Available online at www.sciencedirect.com
SCIENCE @ DIRECT®
Earth-Science Reviews 66 (2004) 143–162
EARTH-SCIENCE REVIEWS
www.elsevier.com/locate/earscirev

Origin, signature and palaeoclimatic influence of the Antarctic Circumpolar Current

P.F. Barker[a,*], E. Thomas[b,c]

[a] 25 Church St., Great Gransden, Sandy, Beds SG19 3AF, UK
[b] Department of Earth and Environmental Sciences, Wesleyan University, Middletown, CT 06457, USA
[c] Center for the Study of Global Change, Department of Geology and Geophysics, Yale University, New Haven, CT 06520-1809, USA

Accepted 27 October 2003

Abstract

The Antarctic Circumpolar Current (ACC) is today the strongest current in the world's ocean, with a significant influence on global climate. Its assumed history and influence on palaeoclimate, while almost certainly equally profound, are here called into question. In this paper, we review 30 years of accumulated data, interpretation and speculation about the ACC, deriving mainly from DSDP and ODP drilling in the Southern Ocean. For most of this time, a conventional view of ACC development, signature and influence has held sway among palaeoceanographers and marine geologists. In this view, the ACC began at about 34 Ma, close to the Eocene–Oligocene boundary, the time of onset of significant Antarctic glaciation and the time of creation of a deep-water gap (Tasmanian Seaway) between Australia and Antarctica as the South Tasman Rise separated from North Victoria Land. This is the "smoking gun" of synchroneity. The Southern Ocean sediment record shows a latest Eocene development and subsequent geographic expansion of a siliceous biofacies, its northern limit taken to indicate the palaeo-position of the ACC axis. In addition, the ACC was considered to have caused Antarctic glaciation by isolating the continent within a cold-water annulus, reducing north–south heat transport. A different (and later) date for Antarctic–South American opening ("Drake Passage") was proposed, but the timing of ACC onset there was disputed, and the simple story survived. Recent developments, however, call it into question. Modern physical oceanography shows that all or most of present-day ACC transport is confined to narrow jets within deep-reaching circumpolar fronts, and numerical modelling has suggested that a steady reduction in greenhouse gas concentration through the Cenozoic could cause Antarctic glaciation, with or without a contribution from ocean circulation change. The rapidity of Antarctic glacial onset at the Eocene–Oligocene boundary and coeval creation of a deep-water gap south of Tasmania both survive but, in light of the new information, the presence of a siliceous biofacies cannot be claimed as evidence of the existence of a continuous, deep-reaching oceanic front and therefore of an ACC, and the possibility arises that cool and cold sea-surface temperatures were effects of Antarctic glaciation rather than evidence of a major contributor to its cause. In considering future work, we emphasise the importance of additional information from ancillary fields—better definition of the necessary and sufficient properties of oceanic fronts, additional determinations of Cenozoic atmospheric pCO_2 and further developments in models of Antarctic glaciation—but also suggest the way forward in marine geology. Our knowledge of the development and palaeoclimatic significance of the ACC will be best served by grain-size studies of bottom current strength at selected locations, and geochemical or mineralogical studies of clays and IRD as a way of examining provenance and therefore surface and bottom current directions and the existence of interocean connections.

* Corresponding author. Tel.: +44-1-767-677081.
E-mail address: pfbarker@tiscali.co.uk (P.F. Barker).

0012-8252/$ - see front matter
doi:10.1016/j.earscirev.2003.10.003

Studies of biogenic assemblages within the same sediments may be able to recover a value for the many such studies undertaken in the past and interpreted, probably erroneously, as evidence for an ACC. Mainly in view of the timing uncertainties, we propose the region south of South America as the best initial focus of future investigation.

Keywords: Antarctic Circumpolar Current; ACC; Palaeoclimate; Palaeocirculation; Oceanic fronts; DSDP; ODP

1. Introduction

The Antarctic Circumpolar Current (ACC) is an influential component of the present-day global climate system. The largest ocean current in the world, it averages about 130 Sv (10^6 m^3 s^{-1}) volume transport through Drake Passage. It is considered to be mainly or entirely wind-driven, but extends to the seabed in most places. It is closely associated with one or more oceanic frontal systems, and an approximation to the mean position of the ACC "axis" is given by the locus of the Polar Front (PF) in Fig. 1a. The sea-surface expression of most of the frontal systems is a sharp southward temperature drop. The association of

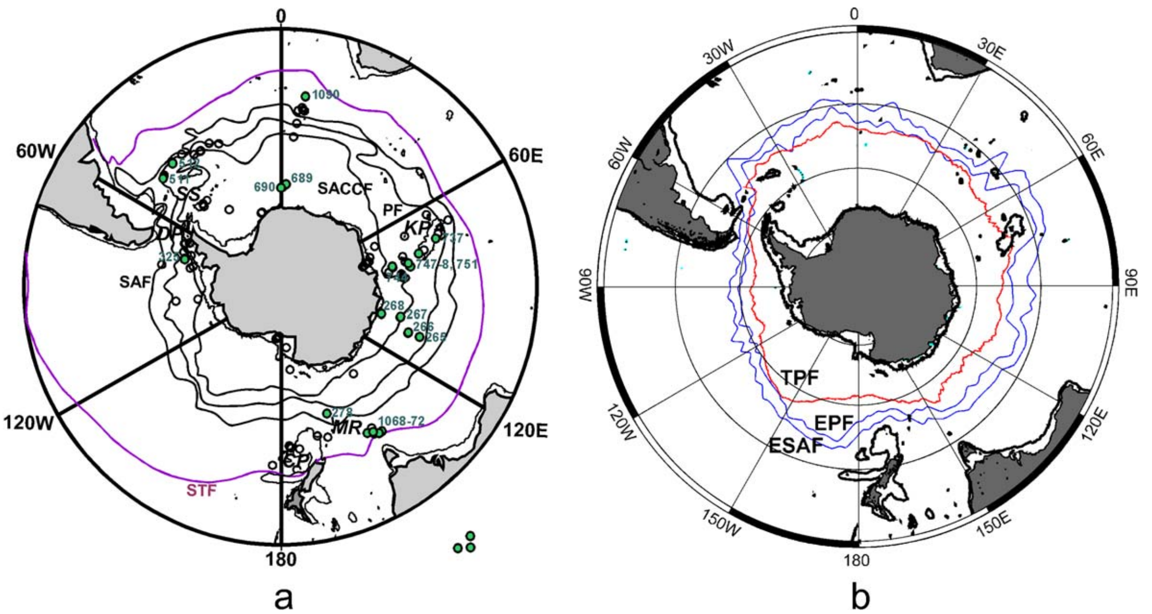

Fig. 1. (a) Paths of Southern Ocean fronts (Sub-Tropical Front STF, Sub-Antarctic Front SAF, Polar Front PF, and southern ACC front SACCF) determined by Orsi et al. (1995), from water property variations measured at stations on approximately 100 shipboard transects accumulated over 30 years. The STF, not continuous around Antarctica, is purple. Also marked are the Kerguelen Plateau (KP), Macquarie Ridge Complex (MR), Scotia Sea (SS) and Drake Passage (DP), and the 1000-m isobath. DSDP and ODP drill sites in the Southern Ocean, within the realms of the ACC fronts and to the south, are open circles, filled and labelled green if mentioned specifically in the text. (b) Comparison of the mean positions of the Polar Front and Sub-Antarctic Front (EPF and ESAF) determined by Gille (1994) from satellite-derived sea-surface elevation variation, and of the Polar Front (TPF) determined by Moore et al. (1999) from satellite-derived measurements of sea-surface temperature. The 1000-m isobath is shown. Satellite and ship (a) measurements are complementary. Note that mean positions of the Polar Front (PF, TPF, EPF) usually coincide, but may differ: where the shallow TPF is displaced from the others (as near the Kerguelen Plateau near 70°E), the shallow front may cross the shallow plateau, whereas its deep component requires the deep path to the north. The satellite-based sea-surface elevations (Gille, 1994) measure eddy kinetic energy, and where their recomputed mean (EPF) departs from the others, eddies may have spun off the front upstream, the main front being guided by seabed topography (such as a mid-ocean ridge) and thus positionally invariant and undetectable by the elevation-based technique. In addition, the TPF positions are inevitably slightly offset southward, marking the poleward edge of a frontal zone that may be 40–70 km wide (Moore et al., 1999).

the ACC with sea-surface temperature change and consequent change in planktonic biotic assemblages, and its extension to the seabed, have suggested there should be a sedimentary record of its past variation (e.g., papers in Prothero and Berggren, 1992).

