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Fossil proxies of near-shore sea surface temperatures and seasonality from the late Neogene Antarctic shelf

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We evaluate the available palaeontological and geochemical proxy data from bivalves, bryozoans, silicoflagellates, diatoms and cetaceans for sea surface temperature (SST) regimes around the nearshore Antarctic coast during the late Neogene. These fossils can be found in a number of shallow marine sedimentary settings from three regions of the Antarctic continent, the northern Antarctic Peninsula, the Prydz Bay region and the western Ross Sea. Many of the proxies suggest maximum spring-summer SSTs that are warmer than present by up to 5C°, that would result in reduced seasonal sea ice. The evidence suggests that the summers on the Antarctic shelf during the late Neogene experienced most of the warming, while winter SSTs were little changed from present. Feedbacks from changes in summer sea ice cover may have driven much of the late Neogene ocean warming seen in stratigraphic records. Synthesized late Neogene and earliest Quaternary Antarctic shelf proxy data are compared to the multi-model SST estimates of the Pliocene Model Intercomparison Project (PlioMIP) Experiment 2. Despite the fragmentary geographical and temporal context for the SST data, comparisons between the SST warming in each of the three regions represented in the marine palaeontological record of the Antarctic shelf and the PlioMIP climate simulations show a good concordance.

Key words: Antarctica, Palaeoclimate, Pliocene, Proxy, Climate models

1. Introduction

We review fossil evidence for sea surface temperatures and seasonality for the Antarctic shallow marine shelf during the latest Miocene, Pliocene and early Pleistocene (~6 to 2 Ma). For our study, we compare fossil materials sourced from a range of glacigenic and interglacial shallow marine sedimentary deposits from the terrestrial Antarctic Peninsula and East Antarctic regions include invertebrates (bivalve molluscs, bryozoans), vertebrates (cetaceans) and unicellular eukaryotes (silicoflagellates, diatoms). These are also compared to fossil data from local offshore marine sites, allowing a multi-proxy approach where samples are limited. Our review includes but is not exclusive to data from two of the broadly defined warmer climate intervals from the Antarctic Peninsula documented by Smellie et al. (2006) using $^{40}Ar/^{39}Ar$ and $^{87}Sr/^{86}Sr$ methods: the late Miocene (6.5 to 5.9 Ma) and the early Pliocene (5.03 to 4.22 Ma). Though the stratigraphic record of the source deposits and of the fossil materials themselves are often fragmentary, we have compared these with reconstructed sea temperatures for several time intervals in the late Neogene (e.g. Ciesielski and Weaver 1974; Clark et al. 2010; Williams et al. 2010). We also assess peak summer temperatures and sea ice extent during warm climate intervals in the late Neogene. Thus our study provides a calibration for palaeoclimate reconstructions of warm intervals of the Pliocene world (e.g. Dowsett et al. 2010), and potentially for comparing with the Antarctic climate records that span key intervals of West Antarctic ice sheet collapse and regrowth (Naish et al. 2009).

In this paper we aim to: (1) summarise modern marine environmental conditions and faunal distribution adjacent to the fossil sites; (2) provide a summary stratigraphic correlation for the latest Miocene, Pliocene and earliest Pleistocene coastal marine successions of the Antarctic that contain fossil assemblages; (3) summarise the range of palaeoclimate data that can be gleaned from these successions for specific intervals; (4) assess the veracity of the available data and their implications for peak summer warming; and (5) compare these data with the latest palaeoclimate simulations for intervals of late Neogene warmth.

2. Climate and Seasonality in Modern Antarctic Shelf Seas

The Southern Ocean has an area of 35 million km^2 and is covered by 60% sea ice in winter and 20% during the summer (Gutt et al. 2010). Sea ice extent around the Antarctic continent varies throughout the year with an almost complete coastal coverage during the austral winter. The only exceptions are polynyas, persistently open, large (up to 350,000 km^2) areas of waters caused by upwelling of relatively warm water or formed by katabatic winds forcing ice away from fixed boundaries such as the coastline. There are many regions where large volumes of sea ice also persist into the summer months, such as the Weddell Sea and around parts of the East Antarctic coastline (Fig 1).

Modern average seasonal SSTs around the Antarctic coast vary from ca. −1.8°C, at which point sea ice forms in water of normal salinity, to +1.4°C during the summer in particularly warm years (Table 1). The Antarctic Surface Water (AASW) can be stratified by solar heating leading to anomalously high temperatures; for example, temperatures of +5°C were recorded in Marguerite Bay in 2002 (BAS unpublished; Barnes et al. 2006). The temperature range is influenced partly by strong, cold katabatic winds blowing off the Antarctic continent and cold polar waters contained by the Antarctic Circumpolar Current (ACC). These conditions allow for sea ice to be maintained around parts of the Antarctic coast even during the austral summer months. As our synthesis of data focuses on the James Ross and Cockburn islands in the Weddell Sea, and the Vestfold Hills and McMurdo Sound in East Antarctica, we summarize modern oceanographic data for SST variation in these regions. As shown in Table 1 seasonal temperature variations around the Antarctic coast are very small.

2.1 Weddell Sea

The Weddell Sea is an important source region for Antarctic bottom water (AABW), formed when warmer deep water mixes with cooler shelf or ice shelf waters (Deacon 1937; Robertson et al. 2002). Measured potential temperatures from surface waters in the Weddell Sea range from −1.80°C (surface freezing temperature at the present salinity) to +1.15°C (see Table 1 and references therein). Salinity in the same water mass ranges from 33.0 to 34.5 psu in the winter (Weiss et al. 1979; Robertson et al. 2002).

Annual summer flux of glacial meltwater, which has a low $\delta^{18}O$, into the northern Weddell Sea around James Ross Island affects the salinity of surface water. There are no modern detailed studies of meltwater flux around James Ross Island, from which the most extensive Neogene fossil assemblages are derived. By contrast, meltwater fluxes are well constrained for surface water in the western Antarctic Peninsula region in Marguerite Bay at 68°S (Meredith et al. 2008; Williams et al. 2010). The north end of Marguerite Bay is covered by winter sea ice for several months, which provides a useful comparison for seasonal meltwater fluxes from sea ice and glacial sources into the modern James Ross Island area, despite different geographical settings. In Marguerite Bay, between 3 and 5% of the near-surface ocean is estimated as formed by glacial meltwater (Jenkins and Jacobs 2008; Meredith et al. 2008), with sea ice-melt accounting for a much smaller percentage (ca. 1%) (Meredith et al. 2008). The effects of seasonal sea ice-melt on the $\delta^{18}O$ of seawater, and therefore salinity, are minimal (Meredith et al. 2008) but those of glacial ice-melt are considerably more significant as high latitude ice has very low $\delta^{18}O$, with values as low as −50‰ (Weiss et al. 1979; Meredith et al. 2008). The meltwater fraction that we assume for modern waters around James Ross Island may be a little overestimated because here warm water upwelling onto the shelf (Fahrbach et al. 1995) is less significant than in Marguerite Bay, which is affected by upwelling of relatively warm Circumpolar Deep Water (CDW) (e.g., Klinck et al. 2004).

2.2 East Antarctica

SSTs vary around the East Antarctic coast, but for this study we focus on two main areas, Prydz Bay and the Ross Sea. The East Antarctic coastline is very extensive so direct temperature measurements are sparse (see Table 1 for data and references). On average, present-day SSTs around the coast of East Antarctica range from −1.7 to +1°C according to the National Oceanic and Atmospheric Administration (NOAA; www.emc.ncep.noaa.gov/research/cmb/sst_analysis/). In the Ross Sea, close to the coastline, sea surface temperatures have a smaller overall range during the summer (Table 1) but in winter the area is covered by extensive sea ice (Fig 1). A study by Trevisiol et al. (2012) in Terra Nova Bay, Ross Sea demonstrated that Adamussium colbecki bivalve shells could be used to measure, within error, the summer sea temperature of the local environment.

Modern measurements for freshwater influx into the coastal waters around East Antarctica are sparse. Glacial meltwater in Prydz Bay during two Antarctic cruises (13^{th} and 14^{th} Chinese National Antarctic Research expeditions; CHINARE) was calculated to be between 0% and 3.98% (Cai et al. 2003). Mean meltwater concentrations at the front of the Ross Ice Shelf were measured by Loose et al. (2009) and these showed inter-annual variation (for example; 2.2±0.36‰ in 2000 and 0.25±0.1‰ in 2001). Burgess et al. (2010) also interpreted variations in $\delta^{18}O$ from late Holocene barnacle plates (Bathylasma corolliforme) as representing fluctuations in meltwater under the McMurdo Ice Shelf. Past influxes may have been even greater, something which can potentially be tested by looking at variations in strontium ratios in bivalve shells (see section 4.1 for further detail).

In McMurdo Sound peak SSTs were recorded by in situ temperature loggers in mid-January and early February during 2000 and 2001 (see Table 1).

Locality	Modern annual SST range (°C)	Modern seasonal range (°C)	Glacial melt water (Gt a^{-1})	Modern salinity (psu)	Reference
Weddell Sea	−1.75 to +1.15	2.9		34.3 to 34.4	Weiss et al. 1979
			126		Schlosser et al. 1990
	−1.8 to +1.0	2.8		33.0 to 34.5	Robertson et al. 2002
	−1.8 to −1.0	0.8		33.6 to 34.4	Nicholls et al. 2009
	−1.7 to +1.0	2.7			NOAA*
East Antarctic coast	−1.7 to +1.0	2.7			NOAA*
Prydz Bay	−1.1 to −0.6	0.5 to 2.49			Kerry 1987; Kerry et al. 1987
	−1.87 (min)			33.5 to 34.3	Smith & Tréguer 1994
	+1.39				Gibson et al. 1998
			10.7 to 21.9		Wong et al. 1998
Ross Sea/ McMurdo Sound	−2.14 to −0.96 °C	1.18		34.12 to 34.84	Tressler,& Ommundsen, 1962
	0.2 ±0.50 **				Eggers 1979
	−0.35 to −0.65 (peak)				Hunt et al. 2003
	< −1.5			34.30 to 34.36	Orsi & Wiederwohl 2009
			14.2		Pritchard et al. 2012
	−1.7 to 0 (only summer)	1.7			NOAA*
Pacific sector			200 to 215		Hohmann et al. 2002
	−1.6 to +1.0 °C	2.6		<34.2	Smethie Jr & Jacobs 2005

Table 1. Modern oceanographic data measured close to sites where fossil material has been collected. Note that a variety of methods have been used and measurements were taken from different locations.

*****NOAA Optimum Interpolation SST Analysis (www.emc.ncep.noaa.gov/research/cmb/sst_analysis)

** Single temperature from A. colbecki

2.3 Sea Ice

Modern sea ice extent varies from year to year and can be measured using satellite techniques (see Fig 1). It is also very seasonal, with an average summer minimum extent of 3 x 10^6 km² to a winter maximum extent of 18 x 10^6 km² (Zwally et al. 2002; Comiso 2010; Comiso et al. 2011). Based on new passive microwave satellite data analysis, Turner et al. (2009) show that since the late 1970s, Antarctic annual mean sea ice extent has increased at a rate of 0.97% per decade. This pattern agrees with studies by Comiso et al. (2011) who calculated an increase in sea ice concentration at a rate of 0.8% dec^{-1} based on data from 1992 to 2008.

