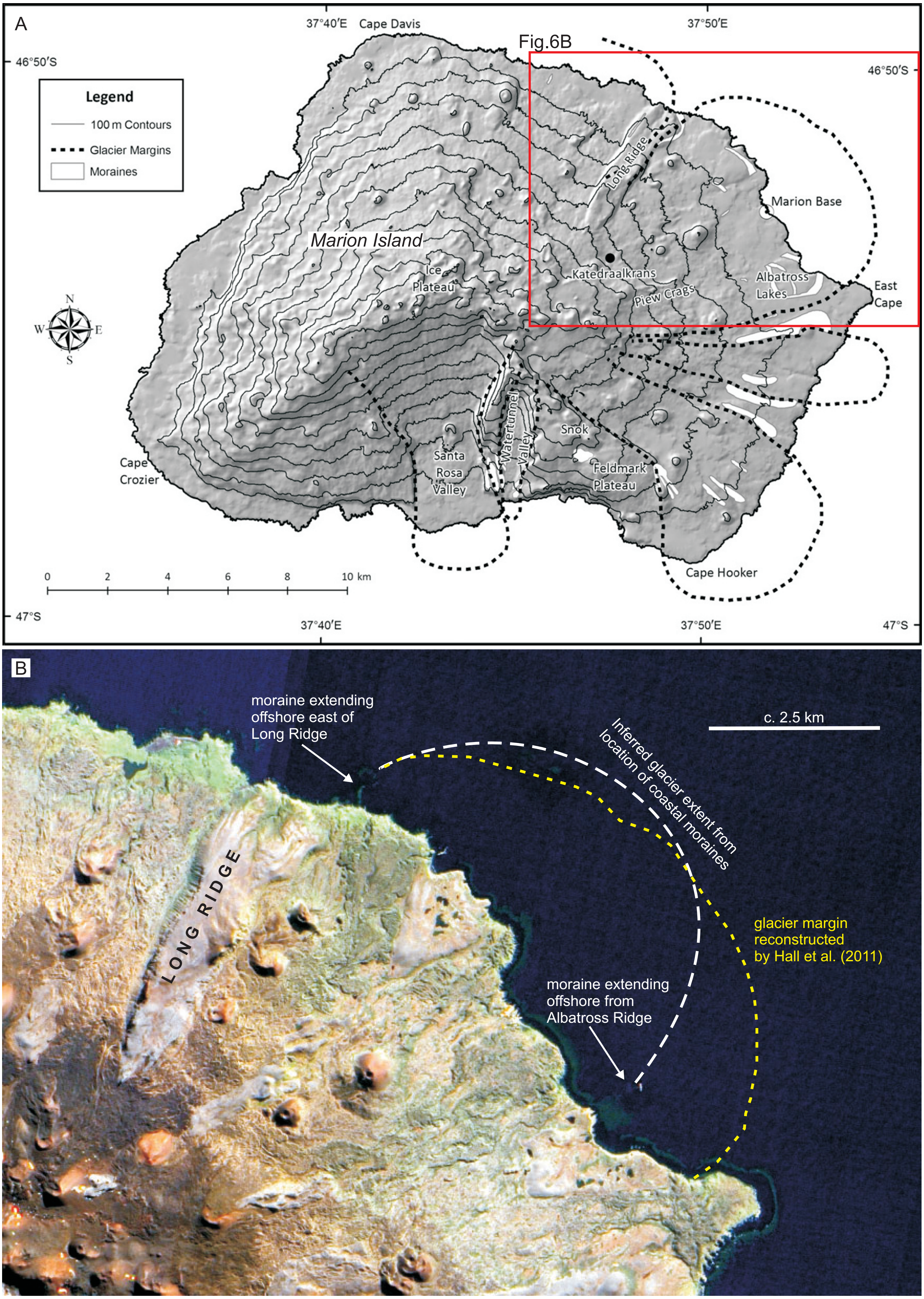

Figure 6, Hodgson et al.