The ACC is important for an understanding of global palaeoclimate and ocean circulation. It developed at some time during the Cenozoic, as a series of deep-water gaps opened around Antarctica, and has been widely viewed as having reduced meridional heat transport, isolating the continent within an annulus of cold water and thus being at least partly responsible for Antarctic glaciation (e.g., Kennett, 1977). It is today almost the only means of exchange of water between the major oceans, and its onset may also have significantly modified Northern Hemisphere climate (Toggweiler and Bjornsson, 1999). Some consider the concentration of greenhouse gases (principally CO_2) in the atmosphere to have been the main influence on palaeoclimate (see, for example De Conto and Pollard, 2003a,b), and the change in meridional heat transport associated with ACC onset to have been insignificant (e.g., Huber and Sloan, 2001), but an understanding of ACC variation remains important, so that the relative contributions of ocean circulation and greenhouse gases as climate influences may be better understood.

The time of onset of the ACC is uncertain, and abundant speculation exists about its onset and effects. As knowledge of modern ocean circulation and the uses of the geological record has developed, and as the palaeoceanographic data base has grown, so speculation has been refined, but the uncertainties in ACC onset and subsequent variation remain significant. Here, we review the field, with the aims of reassessing the evidence of ACC behaviour already published and identifying the best approach to determining ACC history.

Before reviewing geological data and speculation, we describe progress in our understanding of the main controls or influences on, and indicators of, ACC onset and development: regional tectonic evolution, the modern physical oceanography of the Southern Ocean (including numerical modelling) and Antarctic glaciation. Progress in all fields of research in this region has been considerable over the past 30 years, although slowed by its logistic remoteness, and the influence of such progress has been profound. However, some of the older geological observations have a validity beyond the limitations imposed on their interpretation by the context of the time.

2. Regional tectonics and topography

The ACC as we know it today could not have existed in the absence of a continuous deep-water path. Neither a partial circumpolar path nor one with a very shallow barrier (in particular, one crossing a marine continental shelf) would permit a significant circumpolar circulation (see Section 3, below). This assumption is convenient tectonically, since past variations in the relative distributions of land and continental shelf (barring or permitting shallow-water flow) are much more difficult to determine than are ocean floor ages, but there is little understanding of the nature of ocean circulation with a partial path or shallow barrier, prior to creation of a continuous deep-water circumpolar path.

By the late 1970s, the age of ocean floor around the Southern Ocean was almost completely known. Barker and Burrell (1977) showed that the deep-water gateway south of South America opened last, later than those south of Australia and the Macquarie Ridge and north of the Kerguelen Plateau, all of which are now within the path of the ACC. To the extent that the ACC has been wind-driven, meridional migration of the circumpolar westerly winds could have dictated that other pathways were more important in the past. The present dramatic northward diversion of the ACC north of the Kerguelen Plateau, in preference to a more direct southerly path over older but shallower ocean floor south of the plateau, may be a case in point (see Wei and Wise, 1992), but the lateness of the critical opening south of South America makes the present path of the ACC in almost all sectors the most likely, throughout its life.

South of South America, both regional tectonics and the present path of the ACC are complicated. The ACC flows east through Drake Passage, then most of it veers north, through gaps in the North Scotia Ridge (see Barker, 2001). The age of ocean floor in Drake Passage (at least 28 Ma, probably older) is not disputed, but more recent work retains the age ambiguity of ACC onset. Barker and Burrell (1977) argued that because of prominent ridges extending along part of the Shackleton Fracture Zone in Drake Passage, a deep-water pathway would not have been created until they cleared, at 22 ± 2 Ma (timescale of Cande and Kent, 1995), close to the Oligocene–Miocene boundary. The prior existence of those ridges was based on the contrasting character of sedimentation in their lee. Barker and Burrell (1982) suggested also the possibility of further delay in ACC development, because of relict obstruction to the east by continental fragments and subduction-related volcanoes around the Scotia Sea. Barker (2001), based on a reconstruction of regional tectonic evolution, pointed to the unknown but crucial natures and origins of Davis Bank and Aurora Bank, now lying along the North Scotia Ridge but possibly having formed a barrier to ACC development after the Shackleton Fracture Zone ridges had separated (see Fig. 2). He noted also that the floor of the Scotia Sea itself would have been rougher and shallower than today during the early stages of opening, which may have led to additional delay in ACC development, and suggested a correlation between growth of the ACC and mid-Miocene cooling. Barker and Burrell (1977) pointed out that a widening deep-water path would eventually reach the state at which further widening would no longer lead to an increase in ACC transport.

In contrast, Lawver et al. (1992) suggested a deep-water gap probably existed at Drake Passage before 30 Ma, on the basis of major plate (South American–Antarctic) motion, but allowed the possibility of delayed ACC onset because of a more compact arrangement of continental fragments to the east, citing Barker and Burrell (1982). Subsequently, Lawver and Gahagan (1998) posited the early Oligo-

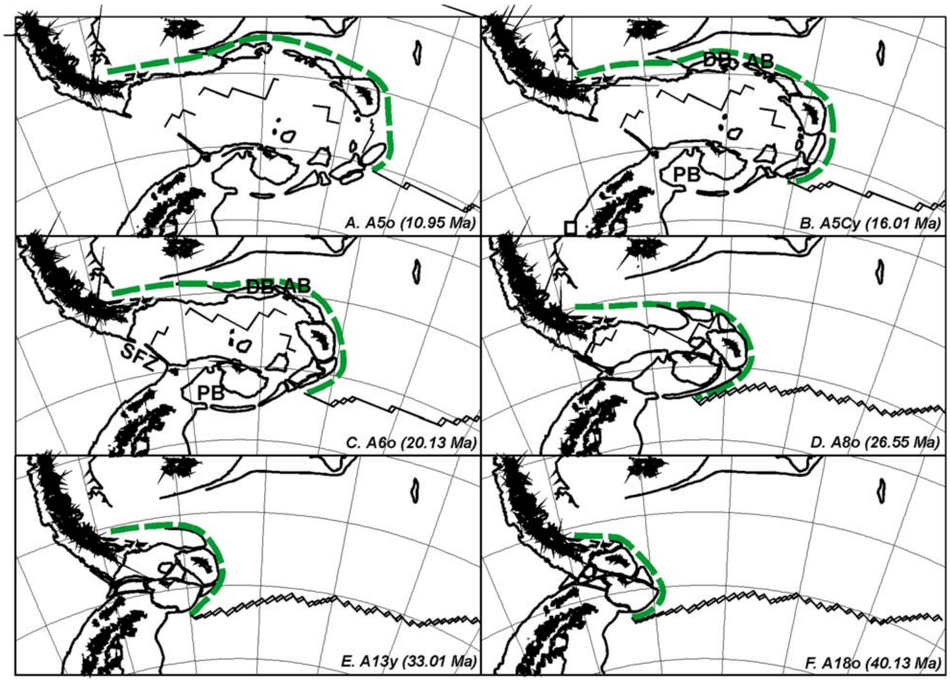

Fig. 2. Reconstructions of Scotia Sea tectonic evolution at six times (A–F) within the last 40 m.y., drawn with respect to a fixed South America (Falkland Islands, Falkland Plateau, Atlantic coast of Argentina), adapted from Barker (2001). Magnetic anomaly ages from Cande and Kent (1995). Major components are identified by coastline and 2000 m isobath. Also showing schematic trench (grey (green in electronic version) dashed line) and trace of spreading centres (within Scotia Sea and at SAM-ANT plate boundary) where known. The reconstructions suggest that a deep-water passage capable of use by an ACC resembling that of the present day was not produced until the ends of the Shackleton Fracture Zone ridges [SFZ marked in (C)] cleared at about 22 Ma, and possibly not until several million years later, depending on the composition and palaeodepth histories of Davis Bank and Aurora Bank along the North Scotia Ridge [DB and AB in (B) and (C)] and the sensitivity of the ACC to ocean floor depth and roughness within Drake Passage and the Central Scotia Sea. Powell Basin is also marked (PB) in (B) and (C).

cene creation of a deep-water pathway in the Drake Passage region via a small basin (Powell Basin) within the South Scotia Ridge. However, Eagles and Livermore (2002) have determined younger ages than were assumed by Lawver and Gahagan (1998) for Powell Basin opening and conclude (with Barker, 2001) that continental parts of the South Scotia Ridge would in the past have barred a deep connection between Powell Basin and the Pacific, as they do today. Most recently, once more using mainly the major plate separations, Lawver and Gahagan (2003) argued that Drake Passage opened for deep-water circulation via Powell Basin before 28 Ma and proposed a 31-Ma onset for the ACC.