3. Modern Marine faunal distribution and environmental conditions adjacent to the fossil sites

The average depth of the Antarctic continental shelf is 450 m, unusually deep due to subglacial erosion during previous glacial maxima and isostatic depression of the continent because of ice-sheet loading (Clarke and Crame 2010). In comparison to other marine shelves around the world there are no river inputs and the small terrestrial freshwater input comes from glacial processes or wind (Clarke and Crame 2010). Most of the Antarctic coastline is ice, with only a small percentage of rock exposed. The shallow seabed along this coastline is subjected to intense scouring by icebergs, making it hard for many intertidal fauna to establish themselves (Clarke and Johnston 2003). There are currently over 4100 benthic species known in the Southern Ocean (Clarke and Johnston 2003), though this diversity may have increased in recent years because of species invasions (for example king crabs in Palmer Deep; Smith et al. 2012). The total macrofaunal diversity on the continental shelf likely exceeds 15,000 species (Gutt et al. 2004; Aronson et al. 2007); this is under evaluation by the Census of Antarctic Marine Life (CAML: www.caml.aq/). The benthic communities that inhabit the shallow-shelf (<100 m depth) surrounding the vast continent are structurally and functionally archaic when compared to similar communities in other parts of the globe (Aronson et al. 2007). Dense populations of epifaunal suspension feeders such as crinoids, bryozoans and brachiopods are dominant in benthic communities (Aronson et al. 2007) and are associated with coarse glacial sand and gravel glacial substrates (Clarke et al. 2004).

Modern warming is apparent in these waters as crabs reinvading Antarctica that, along with other durophagous predators, could potentially alter the character of the present ecosystem there (Aronson et al. 2007). These organisms are thought to have been carried to the Antarctic in their larval stage by eddies or by Sub-Antarctic water injected into warm deep water masses that impinge onto the Antarctic shelf (Thatje and Fuentes 2003). Other possible modes of transport include driftwood or kelp rafts (Highsmith 1985; Helmuth et al. 1994) as well as ships crossing the Polar Front (Barnes et al. 2006).

3.1 Weddell Sea

Modern coastal environments of James Ross Island and Cockburn Island, where seasonal sea ice is well established (see Fig 1), are characterised by the slow-growing, thin-shelled scallop Adamussium colbecki (see Berkman et al. 2004). This bivalve is thought to have originated in cold, deeper ocean waters and migrated onto the Antarctic shelf during the late Pliocene as conditions cooled (Berkman et al. 2004). A. colbecki lives below sea ice, in similar conditions to the deep ocean, with sustained darkness and limited primary production.

Shores in the Antarctic are usually characterised by a thin algal growth and variable concentrations of limpets (Berry and Rudge 1973 and references therein). On the western Antarctic Peninsula shelf, limpets can be found occupying the upper part of the littoral zone and are mainly *Patinigera polaris*, a species only found south of the Antarctic Convergence Zone (Berry and Rudge 1973).

3.2 East Antarctica

The fossil-rich Sørsdal Formation crops out in the Vestfold Hills, which are flanked by the Sørsdal Glacier 2 km to the south and slowly flowing ice draining the East Antarctic Ice Sheet (EAIS) 15 km to the east (Whitehead et al. 2001). The Lambert Glacier-Amery Ice Shelf system drains 13% of the EAIS into Prydz Bay (Hambrey et al. 1991), about 100 and 200 km away from the Larsemann Hills and the Vestfold Hills respectively. Modern shallow-water bays in the area contain abundant red algae and a single species of bivalve, *Laternula elliptica* (King and Broderip, 1831) (Quilty et al. 2000).

To the west of Prydz Bay is the Mac. Robertson shelf, a relatively narrow part of the continental shelf. This has been heavily eroded and sculpted by grounded ice during glacial maxima, and deposition took place mainly during interglacials (Truswell et al. 1999; Mackintosh et al. 2011); it is very different from the deposition-dominated shelf in Prydz Bay (O'Brien et al. 2007). Strong west-flowing currents transporting sediment and icebergs are dominant at present (Harris and O'Brien 1998; Truswell et al. 1999).

4. Stratigraphic setting of Late Neogene shelf fossil assemblages

There are few fossil-bearing marine deposits of late Neogene age on the Antarctic landmass (e.g. Webb 1991; Taviani & Beu 2003; Stilwell & Long 2011) (Figs 2 and 3). This is invariably due to the scarcity of exposure, which is limited to the coastal edges and some ice-free inland areas. These few onshore sedimentary deposits are much less stratigraphically complete than the Neogene and Quaternary sedimentary record of the adjacent Southern Ocean and Ross Sea (e.g. Jonkers 1998; Naish et al. 2009), but nevertheless provide a unique source of fossil material that can give an indication of the temperature of and seasonality in shelf seas bordering Antarctica in the latest Miocene, Pliocene and earliest Pleistocene. Below, we summarise the main geographical settings for each of the late Neogene fossil-bearing shallow shelf Antarctic sites. We comment on the resolution of the available stratigraphic dating, which involves a combination of biostratigraphic, magnetostratigraphic, chemostratigraphic and radiometric techniques. Many of the sedimentary deposits are glacigenic and therefore have the added taphonomic problem of the reworking of older fossil materials into younger deposits.

4.1 James Ross Island, Antarctic Peninsula

James Ross Island is situated in the NW Weddell Sea, just to the East of the northern tip of the Antarctic Peninsula, at 64.17°S and 57.75°W (Fig 2). The James Ross Island Volcanic Group (JRIVG) dominates the outcrop geology of the island (Smellie et al. 2013). The volcanic rocks unconformably overlie relatively unconsolidated Cretaceous marine sedimentary deposits. Some 10 million years of late Neogene and Quaternary history is recorded in the JRIVG (Smellie et al. 2006; Smellie et al. 2008; Hambrey et al. 2008; Smellie et al. 2009). Sedimentary deposits in the JRIVG are dominated by diamict, conglomerate and minor sandstone, sections of which contain fossil material (Smellie et al. 2006; Williams et al. 2006; Hambrey et al. 2008; Nelson et al. 2009) (Table 3). Collectively these fossils occur in strata of late Miocene (ca. 6 Ma) through early Pleistocene age (ca. 2 Ma). Stratigraphic analysis of the JRIVG has identified three broadly defined intervals of relative climate warmth in the northern peninsula region, when volcanic rocks were erupted into marine environments (Smellie et al. 2006). Radiometric ($^{40}Ar/^{39}Ar$) chronology from the volcanic rocks, together with $^{87}Sr/^{86}Sr$ chronology from the molluscs in the intervening sedimentary deposits, have produced a well resolved stratigraphy which constrains the warm intervals from 6.5 to 5.9 Ma (late Miocene), 5.03 to 4.22 Ma (early Pliocene), and ca 0.88 Ma (late Pleistocene) (the late Pleistocene interval will not be discussed in this paper).

Although marine fossil material is widespread in the sedimentary deposits of the JRIVG (Smellie et al. 2006; Nelson et al. 2009), most material that is suitable for environmental analysis has been sourced from three key localities on James Ross Island in the Hobbs Glacier Formation (labelled D6.404, D6.405 and D6.407 in Table 2). Where data are available, locality ages are bracketed by $^{40}Ar/^{39}Ar$ isotopic ages from overlying and underlying volcanic rocks or can be assigned a minimum age (e.g. at Forster Cliffs this is constrained to 2.5 Ma by a date from an overlying lava-fed delta; Table 2). $^{87}Sr/^{86}Sr$ chronology has also been used on calcitic bivalve shells collected from diamicts (Table 2). Discrepancies occur when comparing $^{40}Ar/^{39}Ar$ ages with $^{87}Sr/^{86}Sr$ ages and $^{87}Sr/^{86}Sr$ ratios obtained from individual shells. Our fossil samples show no adhering terrigenous matrix nor are they abraded, which suggest that reworking from older deposits is probably minor or absent (Nelson et al. 2009). Thus, the range of ages for bivalve molluscs collected from the same layer may indicate that the primary marine $^{87}Sr/^{86}Sr$ ratios have been modified, perhaps by freshwater flux with a different strontium signal during summer ice proximal melting at the time the bivalves were living (see Huang et al. 2011 for evidence of $^{87}Sr/^{86}Sr$ variations due to freshwater input). The presence of a few fossil ages considerably younger than overlying volcanic rocks at some sites (Table 2) is also an indication that some primary Sr ratios have been modified. Samples D6.407.4 and D6.407.8 included in Table 2 are an example of this with a mean Sr age of 2.05 and 1.57 Ma respectively. The preservation protocols that were used to identify altered shells indicated that these individuals showed signs of being recrystallized. The young Sr ages are consistent with this inference and highlight the need for careful assessment of preservation. Wilson et al. (2007) pointed out that Sr-ages on a number of shell fragments sampled from the ANDRILL AND-1B drill core (McMurdo Ice Shelf) do not show the expected down-core increase but provide much older ages for bivalve fragments found in stratigraphically higher sections. The authors concluded that the most likely major contributor to this discrepancy was contamination by less radiogenic matrix sediment. Marcano et al. (2009) noted that different bivalves within the SMS AND-2A core (southern McMurdo Sound) had different Sr isotopic values, with the calcitic pectinid samples having younger ages compared to the associated aragonitic Veneridae bivalves. This was also noted in Acton et al. (2009) who concluded that improved age constraints could be determined by dating using strontium ratios from calcitic bivalves. Although unrecognized recycling of fossils from several late Neogene deposits may have occurred in the JRIVG sedimentary deposits, the presence of articulated and well-preserved shells at some sites is also consistent with minimal or no reworking.

References	Shell Fragment	Locality	GPS	$^{87}Sr/^{86}Sr$	Mean Age	Error (+)	Error (-)
Smellie et al. 2006 (Data	DJ.1741.6	1) Below Forster Cliffs main delta	63°59.8′S, 57°35.6′W	0.709051	4.23	0.51	0.95
calibrated from J McArthur,	DJ.1745.4	3 km SW of Terrapin Hill	63°59.6′S, 57°34.7′W	0.709034	5.06	0.35	0.41
University College London)	DJ.1754.7	2) Below Forster Cliffs main delta	63°59.9′S, 57°29.5′W	0.708971	6.45	0.31	0.21
	DJ.1755.6	Eastern Forster Cliffs	63°59.8′S, 57°30.1′W	0.709038	4.93	0.35	0.51
From Ian Millar, BGS	D6.404.2	3) Northwest Forster Cliffs,	63°99.673’S, 57°58.917’W	0.709071	1.13	0.13	0.17
	D6.404.6	immediately south of the waterfall		0.709083	2.30	0.30	0.25
	D6.404.12			0.709071	2.63	0.83	0.38
From John McArthur, UCL	D6.407	4) South Blancmange Hill	63°59.987’S, 57°37.964’W	0.709038	4.94	0.31	0.45
From Ian Millar, BGS	D6.407.1		63°59.987’S, 57°37.964’W	0.708987	6.25	0.20	0.15
	D6.407.2			0.709017	5.70	0.15	0.14
	D6.407.3			0.709050	4.75	0.25	0.55
	D6.407.4*			0.709088	2.05	0.20	0.35
	D6.407.8*			0.709105	1.57	0.22	0.19
Smellie et al. 2008	D5.7	5) Blancmange Hill	63°99995’S, 57°63315’W	0.709039	4.89	0.35	0.53
From Ian Millar, BGS	D6.405.1	6) Cascade Cliffs	63°59.855’S, 57°35.680’W	0.709071	2.63	0.83	0.38
	D6.405.2			0.709087	2.08	0.30	0.36
	D6.405.3			0.709076	2.55	0.50	0.25
	D6.405.4			0.709059	4.15	0.55	0.95
	D6.405.4			0.709056	4.11	0.66	1.23
	D6.405.6			0.709066	3.30	0.95	0.65
Nỳvlt et al. 2011	PC05-1a	7) Ulu Peninsula	63°48.1222’S, 57°52.5285’W	0.708992	5.93	0.09	0.11
	PC05-1b			0.709012	5.65	0.15	0.24
Pirrie et al. 2011		8) Brandy Bay outcrop	63°50.64’S, 58°01.57’W	0.709050	4.33	0.56	1.31