Figure 7

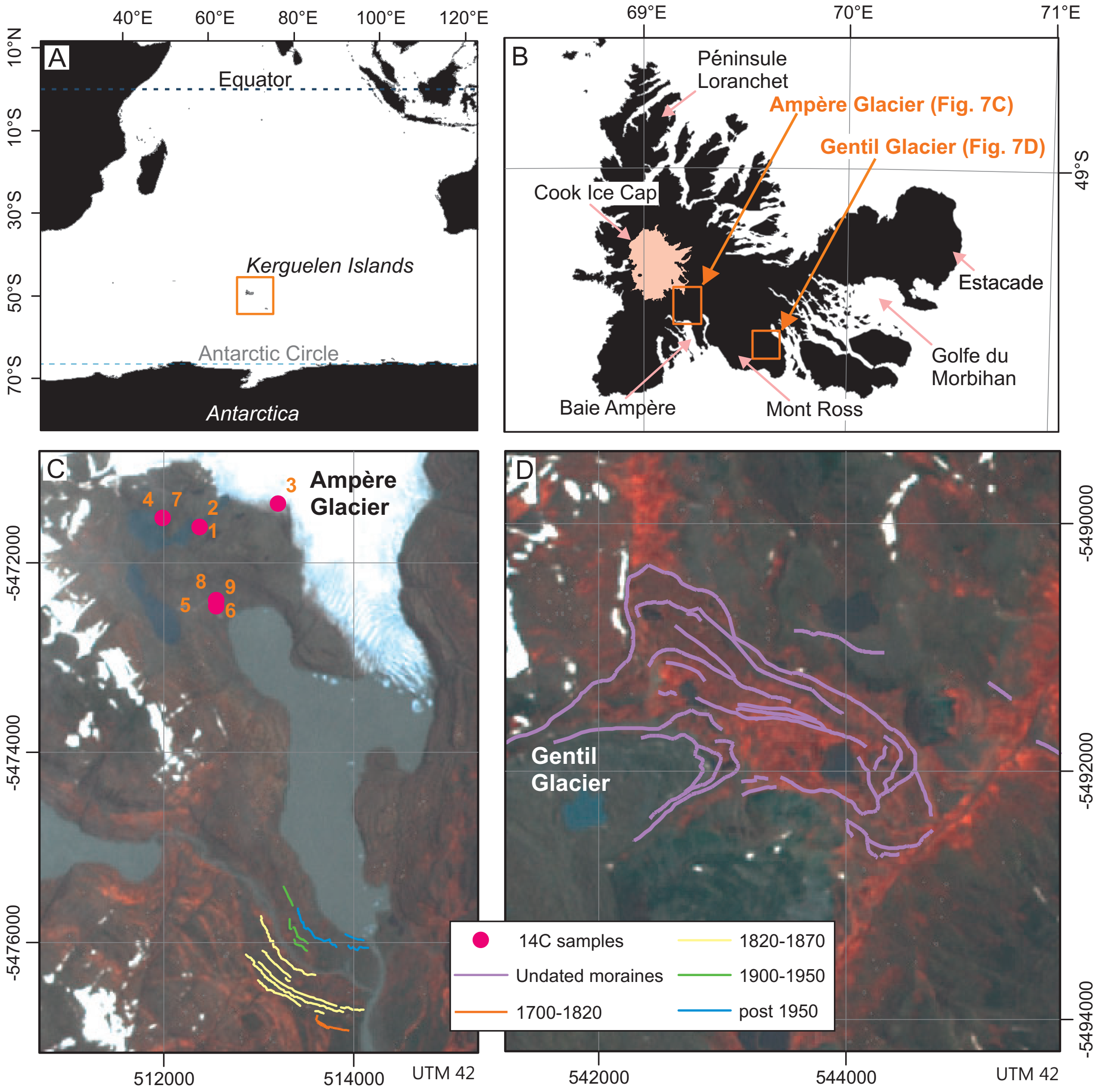

Figure 7, Hodgson et al.