Other options are less widely accepted. It has been suggested that a deep-water pathway existed between East and West Antarctica during the early Cenozoic, on the basis of inferred palaeo-depths of Ross Sea sediments (Webb, 1979; see also Nelson and Cooke, 2001; Lawver and Gahagan, 2003), but this seems unlikely given the continental crustal structure beneath large areas of the southern Ross and Weddell Seas and intervening region (e.g., Bentley and Clough, 1972; Cooper et al., 1995; Hübscher et al., 1996), absence of Cenozoic plate motion in the Weddell Sea region and lack of evidence in seismic reflection profiles of strong bottom currents (although a shallow-water pathway is not ruled out). The effect on ACC development of the conjecture of DeWit (1977), that the Central Scotia Sea contains captured Mesozoic ocean floor, has not been assessed.

Most of those basing speculation about ACC onset on the tectonic evidence have assumed additionally that ocean circulation will quickly exploit a newly created deep-water path. Certainly, since the present ACC is largely or entirely wind-driven, and the driving westerlies are generally considered an indirect consequence of the meridional atmospheric temperature gradient, winds (and thus the ACC) have probably intensified through the Cenozoic as the poles have cooled. The suggestion of Lawver and Gahagan (1998) that a deep-water pathway may not have been exploited immediately by a proto-ACC is hard to accept, because the complete absence in the early Cenozoic of an equator-pole temperature gradient (which might have prevented the driving winds) is contrary to the evidence of a cold Antarctica at that time (see below and Huber and Sloan, 2001).

3. Physical oceanography

Knowledge of the present-day ACC has grown with time (see, for example, Deacon, 1937; Tolstikov, 1966; Nowlin and Klinck, 1986; Orsi et al., 1995; Moore et al., 1999). Shipborne data have accumulated and are now complemented by systematic satellite measurements of sea-surface temperature and elevation (Fig. 1b). The ACC is the most powerful ocean current on earth, having been estimated at 130 Sv (10^6 m^3 s^{-1}) relative to 2500 m in Drake Passage (Whitworth and Peterson, 1985; see also Ganachaud and Wunsch, 2000; Cunningham et al., 2003). It is generally considered to be largely or entirely wind-driven (by the belt of westerly winds at 45–55°S), but flow continues (though usually attenuated) to depth, in most places to the sea bed. However, the concept of an ACC "axis" (implying a single, broad, Gaussian or similar, speed distribution across the ACC) is of limited value since the bulk of ACC transport is now considered (e.g., Nowlin and Klinck, 1986; Gille, 1994) to occur within deep-reaching, narrow jets associated with two or more distinct fronts, the Polar Front (PF) and the Sub-Antarctic Front (SAF, north of the PF), which are continuous around the continent. Orsi et al. (1995) identified a third, southern ACC front, also deep-reaching and continuous (Fig. 1a). Other fronts (such as the Sub-Tropical Front—STF in Fig. 1a) are not continuous around the continent, so do not contribute to the ACC. The fronts are usually water mass boundaries: the subsurface Polar Front is where Antarctic surface water sinks, the basis of its original identification and definition (Deacon, 1933). The fronts meander where there is no constraining seabed topography (e.g., Moore et al., 1999), and mesoscale eddies and rings that include them are common, particularly downstream of topography (e.g., Legeckis, 1977; Gouretski and Danilov, 1994).

Surface expression of the fronts is usually a sharp southward temperature decrease, but in places this shallow expression is displaced from the boundary at depth (notably for the PF, to the south across the Kerguelen Plateau near 70°E, Fig. 1a). Generally, the separate effects of the fronts on sea-surface height can be distinguished, but in places they may merge (Gille, 1994). Causes of the development and maintenance of the fronts are poorly understood. Gille (1994) estimated that 40–70% of ACC transport occurred within the

PF and SAF, but it seems likely that a far higher proportion might be so carried within frontal zones, if Orsi et al.'s (1995) southern ACC front also has west–east transport (as measured by Heywood and King, 2002), and that total transport is greater if the water velocity below 2500 m is not zero, as much evidence now indicates. For example, many seismic reflection profiles crossing deep-water areas show nondeposition or severe reworking of sediments by bottom currents (e.g., Fig. 7 of Barker and Burrell, 1977), and several current meter moorings have measured strong (though varying) bottom currents directly. An example is given in Fig. 3, from a current meter mooring (Mooring 11) in the Central Scotia Sea, described by Pudsey and Howe (2002). The site is at 3660 m water depth, close to the southern ACC front of Orsi et al. (1995).

These and other data from the Southern Ocean (see, for example, other moorings described by Pudsey and Howe, 2002) confirm the geographically focussed nature of fast flow, and the slow speed (scarcely above tidal speeds) of flow otherwise. For a total transport of 130 Sv over a mean water depth of 3 km, the ACC would sustain entirely focussed fast eastward flow (say 30 cm s^{-1} from the surface to full depth) across a width much less (only 140 km in this example) than the ACC width generally assumed within palaeoceanography.

Diffuse upwelling south of the PF complements the sinking of Antarctic surface water there and close to the Antarctic margin (see below). Although there is calcareous as well as siliceous biogenic primary production south of the PF (see, for example, Fig. 1 of Iglesias-Rodriguez et al., 2002), stimulated by the upwelled nutrients, calcareous tests are almost entirely dissolved within the water column, leading to a mainly siliceous residual biofacies in deep-sea sediments to the south (e.g., Goodell, 1973). Close to the fronts, primary production is generally high (see, for example, papers in Smetacek et al., 1997).

The other main component of Southern Ocean circulation is a discontinuous west-flowing, wind-driven current close to the continent and south of the ACC (Deacon, 1937), which develops where possible into clockwise gyres such as the Weddell Gyre. The Weddell Gyre is enhanced by the production of dense, cold and salty bottom water in the southern Weddell Sea, by a combination of ice-related processes: increased salinity of surface waters because of sea-ice formation and super-cooling by contact with the cold underside of a floating ice shelf (e.g., Brennecke, 1921; Foster and Middleton, 1980; Foldvik and Gammelsrod, 1988). Such water mass modification, by one or both processes, probably will have taken place since the onset of Antarctic glaciation. The very dense bottom water so produced today is termed Weddell Sea Bottom Water (WSBW) and

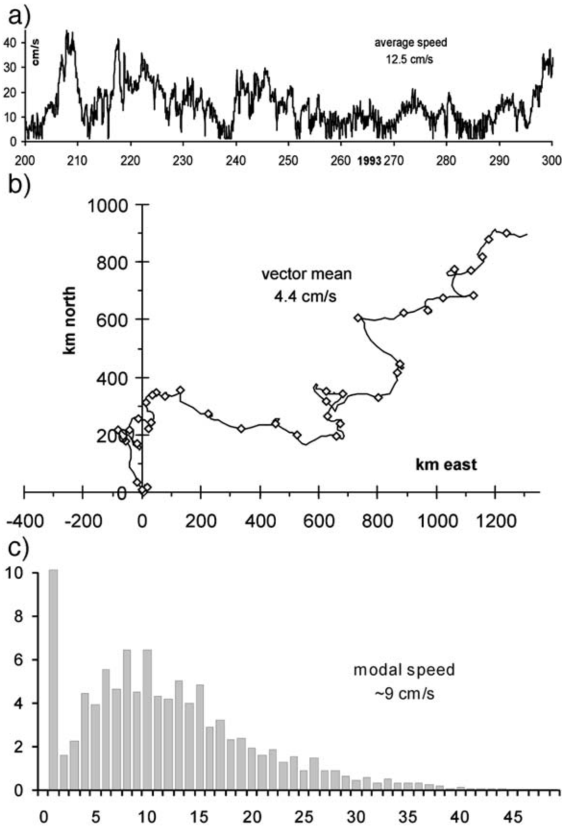

Fig. 3. Bottom-current data from BAS Mooring 11, 50 m above the seabed (part of Fig. 2 of Pudsey and Howe, 2002) showing (a) current speed over a 100-day period during 1993, (b) a cumulative vector representation of current velocity over the life of the mooring (ca. 700 days) with crosses at 10-day intervals and (c) the relative abundances of current speeds (the peak at zero speed represents a stalled current meter, when speeds were below 1.5 cm s^{-1}). Speed at this site rose above 30 cm s^{-1} for about 10 days within a year's record, but otherwise was mainly below 15 cm s^{-1}. Mean speed and velocity here may be compared with mean speed and velocity of 22 and 17 cm s^{-1} near the seabed in a nondepositional area within the deep-water gap in the North Scotia Ridge at 48°W, close to the ACC "axis" (Zenk, 1981).