References	Shell Fragment	Locality	Relation to shell-bearing deposit	GPS	Age (Ma) $^{40}Ar/^{39}Ar$	±2σ
Smellie et al. 2006	DJ.1752.3	1,2) Forster Cliffs	Lava-fed delta - above	64°00.4′S, 57°36.6′W	2.50	0.07
Smellie et al. 2008	DJ.1745.2	1,2) Foster Cliffs	Basal delta - below	63 59.61′S, 57 34.68’W	5.47	0.11
Nỳvlt et al. 2011	PC05-1a	7) Ulu Peninsula	Pillow lava - below	63°48.1222’S,		
	PC05-1b	Mendel Formation		57°52.5285’W	5.85	0.31
Smellie et al. 2008	DJ.1715.1		Basal volcanic delta - above	63 49.63′S, 57 50.48′W	5.32	0.16

Table 2 Compilation of $^{87}Sr/^{86}Sr$ data for Austrochlamys shells (each number indicates a fragment from a different individual specimen) from the James Ross Island region: locality numbers (1-8) are reference numbers for this paper only. $^{40}Ar/^{39}Ar$ ages have been included where available as a comparison (see included references for complete methods and calculations used). Note that in some cases there is a wide range of values for shells from the same deposits. $^{87}Sr/^{86}Sr$ values and ages from Smellie et al. (2006) and Nelson et al. (2009) are mean values calculated from a number of shells from the same locality. Ages as young as 1.57 Ma from South Blancmange Hill are very unlikely to be valid and suggest there may be another contributing factor to the strontium ratios as highlighted above. (Smellie et al. 2006; Smellie et al. 2008; Nelson et al. 2009; Nỳvlt et al. 2011; John McArthur, University College London and Ian Millar, British Geological Survey).
*Individual shell is recrystallized.

More recently there have been further discoveries of Austrochlamys shells on northwestern James Ross Island (Locality DJ.1501; Pirrie et al. 2011). Although reworked into the active layer of modern periglacial sediments, the shelly fossils are indistinguishable in age from those on Cockburn Island (see below).

Bivalves are also reported from the Mendel Formation (Nývlt et al. 2011), found on the Ulu Peninsula in the northern part of James Ross Island, which has an age of 5.9-5.4 Ma according to $^{40}Ar/^{39}Ar$ dates on associated volcanic rocks and $^{87}Sr/^{86}Sr$ dates on individual bivalve shells. It has a thickness of at least 80 m and is restricted in outcrop to the northwestern tip of Ulu Peninsula (Nývlt et al. 2011). The formation is made up of terrestrial glacigenic lodgement and melt-out till, glaciofluvial sandstone, glacimarine debris flow deposits, diamict, tuffaceous sand and siltstone.

4.2 Cockburn Island, Antarctic Peninsula

The JRIVG also crops out on the summit of Cockburn Island and contains the richest late Neogene fossil assemblages from the Antarctic Peninsula region. These are contained within the interglacial marine Cockburn Island Formation (CIF), and include abundant large molluscs, especially Austrochlamys ('Zygochlamys' of Jonkers et al. 2002; see Jonkers (2003) for a detailed taxonomic appraisal). The formation consists of rusty-brown sandstone and conglomerate containing clasts exclusively derived from the JRIVG (Jonkers 1998). $^{40}Ar/^{39}Ar$ ages from the underlying volcanic rocks provide a maximum age of ca. 4.9 Ma for the CIF (Jonkers and Kelley 1998). Ages determined using mean Sr isotope ratios from Dingle et al. (1997); based on Howarth and McArthur (1997) and Sr isotope ratios from bivalve shells (McArthur et al. 2006) suggest a depositional age of 4.7 (+0.6/ −1.2) and 4.66 (+0.17/−0.24) Ma respectively (Pirrie et al. 2011). These ages are stratigraphically consistent with the $^{40}Ar/^{39}Ar$ age of the underlying volcanic rocks. In contrast a biostratigraphic age based on the diatom assemblage suggested that the deposit was formed around 3 Ma (Jonkers and Kelley 1998; Pirrie et al. 2011) but an updated biostratigraphy places this at 2.8 to 2.4 Ma (Cody et al. 2008; 2012; Levy et al. 2012). Ages based on diatom biostratigraphy can be problematic as aeolian transportation of diatoms by wind may have occurred (e.g. Kellogg and Kellogg 1996; Kellogg et al. 1997; Stoeven et al. 1997; McKay et al. 2008). It has also been suggested that marine diatoms of Eocene to Pliocene age were atmospherically transported to Antarctica by the Eltanin asteroid impact in the South Pacific at ~2.15 Ma (Gersonde et al. 1997).

4.3 Vestfold Hills, East Antarctica

The Sørsdal Formation crops out in the Marine Plain area of the Vestfold Hills, covers ~10 km^2 of this area (Fig 4), and consists of horizontal diatomaceous siltstone and sandstone with dark limestone lenses (Quilty et al. 2000; see Table 3 for details). These deposits unconformably infill valleys within Precambrian metamorphic basement rock (Adamson and Pickard 1983). The part of the Sørsdal Formation that contains abundant bivalve material has been dated to the early Pliocene (4.5-4.1 Ma) using diatom and bivalve biostratigraphic records (Adamson and Pickard 1986; Pickard et al. 1986, 1988; Quilty et al. 2000; Harwood et al. 2000) and is supported by magnetic polarity data from the site (Berggren et al. 1995; Quilty et al. 2000). This age has been further constrained to 4.2 to 4.1 Ma (Whitehead et al 2004; Whitehead et al. 2006b). Shelly lenses within diatomaceous sandstone beds contain numerous hiatellid bivalves, dominantly of Hiatella cf. arctica (Quilty et al. 2000) along with fossil dolphin and whale bones (Adamson and Pickard 1983; Quilty 1991); apart from early fracturing, there is no evidence for major diagenetic effects on the vertebrate material (Fordyce et al. 2002). The outcrop is also distinguished by a lack of volcanism, tectonics or reworking, which is unusual for Neogene in coastal Antarctica (Quilty et al. 2000).

To the NE of Marine Plain, Vestfold Hills, younger sedimentary macrofossil-bearing deposits are recorded in the Heidemann Valley (Fig 5). These have been dated as Pliocene in age (3.5 to 2.6 Ma), determined from amino acid racemisation, diatom and foraminifera biostratigraphy and magnetic polarity data (Colhoun et al. 2010). There are uncertainties associated with both the magnetic data, due to unreliable interpretation of polarity changes in the absence of absolute dating, and amino acid dating of molluscs due to assumptions about temperatures controlling the rate of amino acid diagenesis. The foraminifera biostratigraphy only provides an upper age limit due to the absence of Ammoelphidiella antarctica and the fact that all other foraminifera reported are extant. The diatom biostratigraphy used by Colhoun et al. (2010) is based on the presence of Thalassiosira insigna (Jousé) Harwood & Maruyama, 1992 and Fragilariopsis kerguelensis (O'Meara) Hasle, 1958, indicating an age of >2.6 Ma and <3.2 Ma respectively. However, using these same species and biostratigraphy from Cody et al. (2008), the age range would be 2.29 to 2.19 Ma, which would not correlate with the other dating methods used. The Heidemann Valley deposits may correlate with a marine incursion (M10 of Whitehead et al. 2006b) dated at 3.2 to 2.5 Ma that can be traced

500 km to the SW at Amery Oasis and is part of a major marine incursion onto the East Antarctic craton (Whitehead et al. 2006b; Quilty 2010). The deposit at Heidemann Valley is just 4 m thick and it can only be reached by digging trenches through regolith (see Colhoun et al. 2010).

4.4 Larsemann Hills, East Antarctica

Pliocene deposits also occur in the Larsemann Hills, about 100 km SW of the Vestfold Hills (see Fig 5). They consist of grey clayey sand that contains in situ fossiliferous material similar to that obtained in the Vestfold Hills (Quilty et al. 1990). The Larsemann Hills deposits were originally dated as Pliocene using the age range of the foraminifer Ammoelphidiella antarctica (3.8-2.5 Ma; Webb 1974; Quilty et al. 1990), though this was later revised to 4.5 to 3.8 Ma (McMinn and Harwood 1995).

4.5 McMurdo Sound, East Antarctica

The Scallop Hill Formation (SHF; Figs 2 and 3) in the McMurdo Sound region of the Ross Sea was first defined by Speden (1962) and contains pectinid-bearing deposits (Jonkers 1998). This is the older of two main sedimentary successions in this region (the younger being the Taylor Formation) and consists of tuffaceous sandstone and conglomerate containing Austrochlamys anderssoni (Speden 1962; Jonkers 1998). Balanomorph barnacles belonging to an extant taxon Bathylasma corolliforme (Hoek) (Hexelasma antarcticum Borradaile in Speden [1962]) have also been found (Jonkers 1998). Most exposures of the SHF are not in situ, for example, exposures on Ross and White islands. Speden (1962) suggested it is only in situ at the type locality on Scallop Hill, a trachytic dome on Black Island, but further studies by Leckie and Webb (1979) contradicted this, concluding that all localities where the SHF is seen have been displaced from their original position. K-Ar dates from the underlying trachyte at Scallop Hill yield an age of 4.4 ±0.6 Ma (early Pliocene; Leckie and Webb 1979) and 3.8 ±0.2 Ma (Eggers 1979). K-Ar dating of the underlying Aurora trachyte on the Brown Peninsula yielded an age of 2.25 Ma and of the basalt overlying the SHF an age of 2.2 Ma (early Pleistocene; Eggers 1979). This relationship was not replicated by the study of Webb and Andreasen (1986) but K-Ar ages from volcanic boulders in the formation suggested a maximum age of latest Pliocene (2.62 ± 0.04 Ma and 2.58 ±0.09 Ma; see Eggers 1979; Jonkers 1998).

4.6 Wright Valley, East Antarctica

The Prospect Mesa Gravel Formation (PMGF; previously known as the Pecten Gravels; Nichols 1971) has also been called the Prospect Formation (Vucetich and Topping 1972). It forms part of the Wright Valley sedimentary deposits (Fig 2) of Victoria Land, and represents an interval of fossiliferous gravel deposition (Prentice et al. 1993) containing thick-shelled pectinids described as Austrochlamys tuftsensis (Jonkers 2003) (previously known as Chlamys or Zygochlamys tuftsensis; Turner 1967; Jonkers 1998). The presence of the diatoms Fragilariopsis kerguelensis (O'Meara) and Thalassiosira insignis (Jousé) (Prentice et al. 1993) suggested an age of 3 to 2.5 Ma (Jonkers 1998). Recent diatom stratigraphy of Cody et al. (2008) suggests an age around 2.2 Ma. An age of 5.5 ± 0.4 Ma was reported from Sr-dating of A. tuftsensis shells (Prentice et al. 1993; Jonkers 1998). This is very different than that suggested by the diatom biostratigraphy, indicating potential issues with one or both of these methods (as mentioned previously). However, as we have noted for Sr-isotope ages obtained on James Ross Island shells, salinity variations experienced by the shells during growth may have affected the Sr isotopic ratios measured. Jonkers (1998) also queried the reliability of the Sr-isotope approach to dating these deposits. The overlying Peleus Till contains the same marine diatom species as the Prospect Mesa gravels which have been inferred to have been reworked and may thus be younger than 3 Ma (Prentice et al. 1993). Hall et al. (1997) used field relationships and $^{40}Ar/^{39}Ar$ isotope dating of basalt erratics to determine an age of >3.8 Ma for the Peleus Till.