Figure 8

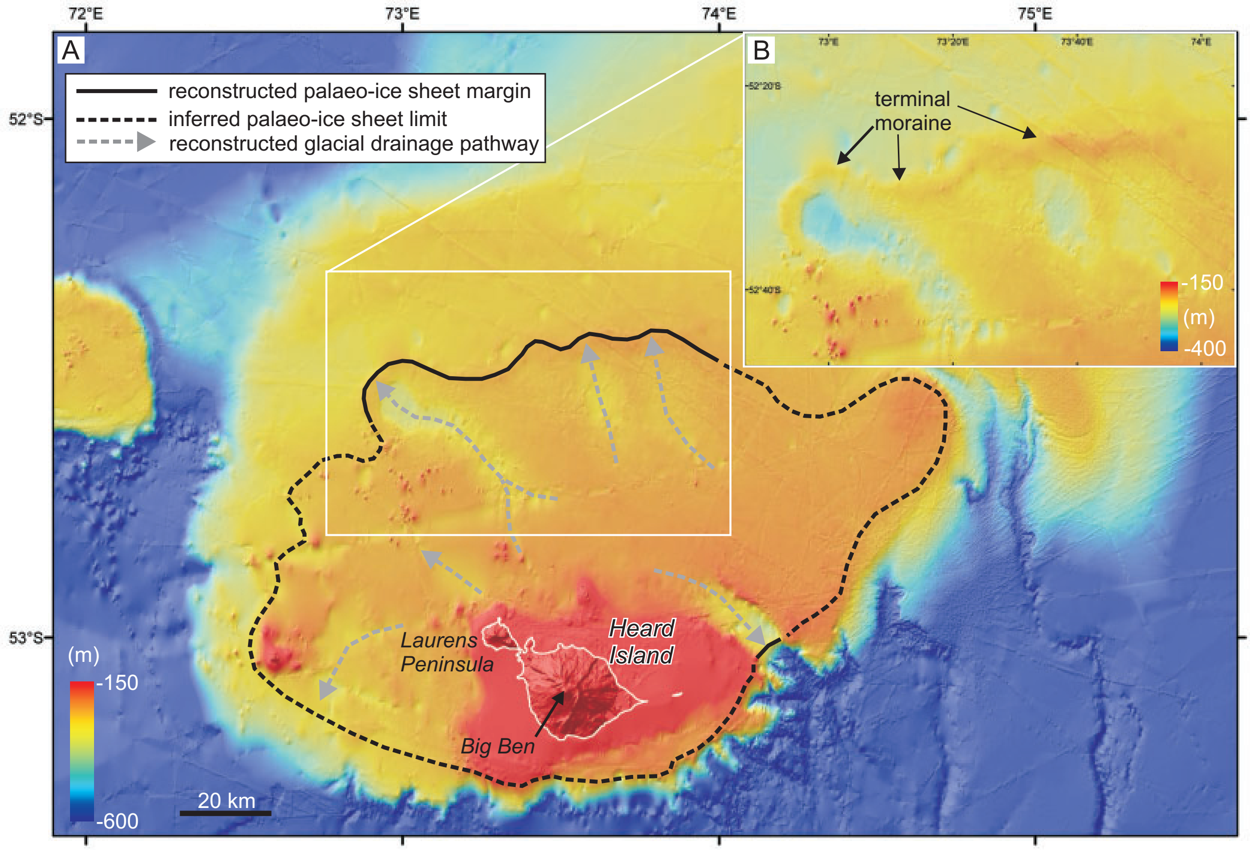

Figure 8, Hodgson et al.

Figure 9

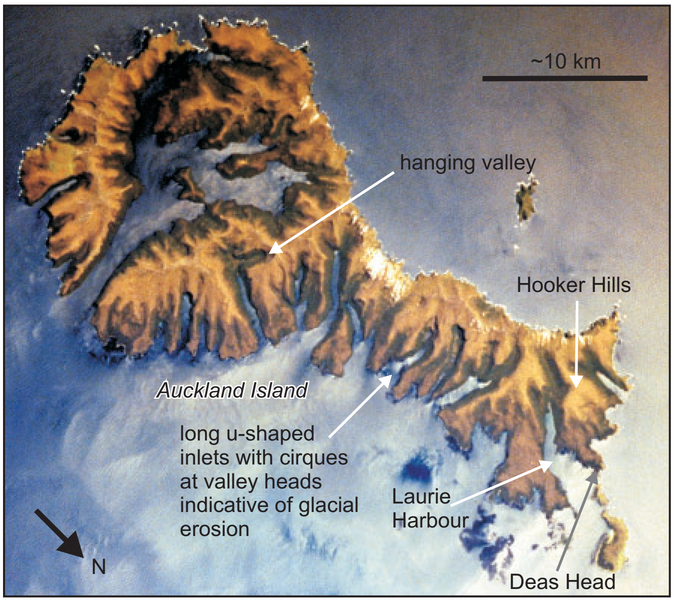

Figure 9, Hodgson et al.