(overlying, slightly warmer and less dense, modified in part by mixing) Weddell Sea Deep Water (WSDW), but in recognition that bottom-water production in the past may not have involved both processes of modification, we use the generic term Southern-Origin Bottom Water (SOBW). WSDW flows beneath the ACC within the eastern Scotia Sea and South Sandwich Trench, and northward through the western South Atlantic (e.g., Mantyla and Reid, 1983), mixing with overlying water masses as it goes, and SOBW has probably behaved similarly in the past. Northward WSDW transport is small (less than 5 Sv), and flow is steady and, except at western boundaries, slow. Similar northward bottom currents flow into the Indian Ocean and SW Pacific, also beneath the ACC, and a lesser water mass modification, similar to WSBW formation, occurs at other regions of the Antarctic continental shelf. Barker and Burrell (1982), Gamboa et al. (1983) and Barker (1992) suggested that ACC development created an oceanographic "barrier" that reduced or greatly modified northward flow of SOBW. Present-day WSBW usually dissolves even the siliceous biofacies, leading to completely barren deep-sea sediments beneath it, but the cool water masses produced by mixing farther north sustain the benthic foraminifera that are used in oxygen isotopic studies to examine palaeotemperature and ice volume (see Section 4, below).

Numerical models of ocean circulation are too many to cite individually. A useful summary of modern, coupled ocean-atmosphere models is provided in IPCC (2001). Nowlin and Klinck (1986) noted that the majority published at that time calculated (from wind stress) ACC transport that was too high and concluded that a most likely explanation for the discrepancy was form drag, a combination of the flow diversions at such large obstructions as southern South America, the Kerguelen and Campbell plateaux and Macquarie Ridge, and those at ocean floor elevations and ocean floor roughness at smaller scales, including smaller than model grid sizes. Despite subsequent reductions in model grid size, much ocean floor roughness is smaller still, and the conclusion remains valid. A subset of models has addressed the question of Drake Passage opening to examine its effects on circulation and climate. For this class of model, from early (Gill and Bryan, 1971) to modern studies (e.g., Toggweiler and Bjornsson, 1999; Nong et al., 2000), the main use of greatly increased computing power has been to incorporate ocean-atmosphere coupling and increase the freedom within the model in other ways: some of the earlier parameterisations and boundary conditions were seen as unduly limiting and possibly misleading. The modelling has provided many useful insights: in particular, it has shown that, most probably, a significant effect of ACC development has been to reduce meridional transport of water (particularly southward surface water) and heat across it, thus cooling water at higher southern latitudes (by up to 3 °C—Toggweiler and Bjornsson, 1999; by about 1.5 °C—Nong et al., 2000). This supports a role for ACC development in the onset or stabilisation of Antarctic glaciation.

Nevertheless, many modelling problems remain, despite improved computing capabilities, and palaeoclimate modelling in the Southern Ocean region has not yet reached the stage of being limited by the level of observation. The modelling community considers there is greatest benefit in what are essentially sensitivity tests, and there has been less emphasis on reproducing more closely the geometric realities of Southern Ocean circulation. Model grid sizes have reduced, but remain too large to simulate oceanic western boundary currents, fronts, mesoscale eddies and rings, or include moderately realistic seabed topography over substantial areas of ocean. Sea ice and related bottom-water formation are not included. Key questions remain unaddressed, such as the nature of Southern Ocean circulation with Drake Passage closed but the Australia–Antarctica gap open.

4. Antarctic glaciation

Antarctica has lain over the South Geographic Pole for perhaps the past 120 Ma (e.g., DiVenere et al., 1994) but became glaciated only more recently: a cool but not glacial early Cenozoic Antarctic climate is well known (see, for example, Dingle et al., 1998; Stilwell and Feldman, 2000; Dutton et al., 2002). Evidence from an ever-increasing range of sources (ice-rafted debris, Antarctic continental shelf drilling, marine benthic oxygen isotopes, clay mineralogy, deep-sea biotic changes and hiatuses indicating SOBW onset) points to the initiation of Antarctic glaciation at sea level close to the Eocene–Oligocene boundary (ca. 34 Ma). Benson (1975) argued that the modern "psychrosphere"—the cold deep ocean below the thermocline—also originated at the Eocene–Oligocene boundary, suggesting an early onset of SOBW formation. A relatively rapid cooling of the deep oceans in the earliest Oligocene, however, is not supported by recent interpretations of the major oxygen isotope increase at the time (Lear et al., 2000).

Although early work (e.g., Kennett, 1977) was taken to suggest that the initial (Oligocene) glaciation was minor, more recent work has intimated that the Oligocene ice sheet was (though probably warmer) virtually as large as the present ice sheet. Measurements of the oxygen isotopic composition of benthic foraminifera in global deep waters, on which these speculations are mainly based, have become more representative with time: a recent compilation by Zachos et al. (2001) forms Fig. 4. This shows a dramatic increase in isotopic ratio (cooling and/or ice volume increase) across the Eocene–Oligocene boundary (ca. 34 Ma), a decrease (warming and/or ice mass decrease) at about 26 Ma, a more gradual increase over the period 16–13 Ma and a large Plio-Pleistocene increase which is mainly the result of development of Northern Hemisphere ice sheets. The ambiguity between cooling deep waters and global ice volume that has limited the value of this proxy may have been resolved by recent determinations of Mg/Ca ratio (Lear et al., 2000), which appears to be temperature-sensitive alone. These support the notion of a large Oligocene ice sheet. The cause of the short Mi-1 "event" at about 23.7 Ma is uncertain, but Paul et al. (2000) have implicated orbital insolation variations (see also Roberts et al., 2003).

Although the earliest glaciations are conventionally referred to East Antarctica, the recent Oligocene dating of glacial debris onshore in the South Shetland Islands, off the Antarctic Peninsula (Dingle and Lavelle, 1998; Troedson and Smellie, 2002), suggests caution is needed in inferring strong zonal differences in post-Eocene circum-Antarctic climate, even before an ACC was fully established. Barker and Camerlenghi (2002) have speculated that the transport of pack ice from the southern Weddell Sea, and of icebergs originating in East Antarctic ice, to the eastern Antarctic Peninsula margin by clockwise circulation of an ancestral Weddell Gyre, would have helped cool the Antarctic Peninsula.