Reference	Locality		GPS	Formation/ Member	Age Range	Lithology	Sedimentary setting/ water depth	Material Collected
Pirrie et al. 1997; Dingle and Lavelle 1998; Jonkers 1998; Smellie et al. 2006; Hambrey et al. 2008; Nelson et al. 2009	James Ross Island	1-3) Forster Cliffs	63°99.673'S 57°58.917'W	Hobbs Glacier Formation	6.2 to 2.5 Ma; Ar* of overlying volcanics indicates diamict is older than 2.5 Ma Sr** 2.63 to 1.13 Ma ***	Diamict-dominated	Glaciomarine close to ice-front	Austrochlamys bivalves, bryozoans
Williams, M per comm	James Ross Island	6) Cascade Cliffs	63°59.855'S 57°35.680'W	JRIVG	Sr** 4.15 to 2.08 Ma***	Diamict		Austrochlamys bivalves, bryozoans
Nelson et al. 2009	James Ross Island	4) South Blancmange Hill	63°59.987'S 57°37.964'W	JRIVG	Sr** 6.25 to 1.57 Ma (6 shells)***	Well-bedded fossiliferous conglomerate	Ice-proximal glacial debris flow; episodes of ice expansion within a relatively warm period	Austrochlamys bivalves, bryozoans
Nỳvlt et al. 2011	James Ross Island	7) Ulu Peninsula	63°48.1222'S 57°52.5285'W	Mendel Formation	5.9 to 5.4 Ma	Dominated by sandy to intermediate glaciomarine diamict	At least two glacials and an interglacial; marine pro-delta with little glacial influence	A. anderssoni, bryozoans, barnacles, foraminifera, fish teeth and ostracods
Pirrie et al. 2011	James Ross Island	8) Brandy Bay outcrop	63°850.64'S 58°801.57'W	Brandy Bay outcrop	Sr** 4.33 Ma	Periglacial reworking of early Pliocene sediments	Marine conditions	A. anderssoni
Jonkers 1998; Jonkers and Kelley 1998; Williams et al. 2010	Cockburn Island		64°12′S 56°51′W	Cockburn Island Formation	3 Ma (diatom biostratigraphy) or 4.7 Ma (Ar* and Sr**)	Interbedded fossiliferous conglomerate and sandstone containing clasts exclusively from the JRIVG	Ice-free interglacial marine, shallower than 100 m	Austrochlamys, brachiopods, barnacles
Pickard et al. 1988; Quilty et al. 1990; Quilty 1991; Quilty et al. 2000; Harwood et al. 2000; Whitehead et al. 2004	Vestfold Hills	Marine Plain	68°38.5'S 78°08'E	Sørsdal Formation/ Graveyard Sandstone Member (GSM)	4.2 to 4.1 Ma	In situ horizontal bedded successions of diatomaceous silt and sandstone and diamicts. GSM: well-rounded, highly lithified, sandy diamict containing bivalves in living position	Shallow, fully marine (within photic zone); maximum depth 25 m. Range of islands and bays relating to modern topography. GSM: higher energy environment; change from to glacial setting	Bivalves including; A. tuftensis, Cyclocardia and hiatellids, diatoms, bryozoans, cetaceans, foraminifera
Colhoun et al. 2010; Quilty 2010	Vestfold Hills	Heidemann Valley	68°34.3’S 78°01.9’E	Heidemann Valley trench	3.5 to 2.6 Ma	Till beds; poorly sorted sand containing cobbles and pebbles with lenses of sand and gravel	Shallow, fully marine bay with locally sourced marine sediments. Low temp. in shallow, narrow embayment	Hiatellids, Laternula, diatoms, foraminifera

Quilty et al. 1990; Quilty 1993; McMinn and Harwood 1995	Larsemann Hills		69°24.7S 76°09.0E	Not formally named	4.5 to 3.8 Ma Possibly coeval with Marine Plain	Grey clayey poorly sorted sand	Shallow, fully marine at or above wave base (depths <50 m; likely 10-20 m). Possibly warmer with higher sea level; indications of less sea ice	Hiatellids, pectinids, foraminifera, diatoms
Speden 1962; Eggers 1979; Jonkers 1998	McMurdo Sound Region	Brown Peninsula and Black, White, Ross Islands		Scallop Hill Formation	2.6 to 2.4 Ma	Cemented, tuffaceous sandstone and conglomerate	Shallow marine environment, probably less than 100 m depth and possibly under an ice sheet at time	A. anderssoni, Balanomorph barnacles, corals, bryozoans, foraminifera, ostracods, plant fragments
Webb 1972; Prentice et al. 1993; Hall et al. 1993; Jonkers 1998	Victoria Land	Wright Valley		Prospect Mesa Gravel Formation	3 to 2.5 Ma Sr** 5.5±0.4 Ma	Fossiliferous sandy gravels and massive mudstone	Fjord deposit	A. tuftsensis, foraminifera, diatoms

Table 3 Stratigraphic information for Pliocene aged land-based marine fossil-bearing sites around the coast of Antarctica. Material collected includes fossils used for palaeoclimatic reconstructions only. Age ranges given are general ranges from the published literature (see Figs 2 and 4 for more detail). Ages from $^{40}Ar/^{39}Ar$ and $^{87}Sr/^{86}Sr$ dating have also been included. JRIVG - James Ross Island Volcanic Group

* $^{40}Ar/^{39}Ar$ dating

** $^{87}Sr/^{86}Sr$ dating of bivalve shells

***The Sr ages may be unreliable due to potential fresh water influences, local or diagenetic effects; the full range of values is included in each case

5. Environmental Setting of Late Neogene Fossil-bearing Shallow Marine Shelf Deposits

5.1 James Ross Island, Antarctic Peninsula

On James Ross Island, late Neogene marine fauna have been recovered from both ice-proximal debris flow deposits and diamicts (Nelson et al. 2009). Fossiliferous glacigenic debris flow deposits at Blancmange Hill, signifying episodes of ice expansion within a relatively warm period, mainly accumulated along the margins of expanding ice masses (Nelson et al. 2009; Table 3). Austrochlamys bivalves appear to have been living on the shelf at the time of this ice expansion, and articulated specimens are included in the debris flow deposits at Blancmange Hill (Fig 6), suggesting minimal transport. This could suggest that the ice sheets were expanding during an interglacial period, possibly due to a 'snow-gun' affect (Bart 2001; Nelson et al. 2009; Smellie et al. 2009) or minor climate fluctuations. Material within diamicts may represent organisms living on the marine shelf immediately prior to the time of ice advance (and diamict formation) as is suggested by the occurrence of articulated bivalves at East Forster Cliffs. The numerous encrusting bryozoans on rounded blocks of local James Ross Island origin in the diamicts suggest local marine conditions and these same bryozoans encrust the inner surfaces of some Austrochlamys shells.

The Mendel Formation in northwestern James Ross Island has a sparse macrofauna of Austrochlamys in massive sandy to intermediate diamict (33-67% sand content in the matrix based on the classification by Moncrieff, 1989) (PC05-1 from Nỳvlt et al. (2011). Marine fossil material is found only within terrestrial deposits indicating that is has been reworked (Nỳvlt et al. 2011).

5.2 Cockburn Island, Antarctic Peninsula

Within the Cockburn Island Formation Jonkers and Kelley (1998) recognised a western 'proximal' or 'littoral' facies and an eastern 'distal' deeper water facies; the latter is estimated to represent original water depths no greater than 100 m (see Table 3). The presence of large thick-shelled pectens, sessile barnacles and a lack of ice-rafted debris suggest deposition in a mainly ice-free environment (Jonkers and Kelley 1998), consistent with reconstructed seasonality and growth patterns from the bivalves (Williams et al. 2010).

5.3 Vestfold Hills, East Antarctica

The Sørsdal Formation (Table 3) represents a shallow fully marine environment within the photic zone in a setting interpreted as representing bays with adjacent islands, comparable to the modern topography (Quilty et al. 2000). The palaeoclimate at the time of deposition (4.5-4.1 Ma) is thought to have been warmer than present with reduced or no sea ice, interpreted from a low abundance of sea-ice diatoms (compared to present) (Harwood 1986; Pickard et al. 1986, 1988; Quilty 1991, 1992, 1993; Fordyce and Quilty 1994; Whitehead 2001). There is no lithological evidence for glacial erosion or re-deposition from the part of the succession composed of diatomaceous siltstone (Quilty et al. 2000). During deposition of the Sørsdal Formation it is assumed that the ice-sheet margin was at least 50 km farther inland compared to its current position, due to the large volume of aeolian sand grains within the formation (Pickard et al. 1988; Whitehead et al. 2001). It is interesting to note that the more open marine sedimentary deposits of the Vestfold Hills bear well-preserved carbonate microfossils whereas diatomites cropping out in the Marine Plain lack any calcareous microfossils. This is consistent with the Marine Plain representing a quieter, more restricted environment, probably in embayments (Quilty 1991). The absence of coccolithophorids has been noted by Whitehead et al. (2001). These have a restricted abundance in waters <+5°C (Dmitriyenko 1989) and limited growth in water <+3°C (Burckle and Pokras 1991), but a higher abundance in waters >+5°C (Goodell 1973; Burckle et al. 1996). The absence of coccolithophores suggests, but does not prove, water temperatures of <+5°C (Whitehead et al. 2001). Specimens of a newly named dolphin species *Australodelphis mirus*, found in Marine Plain, are compared to species that do not permanently occupy coastal Antarctic waters at present, though this is partly based on lack of sightings (Fordyce et al. 2002). There are some modern species of dolphin that have seasonal occurrences in these waters such as the hourglass dolphin (*Lagenorhynchus cruciger*) (Kasamatsu and Joyce 1995). The general absence of sea-ice associated diatoms in the deposits containing *A. mirus* (Pickard et al. 1998) suggests these dolphins lived in an environment free of significant sea-ice (Fordyce et al. 2002). Along with estimates of +4 to +5°C for sea temperatures, the big assumption made for this interpretation is that *A. mirus* were similar to their nearest living relatives today, an assumption that may be inaccurate due to potential environmental drift.

The overlying Graveyard Sandstone Member (GSM, Table 3) is a thin bed that indicates short-lived glacimarine conditions in the latter part of the Marine Plain deposition. From the lower part of the Sørsdal Formation to the upper GSM there is a change of depositional environment from non-glacial to glacial and with this an associated cooling (Quilty et al. 2000). The diatom flora suggests shallowing and cooling, consistent with an expanding ice sheet at that time (4.5-4.1 Ma) (Harwood et al. 2000; Quilty et al. 2000).

At the younger Heidemann Valley locality (Table 3) deposition occurred in a narrow valley under glacial conditions (Quilty 2010) during a marine incursion onto the continent. Water temperatures were close to the freezing point of seawater (−1°C to +1 or +2°C), similar to present. These deposits are nearly coeval with the Wright Valley (Dry Valleys) and probably represent the last time the area was covered by an enlarged EAIS (Colhoun et al. 2010).

5.4 Larsemann Hills, East Antarctica

The sedimentary deposit at Larsemann Hills, thought to be coeval with Marine Plain, Vestfold Hills, contains fragmentary molluscs (Hiatella sp.) and lacks planktonic foraminifer species, suggesting a shallow water, fully marine environment at or above wave base (Quilty et al. 1990). The marine fossils found there are different from the modern faunas found in the region, likely due to its shallower origin (Quilty et al. 1990). Assuming a similar age of deposition to Marine Plain (4.5-4.1 Ma), this suggests the area had less intense glaciation than at present (Pickard et al. 1988) and was probably warmer with a higher sea level (Quilty et al. 1990).