Information about the level of Antarctic glaciation at other times is sparse. Drilling at Cape Roberts (Ross Sea) provided evidence of East Antarctic glacial and interglacial conditions during the Oligocene and early

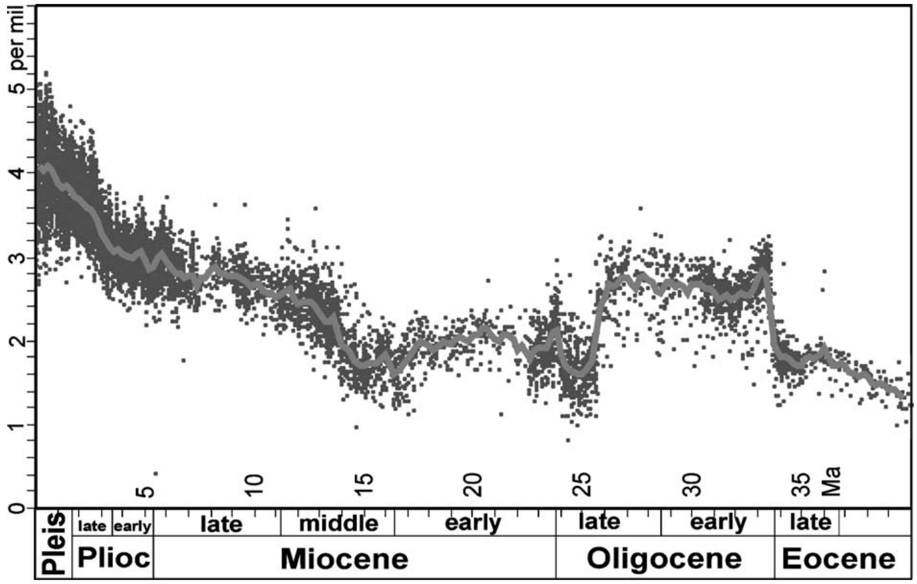

Fig. 4. Raw benthic oxygen isotopic data (Zachos et al., 2001) for the past 40 m.y., and a smoothed version of the same data—revised from Fig. F9 of Barker and Camerlenghi (2002). Timescale from Berggren et al. (1995).

Miocene (Powell et al., 2001; Raine and Askin, 2001), indicating steady cooling. Recent drilling off the Antarctic Peninsula (Barker and Camerlenghi, 2002) supports suggestions (e.g., Barker et al., 1999) that the middle–late Miocene Antarctic ice sheet covered virtually as much of the continent as the present one, and suggests that its volume (though perhaps not its temperature) may have been independent of climate over the past 9 m.y. or so. The temperature dependence of the mean isotopic composition of the ice sheet has not been investigated, and the relative contributions of the ice volume and bottom-water temperature changes to benthic oxygen isotopic variation remain uncertain (though additional measurements of Mg/Ca will be useful in future, e.g., Billups and Schrag, 2002).

The onsets of glaciation and of the ACC have long been linked (e.g., Kennett, 1977). However, De Conto and Pollard (2003a,b) model the onset of glaciation as a response to steadily declining atmospheric pCO_2, as has been measured[1] for the Cenozoic (although no estimates have been published of levels between 25 and 40 Ma; see Pagani et al., 1999; Pearson and Palmer, 2000). Assuming a 20% reduction in meridional heat transport to be the effect of ACC onset (Toggweiler and Bjornsson, 1999; Nong et al., 2000), De Conto and Pollard show that this affects the threshold level of CO_2 required for glacial onset, but is not a requirement for it (Fig. 5). This work also belies the suggestion of Prentice and Matthews (1991) that the principal effect of circum-Antarctic ocean cooling would have been reduced evaporation close to the continent, hence reduced continental precipitation and ice sheet volume.

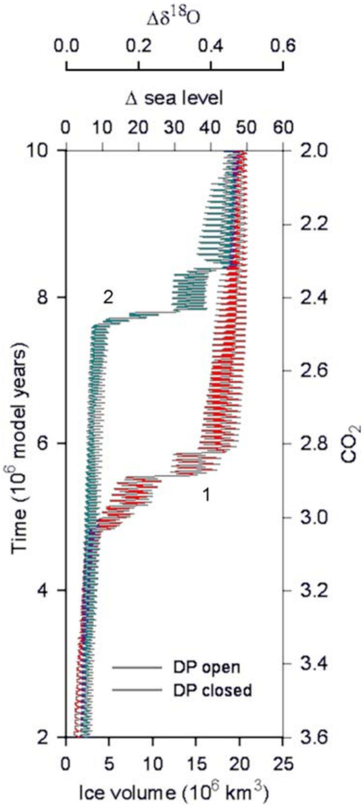

Fig. 5. Size of Antarctic ice sheet as a function of atmospheric pCO_2, with (1, red in electronic version) (red) and without (2, blue in electronic version) (blue) an ACC ("Drake Passage open/closed"), in the model of DeConto and Pollard (2003a: part of their Fig. 2).

5. Geology—speculation on ACC onset

Geological data cited as evidence of ACC onset and development come from DSDP, IPOD and ODP drilling legs (mainly from the deep sea in the Southern Ocean) and from marine (mainly continental shelf) sediments now exposed on land. Insights may derive also from more distant data and from an examination of the modern glacial–interglacial variation: though the glacial–interglacial record does not extend back very far in time, it is more easily accessible (by piston coring) than the older record. In addition, there is relevant evidence from a quite different source—molecular biological studies implying times of development of cold-water characteristics in Antarctic fish.

In all cases, the principal indications are of one or more of

(a) cooling, mainly of surface waters, evidenced through onset of a siliceous biofacies (with in some instances a coeval calcareous biofacies identified to the north) or reduced species diversity to the south;

[1] Reasons for decreasing pCO_2 levels during the Cenozoic do not concern us here, but are most commonly cited as increased weathering of silicate minerals as the result of Himalayan uplift (e.g., Raymo, 1997; Francois and Godderis, 1998). Also mentioned are decreasing activity of large igneous provinces and thus volcanic inputs, and decreased metamorphic degassing (Kerrick and Caldeira, 1998).

(b) zonal species distribution (e.g., spreading of an endemic Pacific species to the Atlantic);
(c) a hiatus indicating strong bottom currents.

All of these must be examined with care, since they may also have non-ACC causes in certain circumstances. In particular, there may be different origins for indications of temperature change and bottom current flow.

5.1. Offshore drilling

Sites in the regions affected by the ACC (Fig. 1a) have been drilled during 5 DSDP legs (28, 29, 35, 36, 71) and 10 ODP legs (113, 114, 119, 120, 177, 178, 181, 183, 188, 189), but this abundance of activity is misleading: only a small minority of the records from those legs have any relevance to ACC onset and subsequent development. Some sites were not usefully located, sediment recovery was often poor and the earlier legs obtained cores only at intervals down-hole. At many sites, no sediment of relevant ages was recovered, or even originally deposited. ODP Leg 189 was drilled (at least in part) to examine the effects of separation of the South Tasman Rise (Australia) and Antarctica, but the other legs were not aimed specifically at the onset and development of the ACC: the relevance of their results was largely fortuitous and their interpretation in many cases speculative.

The major focus of both earlier and more recent drilling activity has been the SE Indian–SW Pacific region, and the relevant sites from the earliest drilling, DSDP Legs 28 and 29, lie in deep water south and southeast of Australia. Kemp et al. (1975), summarising the palaeoceanographic results of drilling on Leg 28, described a northward migration of a nominal Polar Front—a siliceous biofacies and IRD to the south, a calcareous biofacies to the north—which appeared to follow the migrating mid-ocean ridge axis from the late Oligocene until the Pliocene, then accelerate rapidly northward (Sites 265-8). Barker and Burrell (1982) noted a hiatus or near-hiatus preceding the initial or a greatly increased biosiliceous deposition at these and other sites (278, 325) and suggested that an "ACC axis" had accompanied the "Polar Front." Nelson and Cooke (2001) have interpreted data from several drilling legs, including Legs 28 and 29 in the south, to infer the development and successive northward migration of several fronts in the SW Pacific through the Cenozoic.

Exon et al. (2001) have described the initial results of ODP Leg 189 on the South Tasman Rise, which aimed (in part) to examine the Cenozoic development of the Tasmanian Seaway, building on some of the results of DSDP Leg 29 in the same region. A latest Eocene or early Oligocene shallow-water connection between the Pacific and Indian Oceans was inferred from the nature of the record at four of the five sites (1168–1172) drilled during Leg 189. The Eocene–Oligocene boundary did see an abrupt change from siliciclastic to pelagic deposition, indicating subsidence, but the Oligocene record is incomplete except at the most northerly site drilled (north of the Subtropical Convergence), and the effects of an onset of a deep "circumpolar circulation" are entangled with those of rapid tectonic subsidence of the South Tasman Rise and the assumed development of Antarctic glaciation. As with DSDP Leg 29 (e.g., Kennett, 1977; Kennett and Houtz, 1975), the results of drilling were interpreted in the Inital Reports as demonstrating a causal link between development of a Tasmanian Seaway (assumed to allow onset of some kind of ACC—at first shallow and then deep, but not fully circumpolar until Drake Passage opened in the early Miocene) and the development of Antarctic glaciation.

ODP Leg 181 examined the region east of New Zealand (Carter et al., 1996; 1999), where the ACC at present flows over and south of the southern Campbell Plateau and overrides a cold Deep Western Boundary Current (DWBC). The DWBC is essentially Antarctic-derived, with components from the Weddell and Ross Seas, the former via ACC-transported WSDW. Again, drill sites were mainly north of the ACC, and those farther south did not provide Eocene or Oligocene sediments. The more southerly sites from DSDP Leg 29 (Kennett and Houtz, 1975) lie closer to the present PF than those of Leg 181.

In the Indian Ocean sector, the Kerguelen Plateau has seen all or part of three ODP drilling legs (119, 120 and 183). These legs were drilled mainly to sample older sedimentary rocks and basement on the plateau and on the Antarctic shelf in Prydz Bay (also a target of Leg 188). Younger sediments at many of the sites were poorly recovered, largely precluding high-resolution studies. Nevertheless, it is interesting that nannofossils (albeit in restricted assemblages) dominated in preserved sediments at 1000–2000 m depth on the Central Kerguelen Plateau, well south of the present ACC axis, until the late Miocene (Aubry, 1992; Wei and Wise, 1992: sites 737, 744, 747-8, 751), suggesting that the Polar Front was established north of the drill sites only in the late Miocene or Pliocene.

The final barrier in an otherwise complete circumpolar deep-water path almost certainly lay between Antarctica and South America, but comparatively little work has been done on the marine geological record from the South Atlantic sector. Kennett and Barker (1990), summarising the results of ODP Leg 113 in the Weddell Sea, noted the profound differences between Maud Rise (Sites 689 and 690) and the South Orkney microcontinent (Sites 695 and 696), interpreting them in terms of post-Eocene sea-ice formation and water-mass modification in the southern Weddell Sea (as happens today). They argued that biogeographic similarities between Maud Rise and the Falkland Plateau (DSDP Legs 36 and 71, particularly Sites 511-2) during the Oligocene ruled out the existence of an ACC and PF at that time: they identified hiatuses across the Oligocene–Miocene boundary on Maud Rise, coeval with others on the Falkland Plateau and Kerguelen Plateau, as possibly marking ACC development. In contrast, Diester-Haass and Zahn (1996) suggested that Drake Passage had opened (sufficiently to create an intermediate-depth ACC and PF) by 37 Ma, on the basis of a change at that time in the relation between temperature and preserved benthic biomass at ODP Site 689, and proposed that a proto-PF moved south of Maud Rise at times during the Oligocene. Subsequently (e.g., Diester-Haass and Zahn, 2001), these authors have argued for increased productivity at several Southern Ocean locations stemming from ACC initiation around the Eocene–Oligocene transition.

Other sites drilled during ODP Legs 113 and 114 had mainly other concerns (in particular, tectonics, Antarctic glaciation and the development of SOBW). ODP Leg 177 drilled at a number of South Atlantic sites, but most holes did not aim to penetrate below the Pliocene or were located to the north of the relevant water masses and fronts. However, Site 1090 (3700 m water depth) did sample down to the Eocene. Latimer and Filippelli (2002) suggested an age of 32.8 Ma for the opening of Drake Passage and creation of the ACC, based on metallic element ratios of Site 1090 sediments which they considered reflected changes in source regions and biogenic production. This site remains perhaps the closest drill site to the South American–Antarctic sector to be studied in detail with this objective in view, but it lies north of the present-day SAF, and at 8°53′E is some distance from Drake Passage, so the data are capable of other interpretations. A depositional hiatus at the site extends from the early Miocene (ca. 16 Ma) to the Pliocene, making it impossible to examine more recent history. Billups et al. (2002) showed that seabed temperatures at Site 1090 were lower than at sites farther north, during the late Oligocene and early Miocene, but ^{13}C gradients were low, precluding estimates of the directions of deep-water flow. Nevertheless, they believed that there was an ACC at the time, on the basis of the low temperatures at more southerly sites.

Some have inferred ACC development from observations remote from the Southern Ocean. In particular, Gamboa et al. (1983) suggested that an early Miocene increase in sedimentation in the Brazil Basin resulted in part from a reduction in SOBW flow there, caused by its disruption by the ACC. In addition, Pagani et al. (2000) attributed early Miocene changes in carbon isotopic composition of C37 alkenones over the Rio Grande Rise (ODP Site 516) to the presence there of nutrient-rich Antarctic Intermediate Water, formed in the vicinity of the ACC as a result of the sinking and mixing of Antarctic and sub-Antarctic surface water (Gordon, 1971). They associated this with opening and deepening in the Drake Passage region, causing onset of the modern ACC.

Recently, various isotopic measurements have been used to infer patterns of deep-sea circulation and the efficiency of deep-water exchange between ocean basins (review by Frank, 2002). Data extending back into the Paleocene are rare and usually have low time resolution and (in the case of manganese nodules) lack reliable age models. However, the available Pb and Nd data do not show a clear change in circulation efficiency between Atlantic and Pacific oceans over the period (late Eocene to middle Miocene: see Fig. 11a and b of Frank, 2002) within which the ACC may have developed.

5.2. Onshore evidence

Land exposures are of sediments deposited originally on the continental shelf and therefore reflecting shelf conditions, which may not represent those in the deep ocean. Nevertheless, they may prove useful. For example, Jenkins (1974) described the onset of a proto-ACC on the basis of an eastward spread, into the SW Pacific from a region south of Australia, of a planktonic foraminifer *Guembelitria stavensis* at a time now dated (e.g., Berggren et al., 1995) as about 28.5 Ma, using a mixture of onshore and offshore data. Fordyce (1977, 2003), and Fordyce and Barnes (1994) attributed the first occurrences of Mysticetae (baleen whales), in mid-Oligocene shallow-water New Zealand sediments and lowermost Oligocene sediments of Seymour Island, Antarctic Peninsula, respectively, to increased planktonic production resulting from the onset of the ACC and development of the Antarctic ice sheet, via increased nutrient availability and more open water circulation. Foster (1974) proposed a late Eocene age for Drake Passage opening, on the basis of an echinoid assemblage change (taken to indicate cooling) in Australian sediments.

5.3. Glacial–interglacial variation

ACC behaviour through a glacial cycle serves to isolate the effects of certain causes of variation; in particular, the timescale (ca. 10^5 years) is much shorter than that of tectonic variation. In addition, around glacial maxima, the Antarctic ice sheet is grounded to the continental shelf edge, and the fringing sea-ice zone is more extensive. As a result, atmospheric temperatures are reduced during glacials (so the belt of driving winds may move north), and the ocean-atmosphere momentum coupling is reduced because of sea-ice cover. We may expect from this that ACC strength will be reduced during glacials, since its pathway is already topographically constrained to lie south of the zone of westerly winds that drive it. On the other hand, wind strengths may be greater if meridional atmospheric temperature gradients are increased (i.e., if equatorial temperatures are relatively constant and high-latitude temperatures fall). In addition, sea level is lower during glacials, largely as a result of the growth of large Northern Hemisphere ice sheets, which reduces the shallow pathways available to Antarctic fronts and the ACC, making the deeper pathways more important. In addition, the nature and strength of adjacent ocean circulation systems may change: many authors have detected a reduction in North Atlantic Deep Water production during glacials, and if an ice sheet is grounded to the Antarctic shelf edge, we may anticipate reduced production of SOBW (Kellogg, 1987; Pudsey, 1992. Pudsey et al., 1988). Finally, there may be changes in the nature of the available sediment, particularly the terrigenous component.

Observations of ACC variation through a glacial cycle are difficult to interpret, since sample sites are scarce and, in most regions, observations are too few to separate a latitudinal migration from a change in strength. Where the ACC is topographically constrained, as within the Scotia Sea, fewer cores may be required for valid results. Pudsey and Howe (1998; 2002) have proposed a glacial increase in ACC flow from grain-size studies, having assumed no significant glacial–interglacial change in terrigenous sediment source. There is great value in such studies, particularly since the present southward displacement of the ACC from its belt of driving winds, together with suggestions that climatic zones have migrated northward (e.g., Kemp et al., 1975; Wei and Wise, 1992), opens the possibility that in the past the ACC may have been more vigorous than today.