5.5 McMurdo Sound & Wright Valley, East Antarctica

Both the Scallop Hill Formation (SHF) and the Prospect Mesa Gravel Formation have comparable microfaunas similar to present-day species living in fjord, channel and inshore glaciomarine environments in Alaska (Todd and Low 1967) and British Columbia (Cockbain 1963; Eggers 1979). Eggers (1979) therefore concluded that the SHF indicated a shallow marine environment, probably less than 100 m depth and under an ice shelf. He also compared these conditions to the present day conditions under the Ross Ice Shelf (Kennett 1968). Webb (1972) interpreted the fossiliferous Prospect gravels, of limited lateral extent in the Wright Valley, as a fjord deposit (Jonkers 1998). Prentice et al. (1987) inferred water temperatures of less than +5°C using marine diatoms, a much cooler estimate than Webb (1972) who suggested a warmer fjord with temperatures of up to +10°C (Prentice et al. 1993).

6. Late Neogene Climate and Seasonality of Antarctic Coastal Regions

The overall cooling trend seen throughout the Pliocene (Zachos et al. 2001) has influenced many of the species that once lived around coastal parts of the Antarctic continent (e.g. Beu 1995; Clarke et al. 1992). Many, including predators at higher trophic levels, moved to lower latitudes in order to survive. High-latitude Chlamys-like scallops (e.g. Zygochlamys delicatula) similarly migrated northwards during colder periods of the late Pliocene and early Pleistocene (e.g. Beu 1995) demonstrating a sensitivity to climatic variation (Jonkers 1998). As these chlamydinids are not found in modern Southern Ocean ecosystems it is important to understand why they are present in late Cenozoic deposits in this region (Jonkers 1998).

The late Neogene fossil assemblages from the Weddell Sea and East Antarctic coast contain bivalves, bryozoans, microfossils and vertebrates that have been used for environmental analysis. Of these, the bivalves and bryozoans (Fig 7) provide information that enables maximum and minimum sea temperatures and seasonality to be determined from morphological and geochemical data within the organisms' skeletons.

Palaeoclimate data from various marine-shelf fossil organisms preserved in the Neogene deposits of the Antarctic are summarised in Table 4.

	Proxy	Living environment	Locality age	Locality/GPS	Interpreted SST range	Interpretation/Sea ice?	References
Bivalve molluscs	Pectinidae	Shallow waters up to 100 m					
	Austrochlamys anderssoni	Favours open marine, sea ice free conditions facilitating a yearround food supply. Water depths no greater than 100 m. Modern A. natans (closest living relative) occur in high energy sub-littoral and littoral zones of southern Chile and Argentina.	4.7 Ma (Cockburn Island)	James Ross Island and Cockburn Island	−1.1 to +3.5°C (Cockburn Island)	Interglacial marine conditions with little sea ice	Jonkers 1998; Jonkers et al. 2002; Jonkers 2003; Berkman et al. 2004; Dijkstra and Marshall 2008; Williams et al. 2010
				Scallop Hill Fm, McMurdo Sound	+2±0.13°C & +2±0.32°C	Comparable to present day conditions under the Ross Ice Shelf	Eggers 1979
	Austrochlamys tuftsensis		Pliocene		+10.5°C	Maxima temperature not precise (using $\delta^{18}O$ of 0‰)	Quilty 1991
	Hiatellidae	Slow-growing. Found in lower intertidal and subtidal environments; up to depths of 200 m in cool-temperate to polar waters. Can survive salinities of 20‰ to normal marine	4.5 to 3.8 Ma	Vestfold Hills			Peacock 1989; Gordillo 2001;Whitehead et al. 2006a
	Cyclocardia sp.	Known to live in fully marine conditions along rocky shorelines. Found in water depths of 9 m to up to 600 m	4.2 to 4.1 Ma	Marine Plain, Vestfold Hills			Quilty et al. 2000; WoRMS Natural Geography in Shore Areas (NaGISA) database
Bryozoans		Can occur on shallow and deep shelves along the coastline; can live in annually or seasonally ice-free habitats	5.06 to 4.23 Ma	James Ross Island	MART 4 to 10		Schäfer 2008; Clark et al. 2010
Silicoflagellates	Dictyocha/ Distephanus	Dictyocha is rare in regions South of 56°S MART is consistently between <0 and +1.5°C; Distephanus is dominant	4.30 to 3.95 Ma	110°E to 160°W up to 69°S	<+1.5°C to >+10°C	Too warm for sea ice at certain intervals. Displacement to the South of northern Sub-Antarctic waters	Ciesielski and Weaver 1974
			3.7 Ma (event I), 4.4-4.3 Ma (event II) and	64.380°S, 67.219°E - offshore from Prydz Bay	event I: +5°C, event II: −4°C, event III: −4°C (mean annual SST)	Time intervals where Dictyocha becomes dominant are interpreted as warmer. These events may represent times when NADW	Whitehead and Bohaty 2003

			4.80 to 4.55 Ma (event III)			production and OHT into the Southern Ocean exert maximum influence	
			3.58 Ma	Offshore Prydz Bay and West of the Antarctic Peninsula	+2.5C° to +4C° warmer than present	Overall warmer temperatures than present during interglacials with times of reduced sea-ice cover in Prydz Bay and west of the Antarctic Peninsula	Escutia et al. 2009
Diatoms	Eucampia index	Eucampia antarctica: widely distributed in the Southern Ocean and used as a sea ice proxy	Pliocene; 5 to 2 Ma	64.380°S, 67.219°E and 67.696°S, 74.787°E	N/A	Reduced sea ice conditions; 61% to 78% relative reduction in sea ice with possible periods of sea ice absence	Whitehead et al. 2005
	AND-1B diatom record		5.50 to 4.75 Ma	Ross Sea	~+3 to +4°C	Annual sea ice cover limited to absent during interglacial periods	Levy et al. 2012
			4.6 to 3.3 Ma		+4 to +5°C in summer (not above 8°C)	Pliocene ocean circulation and bottom-water formation were significantly different than today	McKay et al. 2012
			2.9 to 2.0 Ma		Up to +3°C		Sjunneskog and Winter 2012
	Extant diatoms		4.5 to 4.1 Ma	Sørsdal Formation	−1.8 to +5°C (summer temperatures >+3°C)	Stratified open water conditions during summer/spring with significant reduction of sea ice	Whitehead et al. 2001
	Sea-ice associated diatoms			Sørsdal Formation	+1 to +2°C	Low abundance of these diatoms suggest reduced sea ice conditions/ similar to present day Kerguelen Plateau values	Whitehead et al. 2001
Vertebrates	Extinct dolphins (Australodelphis mirus)	Similar modern fauna are found in waters +4 to +5°C near the APFZ. Nothing suggests they were functionally adapted to cold water; at least seasonally absent sea ice	4.5 to 4.1 Ma	Marine Plain	Up to +4 to +5°C (based on closest living relative)	Open water environment, based on closest living relative, suggests seasonally absent sea ice	Adamson and Pickard 1983; Quilty et al. 1990; Quilty 1991; Quilty 1993; Fordyce et al. 2002
Barnacles	Bathylasma corolliforme	Relatively large (up to 10 cm) with a benthic habitat. Found in limited localities in the northern part of the Antarctic Peninsula	2.6 to 2.4 Ma	McMurdo Sound		Grounded ice-free environment	Jonkers 1998; Burgess et al. 2010; Smithsonian Natural History Museum Antarctic Invertebrates
Coccolithophorids			4.5 to 4.1 Ma	Sørsdal Formation	<+5°C	Lack of these may indicate cooler temperatures	Whitehead et al. 2001

Table 4. Proxy data from Pliocene coastal Antarctic sites with associated interpretations about temperature and sea ice (where available) during this time. Note that some proxies are described at species-level while others are more general. NADW – North Atlantic Deep Water, OHT – Ocean heat transport, APFZ – Antarctic Polar Frontal Zone

6.1 Sea Surface Temperatures

Many proxies discussed here suggest maximum temperatures warmer than present during particular intervals in the Pliocene. Some of them also indicate, either from direct measurements (e.g. stable oxygen isotope measurements from A. anderssoni in Williams et al. 2010) or indirectly using species tolerances (e.g. the ratio of Dictyocha/Distephanus used by Whitehead and Bohaty 2003), that there was a reduction in sea ice during these warmer intervals. These data have been sourced from various environments both on land (deposits discussed in this paper), in shallow marine deposits, and beneath the ocean from core material. Available data, though, still remain sparse.

A number of proxies have converging temperature data; for example maximum temperatures of up to +5°C during the summer months have been calculated in deposits of overlapping ages (4.6 to 3.3 Ma) in Prydz Bay, the Vestfold Hills, Ross Sea and the western Antarctic Peninsula margin using silicoflagellates (Whitehead and Bohaty 2003; Escutia et al. 2009), diatoms (Whitehead et al. 2001; Scherer et al. 2010), and vertebrates (Quilty et al. 1990; Quilty 1991; Quilty 1993); see Table 4 for details. These same data also indicated a reduction in sea ice and open ocean conditions during the summer months.

The Pliocene sections presented here do not seem to provide evidence for temperature trends (cooling, warming) through the Pliocene, but as the ages of some deposits are only loosely constrained while others are only defined as ‘Pliocene deposits,’ this is to be expected. This is in contrast to evidence from the ANDRILL AND-1B core which shows generally warmer conditions with reduced sea ice extent and a smaller West Antarctic Ice Sheet (WAIS) before 3.3 Ma followed by coastal cooling of ~2.4°C, with increasingly more persistent sea ice between 3.3 and 2.6 Ma (McKay et al. 2012). Land-based data discussed in this paper, however, shows evidence for periods of warmth (4.6 to 3.3 Ma as mentioned previously) that can be seen in different regions at approximately the same time, including one of the warm periods (5.03 to 4.22 Ma) recognised by Smellie et al. (2006).

6.2 Seasonality

Present temperature measurements around the Antarctic coast indicate relatively consistent mean temperatures with slight variability in maximum temperatures recorded and greater seasonal ranges in some areas (See Table 1 for more details). Minimum temperatures are predominantly just above the freezing point of sea water as most coastal regions are only covered by seasonal sea ice. When compared to the proxies from the Pliocene discussed in this paper there are obvious differences, especially between maximum temperatures and seasonal ranges. Some proxies such as the bivalves (based on the range of calculated temperatures from $\delta^{18}O$ isotopes: Williams et al. 2010) and extant diatoms (based on the abundance and known temperature tolerances; Whitehead et al. 2001) have temperatures that overlap with modern day ranges in their minimum values (-1.1°C and -1.8°C respectively) but many of the proxies show significantly higher (up to 5°C) summer maximum temperatures during the warmer intervals of the Pliocene. The greater amplitude between maximum and minimum SSTs inferred for Antarctica for the late Neogene, which today is characterised by only subtle seasonal SST variations throughout the year, indicates that there seasonality during the Pliocene was much more pronounced than today.

6.3 Sea Ice

Evidence for sea ice coverage around the coast of Antarctica during the Pliocene is very limited. The United States Geological Survey PRISM (Pliocene Research, Interpretation, and Synoptic Mapping) palaeoclimate reconstruction shows winter sea ice between 4 and 6° latitude farther south than present and minimal summer sea ice during the warm mid-Piacenzian Stage (3.264-3.025 Ma; Dowsett 2007). This was based on the extrapolation of increased SSTs from a southward shift of the polar fronts, seen in Southern Ocean diatom records (Barron 1996). Reduced sea ice during Pliocene interglacial periods has been inferred from the presence of fast-growing Chlamys bivalves due to their shell morphology and substrate preference (Jonkers 2003; Berkman et al. 2004), detailed morphological and geochemical analysis of Austrochlamys (Williams et al. 2010) and Mean Annual Range of Temperature (MART) analysis from bryozoan colonies (Clark et al. 2010) (see Table 4 for more details). Enhanced opal depositional rates, which have been linked to an increase in biological productivity between ~5 and 3.1 Ma, also suggest a reduction in sea ice cover (Hillenbrand & Ehrmann 2005). In addition, the ANDRILL AND-1B core contains a thick diatomite unit with abundant diatoms that may indicate surface waters warmer than today (McKay et al. 2012). Along with TEX86 SST of up to 5°C this suggests the absence or restriction of sea ice throughout prolonged intervals (McKay et al. 2012). Whitehead et al. (2005) also suggested that winter sea-ice concentrations were up to 78% less, relative to modern conditions, based on the diatom Eucampia antarctica index (McCullagh and Nelder 1989) measured at Ocean Drilling Program (ODP) Sites 1165 and 1166.

7. Climate model simulations of Pliocene environmental conditions on the Antarctic shelf

The Pliocene Model Intercomparison Project (PlioMIP) represents the first coordinated multi-model study of a warmer-than-modern, significantly above pre-industrial atmospheric CO_2 (405 ppmv) palaeoclimate (Haywood et al. 2013). It includes 8 coupled atmosphere-ocean general circulation model (GCM) simulations of the mid-Piacenzian warm period, with boundary conditions taken from the PRISM3 (Pliocene Research, Interpretation and Synoptic Mapping) reconstruction (Haywood et al. 2011). The peak-averaged SST warming in this interval has been quantified from multi-proxy studies by the PRISM project (Dowsett et al. 2010), providing a large number of marine SST estimates. These are globally correlated, if non-contemporaneous, as a target for GCMs (Dowsett et al. 2012; Haywood et al. in press). Comparisons between synthesized late Neogene Antarctic shelf data and the multi-model SST estimates of PlioMIP Experiment 2, although not providing a direct data-model comparison, allow the SST estimates presented here to be placed in the wider context of our understanding of Pliocene warming.

Comparisons between the SST warming in each of the three regions represented in the marine palaeontological record of the Antarctic shelf and the PlioMIP climate simulations show a good concordance (Table 5). SSTs in both the northern Peninsula and the Prydz Bay regions show an excellent match (defined as overlapping temperature ranges) with the palaeontological estimates. Winter temperatures are little changed from modern (Fig 8), primarily as some seasonal sea-ice can be found in both regions. Summer SSTs in the models are significantly increased throughout the Southern Ocean and especially on the Antarctic shelf (Fig 8), where little sea-ice survives throughout the year. The exceptions to this enhanced summer warming are the southern Weddell Sea and, in a few models, the western Ross Sea. In these regions, the proximity of the remaining parts of the large EAIS and the relatively enclosed nature of the embayments mean that sea ice can survive all year.

This feature of the western Ross Sea does not seem to be reflected in the SST estimates from that region, with summer temperatures estimated to have been 4 C° to 5 C° warmer (McKay et al. 2012). The disparity between model and data may show the limitations of the comparisons that are currently possible. Although there may have been ice-free periods at the drill site during the Piacenzian, this interval is characterised by a series of erosional horizons, showing that the grounded ice sheet periodically advanced through this site. In the early

Pliocene, (seasonally) open-water conditions lasted through a series of orbital cycles. Best estimates suggest that these ice-free periods in the western Ross Sea lasted for hundreds of thousands of years in intervals between 4.5 and 3.4 Ma (Scopelliti et al. 2013). Although many circumstances could have caused cooling between the early and late Pliocene (e.g. Sjunneskog and Winter 2012), it has been shown that the closure of the Panama Seaway during the Pliocene would cool the western Ross Sea region by up to 2°C (Lunt et al. 2008). The mean annual SST warming reconstructed from proxy analyses on Pliocene marine sections in onshore outcrops also seems to show slightly more warming than is produced in the PlioMIP models. However, the dating of these deposits is not very well constrained and it is possible that they also refer to warmer periods before the mid-Piacenzian interval.

Antarctic Region	Large Neogene Warming Parameter	Palaeontological Warming Estimate (C°)	PlioMIP Warming (C°)
NE. Peninsula	MASST	-	1 – 6
	SSTmin	1	0 – 5
	SSTmax	0 – 8	2 – 8
Prydz Bay	MASST	2 – 7	0 – 4
	Minimum SST	0	0 – 3
	Maximum SST	2 – 4	0 – 5
W. Ross Sea	MASST	3	0 – 2
	Minimum SST	4 – 5	0 – 1
	Maximum SST	-	0 – 5

Table 5 Comparison between palaeontological proxy estimates of late Neogene Antarctic shelf warming and regional estimates from the PlioMIP Experiment 2 ensemble (Haywood et al. 2013). Palaeontological warming estimates have been standardised to be warming above observed modern SSTs (Table 1). An excellent match is defined by an overlapping of these ranges.

7.1 Possible Explanations of Warmer Antarctic Coastal Conditions during the Late Neogene

The late Neogene was a time of generally warmer global temperatures than present, with persistent fluctuations associated with variations in the Earth’s orbit that were of lower amplitude than in the Pleistocene (e.g. Lisiecki and Raymo 2005; Naish et al. 2009). During warmer intervals, for example the mid-Piacenzian warm period (defined by PRISM as 3.246 to 3.025 Ma; Dowsett et al. 2010), the average global surface temperature was 2 C° to 3 C° (Jansen et al. 2007) warmer than pre-industrial. There are many hypotheses about what conditions may have driven this warming and it is likely that there was more than one forcing mechanism at work. Suggestions for the cause of Pliocene warmth include loss of permanent

Arctic sea ice (Raymo et al. 1990), increased atmospheric CO_2 concentrations (Crowley 1991), lower mountain altitudes in North America and Tibet (Rind and Chandler 1991), increased northwards heat transport by the oceans (Dowsett et al. 1992), closure of the Panama ocean gateway (Haug and Tiedemann 1998), opening of the Indonesian seaway (Cane and Molnar 2001), changes in global ice cover (Haywood and Valdes 2004), permanent El Niño in the tropical Pacific (Wara et al. 2005) and vegetation feedbacks (Haywood & Valdes 2006), perhaps associated with changes in terrestrial biomass (Mudelsee and Raymo 2005). The most widely accepted mechanisms are linked to elevated atmospheric CO_2 concentrations leading to greenhouse warming or changes in ocean heat transport (Haywood et al. 2007).

An increase in meridional ocean heat transport was suggested as being a major contributing factor to Antarctic warming in the Pliocene, based on records of North Atlantic SST (Dowsett et al. 1992; Kwiek and Ravelo 1999; Ravelo and Andreasen 2000; Haywood et al. 2000). Despite the extensive data supporting this process it has not always been replicated by GCMs, which instead show some tropical warming with more moderate polar amplification (Zhang et al. 2013).

The general consensus for atmospheric CO_2 levels during the Pliocene is for a range from 360 to 400 ppmv, significantly higher than pre-industrial values of 280 ppmv, but comparable to present. These values have been established using a number of different proxies including $\delta^{13}C$ ratios of organic compounds and boron isotopes in calcareous foraminifera tests (Raymo and Rau 1992; Raymo et al. 1996; Pagani et al. 2009; Seki et al. 2010; Bartoli et al. 2011) and the stomata density on fossil plant leaves (Van der Burgh et al. 1993; Kürschner et al. 1996). Other studies have conversely suggested that concentrations were lower and closer to that during the Last Glacial Maximum (Pearson & Palmer 2000).

Naish et al. (2009) provide evidence from the AND-1B sediment core that the WAIS was directly influenced by orbitally-induced oscillations causing periodic collapses of the ice sheet, resulting in open water conditions in the Ross embayment. They suggest that obliquity cycles may regulate the southward distribution and upwelling of Circumpolar Deep Water (CDW), inducing melting of the WAIS. Collapse of ice shelves buttressing the EAIS probably resulted in glacier flow acceleration and retreat of this ice sheet, too (Hill et al., 2007). The widespread ice-sheet melting may have led to feedbacks that drove much of the

Neogene ocean warming seen in the fossil records (cf. Joughin and Alley 2011). The exact details of such feedbacks, however, are still unknown (Joughin and Alley 2011).

More speculative theories include Southern Ocean circulation changes such as increases in eddies and jets within the main current of the ACC (Thompson 2008) and the southward migration of the Antarctic Polar Frontal Zone (APFZ) (Whitehead and McMinn 2002). Eddies can be up to 500 km in diameter and persist for months, sometimes transporting warmer water into the Southern Ocean. Hogg et al. (2008) suggest that pole ward eddy heat flux, increased by greater wind stress, has a relatively large effect on the overall temperature in the Southern Ocean and may have contributed to the recent warming. During warmer intervals in the Pliocene it has been suggested that the APFZ migrated south by approximately 900 km, the temperature gradient was significantly shallower, or there was some combination of the two (Whitehead and McMinn 2002). This is thought to be responsible for a temperature increase of 3 C° to 4 C° between 55° and 60°S (Barron 1996; Bohaty and Harwood 1998; Whitehead and McMinn 2002). However, a southward shift in the APFZ is difficult to achieve, especially in regions such as the Kerguelen Plateau due to topographical limitation preventing such a shift (Barker and Thomas 2004).

8. Conclusions

Fossil proxy data have been sourced from a range of shallow marine sedimentary deposits of late Neogene age around the margins of the Antarctic continent, but the database remains sparse and fragmentary both from a temporal and spatial context. Many proxies evaluated in this paper suggest that at times during the late Neogene, SSTs were significantly warmer than at present. Proxies for bivalves (e.g. A. anderssoni in Williams et al. (2010) or indirectly using species tolerances (e.g. dolphin fossils in Quilty et al. (1990), suggest that there was a reduction in sea ice during warmer intervals. Though the temporal dataset is fragmentary, there does not seem to be significant evidence for temperature trends either towards a long-term cooling or warming through the Pliocene. There is, however, evidence for periods of warmth (4.6 to 3.3 Ma) that can be seen in different regions at approximately the same time (assuming the ages of the deposits are reliably constrained). Comparisons between the SST warming in each of the three regions represented in the marine palaeontological record of the Antarctic shelf and the PlioMIP climate simulations show a good concordance (Table 5), particularly SSTs in both the northern Antarctic Peninsula and Prydz Bay regions. Winter temperatures are very similar to modern. Differences between model and proxy data (e.g. in the Weddell Sea and western Ross Sea) reveal the temporal and geographical limitations of the comparisons that are currently possible. The most accepted mechanisms for causing such warming are an increase in atmospheric CO_2 concentrations leading to greenhouse warming and/or an increase in meridional ocean heat transport to the poles.