5.4. Molecular biology

A different line of evidence bears only weakly on the question of ACC origin at present, but should be included. Molecular studies of diversification of (in particular) Antarctic and sub-Antarctic Notothenioid fish species assume a deepening of Drake Passage and creation of the PF since 25–22 Ma (see Bargelloni et al., 2000), and evolution of their "antifreeze" genes (Chen et al., 1997) provides an estimate of 12–5 Ma for their evolutionary response to cold surface waters. However, calibration of the molecular clock is uncertain, and the possibility is noted of diversification long after the changes in the physical environment that would have enabled it.

5.5. Derived studies

It is reasonable that published estimates of Drake Passage opening, or the onset of the ACC, should have been incorporated into other studies of broader scope. We note some of those here, since their great value in many but not necessarily all respects may have led to their wide acceptance. For example, the global study of Berggren and Hollister (1977) quoted Foster's (1974) late Eocene age for Drake Passage opening, but considered a mid-Oligocene separation of Antarctica and the South Tasman Rise (Kennett et al., 1974) as removing the final barrier to ACC development about 30 m.y. ago. They distinguished this from the late Eocene or early Oligocene onset of Antarctic glaciation and related formation of SOBW and associated the later event with Oligocene deep-sea sedimentary hiatuses widespread in and around the Indian and SW Pacific oceans.

In an oft-quoted paper, Kennett (1977) undertook a useful review of Southern Ocean palaeocirculation and palaeoclimate. He followed Weissel and Hayes (1972) in concluding that a deep-water gap south of the South Tasman Rise developed at about the Eocene–Oligocene boundary, coincident with the initiation of sea ice, hence bottom-water production, around Antarctica. Quoting the evidence available at the time, he was uncertain about the time of opening of Drake Passage, but concluded that it (and the onset of full circumpolar circulation) occurred at or shortly after 30 Ma. He proposed an early Miocene age for development of the Polar Front (Antarctic Convergence), defined as the northern margin of a belt of biosiliceous ooze deposition and IRD. Interestingly, he did not conclude that the full, deep-water ACC and Antarctic glaciation both started at the Eocene–Oligocene boundary, as many have inferred he did. He thought they might be related but could not understand why there should be a long gap between the 30 Ma of ACC onset and the 14 Ma of substantial Antarctic glaciation (he interpreted the Eocene–Oligocene boundary increase in oxygen isotopic ratio as a temperature effect). The concept of a "proto-ACC," that could have existed without a complete deep-water path, was introduced at this time, but was never defined.

More recently, Zachos et al. (2001) took Drake Passage opening to have been at about 30 Ma, citing Lawver and Gahagan (1998), but saw no dramatic change in isotopic values at that time. They argued that atmospheric moisture availability, assumed closely correlated with pCO_2, may have been the most important control on ice sheet volume, affecting isotopic values, particularly before 25 Ma when pCO_2 was high, but remarked also the likely influence of Antarctic thermal isolation by widening of oceanic passageways, and speculated that tectonic events such as gateway opening may have been the dominant influence over the past 25 Ma when pCO_2 was comparatively low.

6. Discussion

6.1. Reassessment

The geologic record in the Southern Ocean is poorly known. Very few drill sites in a useful latitude range provide a continuous sedimentary record of even a part of the period needing to be examined in order to determine the origin and development of the ACC, and almost all sites were occupied originally for other purposes. Further, the Southern Ocean record is not as continuous as elsewhere: most DSDP and ODP oceanic drill sites (particularly those preserving a biogenic section) were located on anomalously elevated basement (for example, the Falkland and Kerguelen plateaux, Maud Rise, starved Antarctic margins), to avoid loss of a calcareous record because of a shallow CCD. Hiatuses are more common at such sites than on "normal" ocean floor, because ocean currents accelerate to pass anomalous elevations and are fast at many continental margins. In addition, some identified hiatuses are suggested to result from bio- or magneto-stratigraphic imperfections (see Ramsay and Baldauf, 1999): in considering a sedimentary hiatus, the depositional environment and sediment type must be taken into account.

The classical concept of sub-ACC sedimentation is of deposition of a calcareous biofacies to the north and siliceous biofacies to the south (e.g., Goodell, 1973; Kemp et al., 1975) of a broad axial zone within which, in many cases (Barker and Burrell, 1982), strong bottom currents prevent deposition of any kind. Most of the published interpretations, based on the data described above, place ACC onset as coeval with or slightly later than the onset of substantial Antarctic glaciation, at or close to the Eocene–Oligocene boundary (the "smoking gun" of synchroneity). Whether or not this is correct, or the changes were causally related, substantial glaciation almost certainly involved cooling of both deep (SOBW) and shallow waters (particularly, by sea-ice formation). Preferential preservation of a siliceous biofacies was made more likely by the existence of cold surface and deep waters, with or without an ACC. Surface palaeotemperature zones migrated north with time. The bulk of such conclusions are based on data from drill sites in the SE Indian–SW Pacific sector, close to the South Tasman Rise region where a deep pathway almost certainly was created close to the Eocene–Oligocene boundary. Data from elsewhere, particularly the South Atlantic sector where the final barrier to complete circumpolar deep-water circulation most probably lay, are more sparse and more disparately interpreted.

Here, we reassess the evidence in light of the most recent consensus in each of the fields associated with ACC development (modern physical oceanography, tectonic evolution, Antarctic glaciation). In particular, virtually all ACC data and conclusions must take account of three major insights:

1. that modern ACC transport lies mostly, if not entirely, within deep-reaching, continuous oceanic fronts or frontal zones
2. that, although it may have contributed, ACC onset need not have caused the onset of Antarctic glaciation, given something like the observed decline in the atmospheric concentration of greenhouse gases through the Cenozoic
3. that, although there is a tempting synchroneity between the creation of a deep-water gap south of Australia at the Eocene–Oligocene boundary and the onset of substantial Antarctic glaciation, a similar deep gap south of South America, necessary for an ACC akin to that of today, may not have come into existence until much later.

Of these insights, the most significant is the modern physical oceanographic one that all or most present-day ACC transport is within fronts. Virtually all palaeo-ACC interpretations are based on observations of sea-surface temperature, in particular the transition from biocalcareous to biosiliceous facies, as occurs at present beneath the SAF and PF. However, the distribution of drill sites is too sparse to allow the firm conclusion that these sea-surface observations reflect the presence of large areas of near-constant temperature, separated by fronts. Fronts do appear to be a relatively common feature of the modern ocean, but the conditions that develop and sustain them are not understood.

In addition, if the fronts did exist, it cannot easily be concluded that they were deep-reaching, or the loci of strong, along-front currents, or (particularly) that they were continuous. Relatively few modern frontal zones show these features. Of the two or three known modern, continuous fronts (all of them in the Southern Ocean), we may cite the behaviour of the Polar Front near the Kerguelen Plateau: the shallow expression (as defined by sea-surface temperature changes, Moore et al., 1999) separates from the deep component, to cross the plateau well south of Kerguelen Island, while the (much greater) deep flow (whether determined from measurements of sea-surface elevation—Gille, 1994, or ship-based data—Orsi et al., 1995) is constrained to pass through the deep water north of the plateau. This demonstrates the importance to the present-day ACC transport of an uninterrupted deep pathway.

Again, it may be argued that the observed sea-surface cooling, with or without intervening fronts, is an effect of continental glaciation (and associated sea-ice formation) rather than its cause. Independently of the ACC, much of the Southern Ocean has sea-ice cover for part of the year at present, and at parts of the Antarctic continental margin, sea-ice formation is sufficiently persistent and systematic for it to be involved also in the production of cold, salty bottom water. In the past, even without an ACC, so that water reaching a glaciated Antarctic margin would originally have been much warmer than today, similar processes may well have acted.