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Figures

Fig 1 Maximum (August) and minimum (February) sea-ice extent. Solid line: minimum sea ice extent in 2012, dashed line: maximum sea ice extent in 2011. Winter sea ice has changed very little but summer sea ice has increased since the late 1970s. (Free data from the National Snow and Ice Data Center (Fetterer et al.; www.nsidc.org); data analysis and diagram courtesy of Robert Cooper, British Geological Survey)

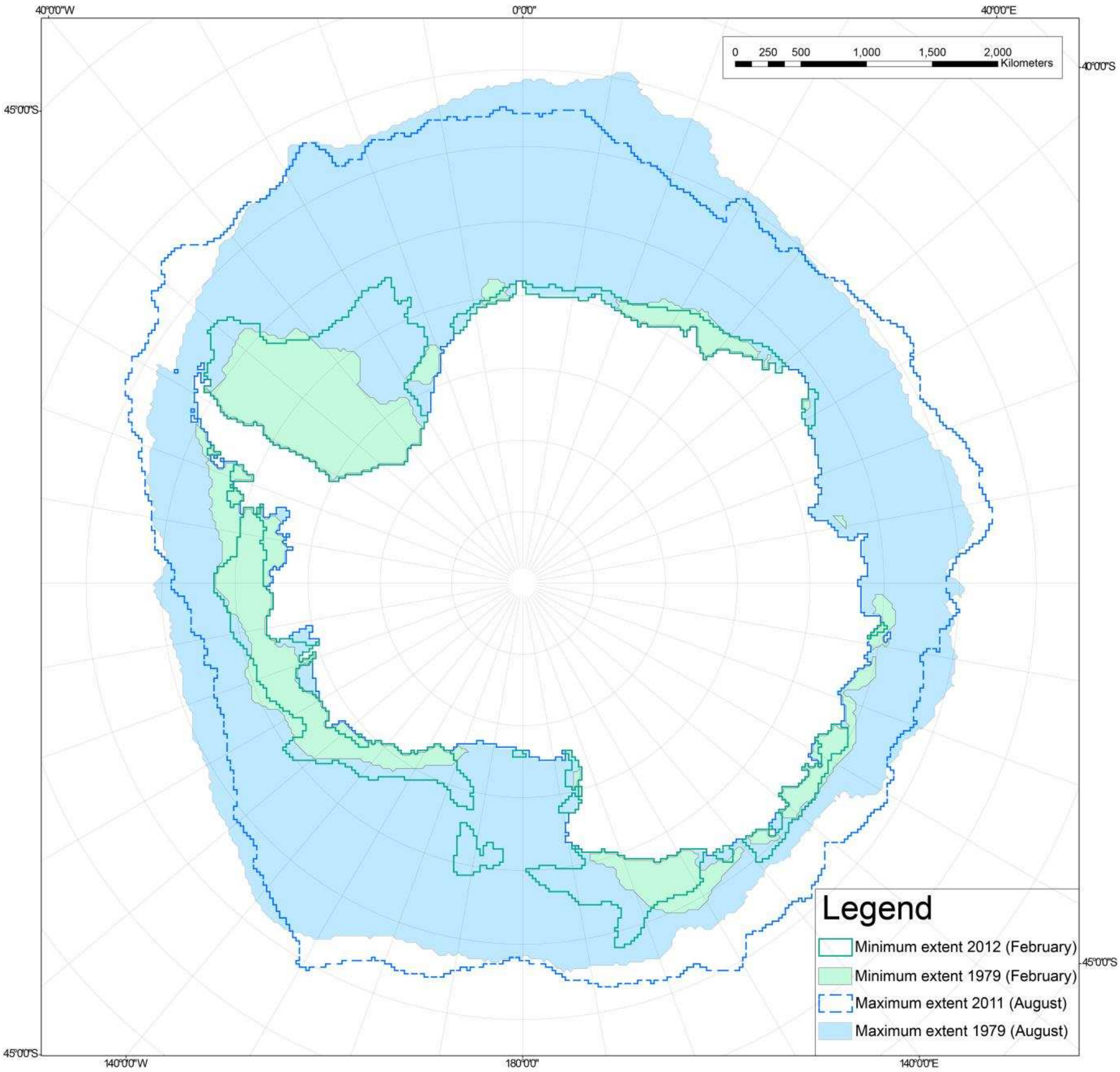

Fig 2 Onshore geographical distribution of late Neogene shallow marine macro-fossil-bearing sites of the Antarctic. EAIS – East Antarctic Ice Sheet, WAIS – West Antarctic Ice Sheet, APIS – Antarctic Peninsula Ice Sheet. Localities summarised from: Jonkers (1998). Schematic stratigraphic columns (not to scale) for each main macro-fossil-bearing locality showing generalised lithology and ages where known (see Fig 2 for more age details). Cockburn Island: (Jonkers and Kelley 1998), James Ross Island: (Pirrie et al. 1997; Smellie et al. 2006; Nỳvlt et al. 2011), Larsemann Hills: (Webb 1974; Quilty et al. 1990; McMinn and Harwood 1995), Vestfold Hills: (Colhoun et al. 2010; Quilty et al. 2000; Whitehead et al. 2001; Whitehead et al. 2004; Whitehead et al. 2006b), Wright Valley: (Prentice et al. 1993), McMurdo Sound: (Eggers 1979; Leckie and Webb 1979)

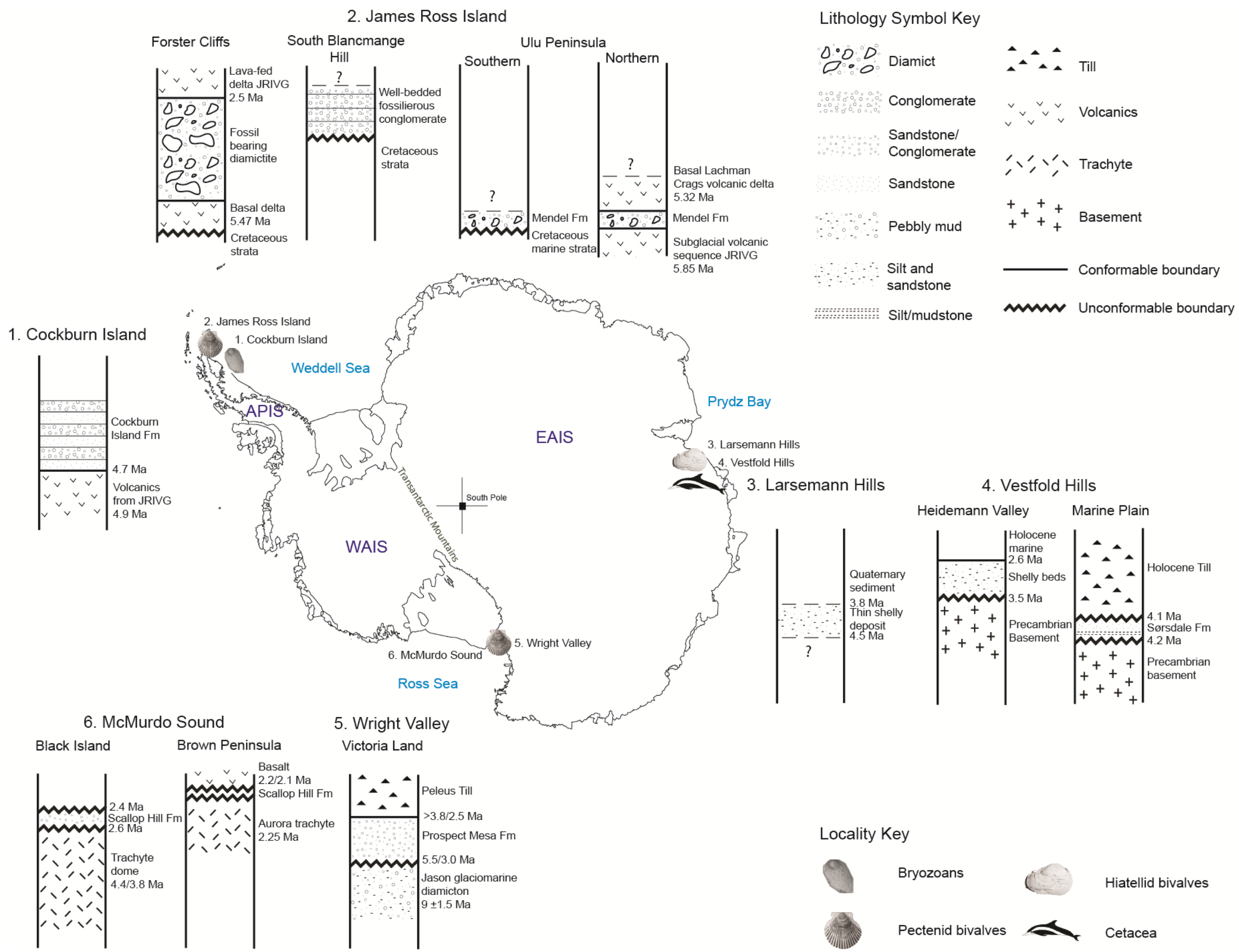

Fig 3 Stratigraphic setting of late Neogene successions that yield shallow marine fossils. Formations are placed in their most likely chronostratigraphic position, but note that the grey arrows denote the degree of stratigraphic uncertainty. Age of boundary shown above or below line. Only sedimentary deposits have been shown as these contain the fossil material discussed. Correlated using data from (Gradstein et al. 2012) and benthic $\delta^{18}O$ isotope stack LR04 (Lisiecki and Raymo 2005). Data for the Antarctic Peninsula is from Smellie et al. (2006); Nỳvlt et al. (2011), Cockburn Island: (Jonkers and Kelley 1998; Pirrie et al. 2011; Levy et al. 2012), Vestfold Hills: (Quilty et al. 1990; Quilty et al. 2000; Whitehead et al. 2001; Whitehead et al. 2004; Whitehead et al. 2006b; Cody et al. 2008; Colhoun et al. 2010), Larsemann Hills: (Webb 1974; Quilty et al. 1990; McMinn and Harwood 1995), McMurdo Sound: (Eggers 1979; Leckie and Webb 1979), Wright Valley: (Haq et al. 1987; Baldauf et al. 1991; Harwood et al. 1992; Prentice et al. 1993; Cody et al. 2008)

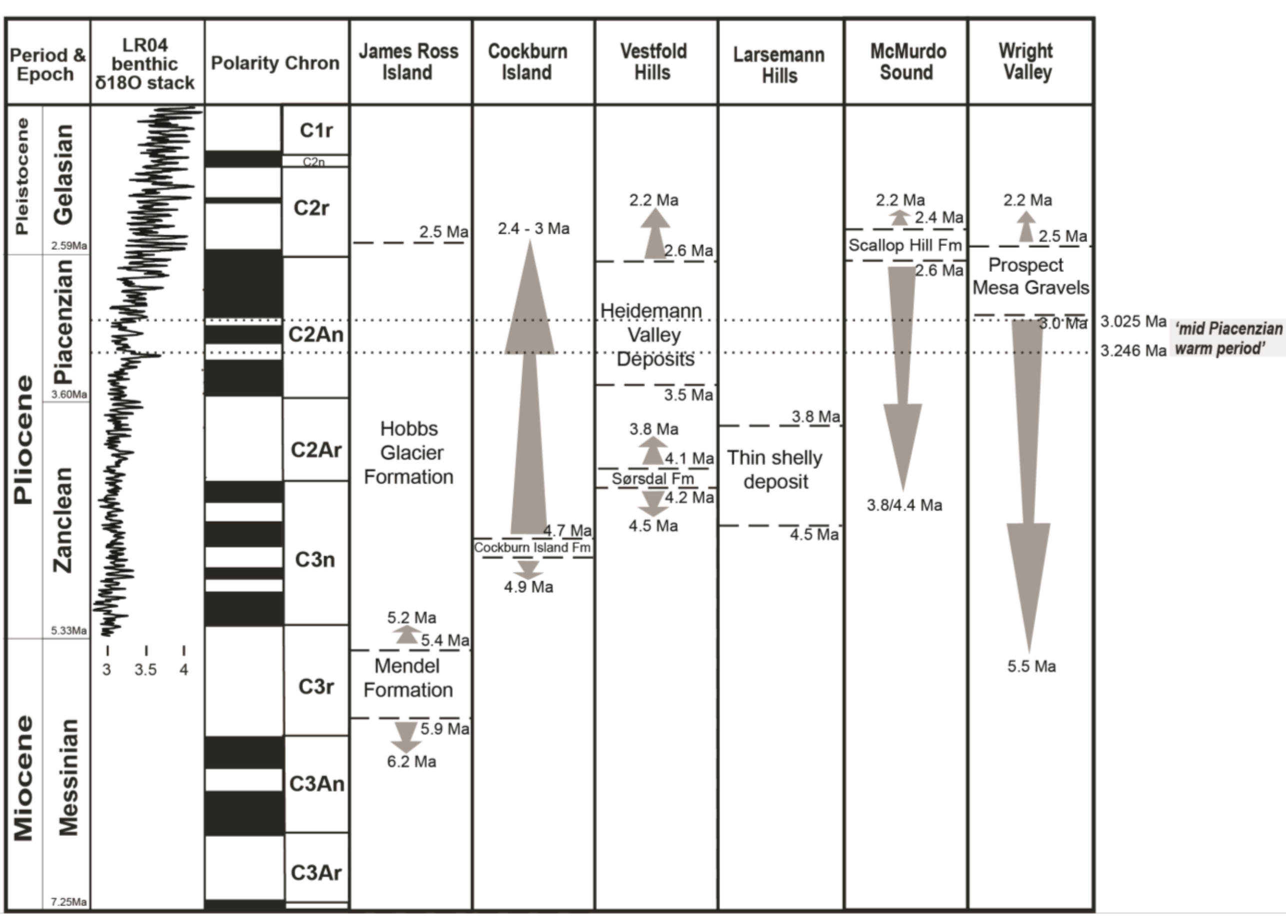

Fig 4 Marine Plain. (a) Coarse Quaternary material overlying the Pliocene succession (Scale: height of circled Adelie penguins ~60 cm), (b) general environment, ice-covered pond in the foreground is the site of the first whale fossil discovery (Scale: circled tents in background)

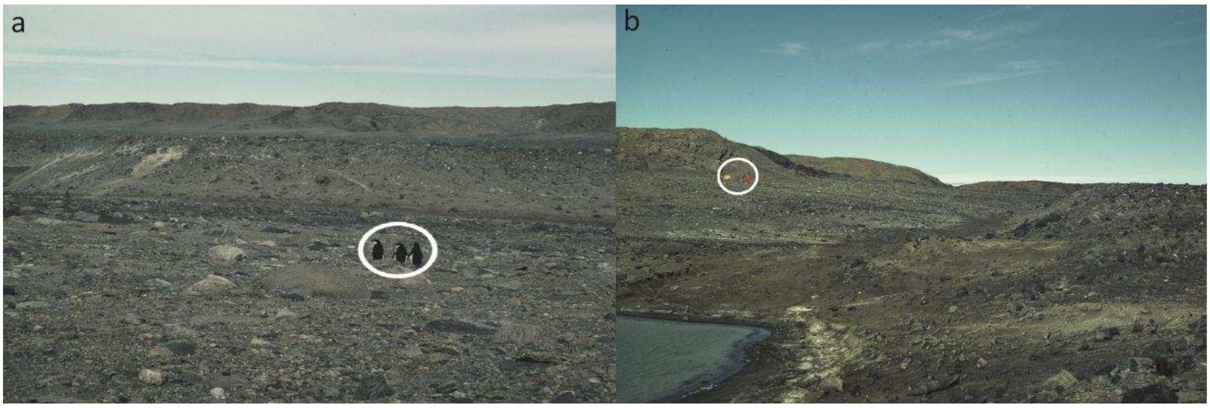

Fig 5 Map of Pliocene fossil-bearing deposits in the Prydz Bay region. Larsemann Hills, Marine Plain and Heidemann Valley (in bold). (Courtesy of Henk Brolsma, Australian Antarctic Division)

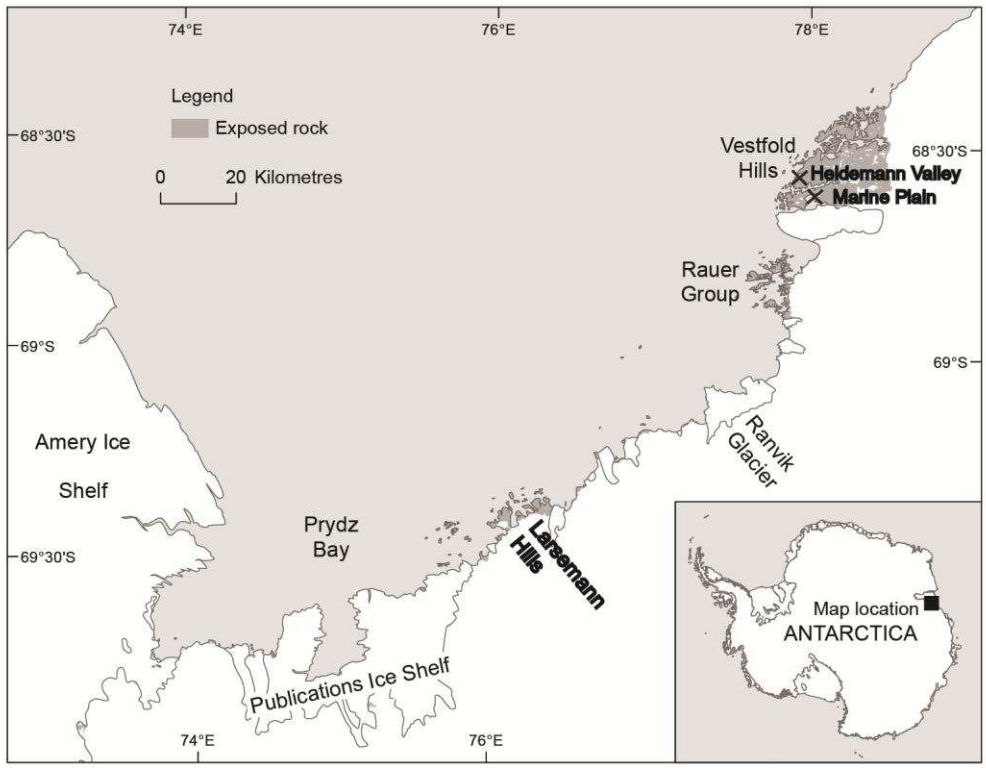

Fig 6 (a) Diamict overlain by volcanic deposits at Forster Cliffs (Loc D6.404; 3 in this paper) (people circled for scale), (b) shell fragment (arrowed) within diamict at Cascade Cliffs (Loc D6.405; 6 in this paper) (scale bar is 2 cm), (c) scallop shells found stacked between volcanic clasts at South Blancmange Hill (Loc D6.407; 4 in this paper), (d) stacked glacigenic debris flows containing macrofossil material at South Blancmange Hill (scale bar is 1 m), (e) articulated Austrochlamys shell suggesting little or no reworking at South Blancmange Hill. (Nelson et al. 2009)

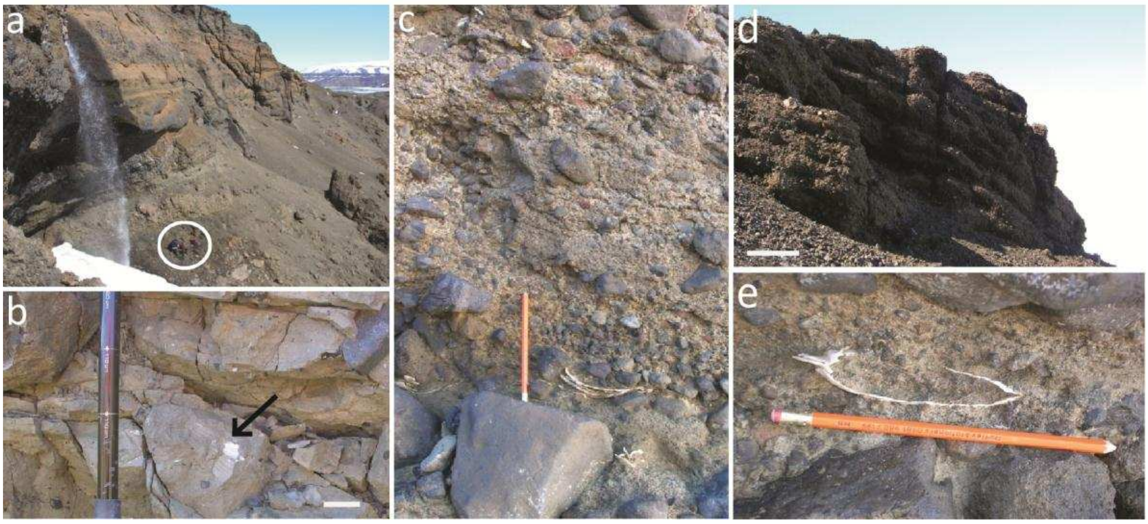

Fig 7 (a) A bryozoan colony encrusting a clast of the James Ross Island Volcanic Group from James Ross Island (Forster Cliffs, locality 3 in this paper), Antarctic Peninsula (Clark et al. 2010), (b) an *Austrochlamys* bivalve from Cockburn Island (Williams et al. 2010). Scale bars are 2 cm

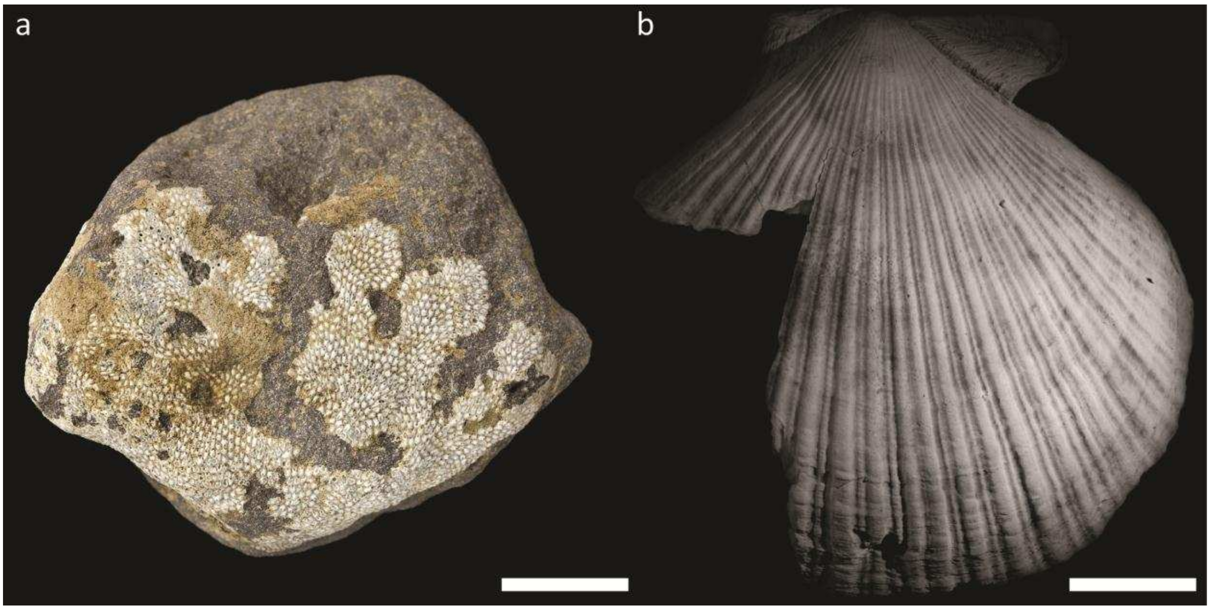

Fig 8 PlioMIP Experiment 2 multi-model mean Pliocene SST warming above pre-industrial around the Antarctic (i.e. Pliocene minus pre-industrial). (a) annual mean warming, (b) February warming and (c) August warming. Sea-ice extent is not shown. For full description of PlioMIP experiments and simulations see Haywood et al. (2013)