We must also reconsider enhanced biogenic production, proposed as a proxy for invigorated circulation as a result of ACC development. Upwelling, carrying nutrients into the photic zone, is a feature of the present-day Southern Ocean south of the Polar Front. However, since cold water sinks also at parts of the glaciated Antarctic margin, upwelling, inferred from signs of enhanced biogenic production in the geologic record, cannot be taken as a response to sinking at the front itself and, therefore, an indicator of the ACC.

The classical concept of a broad axial zone of ACC-related bottom currents also needs to be reexamined. As noted above, the bulk of modern ACC transport is within narrow jets, extending usually to the seabed, rather than a broad zone, but the jets meander and in places generate detached "rings." Thus, the seabed within a zone much broader than the jet width is swept at intervals by strong bottom currents, but experiences much lower bottom current speeds otherwise. Fig. 3, showing the variation in bottom currents at a site within the Scotia Sea, has already been described. Core-top sediment at this site is a mud-bearing diatom ooze (Pudsey and Howe, 1998), presumably deposited under quiet conditions, between transits of front-related eddies or rings. The geological record of high eddy kinetic energy (K_e) regimes such as the ACC has been little studied, and McCave et al. (1995) pointed out that grain-size studies cannot distinguish between different bottom current regimes, nor provide useful estimates of mean current flow (mean kinetic energy K_m) in high K_e regimes. In addition, lower bottom current speeds are needed to prevent deposition than to resuspend the same sediment, and inferences from work on silt-sized and larger particles are more reliable than those from clays. Nevertheless, the geological record will reflect in some way the existence of faster currents over an area broader than the jet width, since the periodicity of variation remains shorter than the time resolution of most sediment sections. In summary, grain-size studies on properly sited cores are capable of indicating an onset and subsequent development of the ACC. Care must be taken, of course, to choose sites that have lain beneath the ACC, rather than beneath a bottom current, such as SOBW, that is clearly related to continental climate. In addition, estimates of flow direction (as from magnetic susceptibility anisotropy, for example) may be worth examination, but could be degraded by the uncertain direction of fast flow within eddies and rings.

It should not be forgotten, of course, that ACC onset is only one of its features (although the estimation of, for example, changes in current strength through the life of the ACC, is much more difficult), and that determinations of sea-surface temperature provide useful palaeoclimate data whether or not they are associated with an ACC. For example, the PF probably migrated north of the Kerguelen Plateau, and of the mid-ocean ridge south of Australia as recently as the late Miocene or early Pliocene (Kemp et al., 1975; Wei and Wise, 1992). In addition, before the advent of the ACC, which can be expected to have made climates in different sectors of the Southern Ocean and Antarctic continent more similar to each other, differences may have been quite dramatic. For example, the benthic oxygen isotopic curve (Zachos et al., 2001 and Fig. 4) suggests a warming or ice volume reduction at the end of the Oligocene, which is in conflict with some Antarctic data (e.g., from Cape Roberts, Ross Sea sector: Powell et al., 2001), but might more easily be reconciled if there was no ACC.

6.2. Future work

The situation is not one of trying to resolve between a number of conflicting but plausible estimates of the time of onset of the ACC. Rather, we must reconsider completely how to determine ACC onset, development and significance—what questions to ask of the present data set, of numerical modelling and in future drilling campaigns, and what parameters will be useful. Clearly, conclusions based on sea-surface palaeotemperatures alone cannot be maintained: the observations and their (sea-surface temperature) interpretations survive, but their greater significance for palaeoclimate is uncertain and must await clarification. It is important to try to establish the time of onset and development of the ACC as we know it today—as a completely circumpolar current—and to see if this development coincided systematically with any of the other apparent changes in sea-surface temperature or the level of glaciation. This would contribute towards an understanding of both the relative contributions of ocean circulation and greenhouse gases to climate change (as exemplified by Antarctic glaciation), and the nature of any interim (Oligocene?) Southern Ocean circulation. In essence, particularly if a complete deep-water circumpolar path developed late, we do need to know the nature of Southern Ocean circulation (including perhaps a "proto-ACC") in the period between the opening of a deep gap south of Australia (and near-synchronous development of a significant Antarctic glaciation) and the creation of the complete path, and the part such circulation played in palaeoclimate.

Numerical modelling of Southern Ocean circulation appears to have some way to go. We do not know the nature of circulation within a realistic Southern Ocean with an Indian–Pacific connection but without a complete deep-water path. We appreciate that, at present, most or all models do not include fronts, eddies and rings, sea-ice or bottom-water formation, but we do not know which conclusions may safely be drawn from such models as do exist. In particular, we do not know how meridional oceanic heat transport has been affected by changes in Southern Ocean temperature and circulation.

Other related fields of study could usefully be extended. We may hope, for example, that additional estimates of atmospheric pCO_2 are made, to fill the 25–40-Ma gap and to test the existing data (Pagani et al., 1999; Pearson and Palmer, 2000). More precise definitions may arise of the necessary and sufficient conditions for the formation and maintenance of oceanic fronts. The 26-Ma warming apparent in the oxygen isotopic compilation (Zachos et al., 2001 and Fig. 4) remains hard to explain in terms of pCO_2 variation, and the numerical model of Antarctic glaciation (De Conto and Pollard, 2003a,b) may benefit from incorporating wet-based ice sheets, a more realistic (circulating and deep) ocean and improved estimates of oceanic heat transport.

We may list the sediment properties to be examined in cores, that will provide strong or unambiguous estimates of ACC existence. Grain-size studies on the coarser grained component of non-biogenic sediments (the effectively pelagic ice-rafted detritus—IRD, for example) are capable of showing gross changes in bottom current strength, and at suitably located deep-water sites should show the time of onset of the ACC. Mineralogy or geochemistry of both clay minerals and IRD provide information on provenance, and therefore on surface and bottom currents, that could be diagnostic of the existence of deep pathways (the uncertainty does not extend back farther than the onset of Antarctic glaciation, so a supply of IRD is assured). Beyond the reach of SOBW, temperatures of both intermediate and deep waters in the more southerly parts of the Southern Ocean could be more diagnostic of ACC existence than the shallow waters of the photic zone, but suitable sites may be hard to find. In any investigation involving these parameters, of course, it would be necessary also to examine the biogenic proxies of sea-surface temperature, so as to be able to assess their value and thus possibly to reinterpret the vast amount of existing data. Given the levels of energy within the band of orbital insolation variation, it will be necessary to determine all properties at intervals sufficiently close to avoid aliasing, and above all, the importance of bottom currents should not lead to the pursuit of the hiatus: a hiatus caused by sediment erosion extends beyond its cause, and in the Southern Ocean, as noted above, hiatuses are more abundant than usual and less useful.

Where to look in the future, for direct rather than far-field evidence? We suggest the region in and around the Scotia Sea, for at least the initial study. The timing of creation of a continuous deep-water gap there remains uncertain, the sector is relatively poorly sampled (but reasonably well known in other respects), and to examine a proximal region may increase the size of the signal and reduce the chances of contamination of the record by the effects of other, unwanted or unknown processes. Ocean floor old enough to resolve between all possible models is preserved more certainly outside the Scotia Sea, to the east. However, if the question of onset can be resolved, the Scotia Sea itself has the additional advantage that ACC variation can be measured there more economically than elsewhere, as the elevation of the North Scotia Ridge restricts the possibility of variation being expressed as meridional migration. A study in this region has the potential not only to resolve the question of onset of the modern ACC, but also to contribute to a more general understanding of the significance of the ACC and of existing data from elsewhere.

7. Conclusions

The sedimentary record of Southern Ocean palaeoclimate and palaeocirculation, derived almost entirely from DSDP and ODP drilling over the past 30 years, is sparse and sited mainly in the SE Indian–SW Pacific sector, but in the context of the ancillary information of the time, appears to have yielded a consistent story. In the past, the great majority of workers have accepted the conclusions that an Antarctic Circumpolar Current (ACC) developed close to the Eocene–Oligocene boundary, at the same time as the Antarctic continent underwent a significant, rapid glaciation, and that the ACC caused or considerably aided glaciation by reducing the oceanic provision of heat to the continental margin. The replacement of a biocalcareous by a biosiliceous facies in ocean sediments to the south was taken as indicating the existence of the ACC and defining the position of its axis.

Almost all of the conclusions outlined above are now open to question: only the time and acknowledged rapidity of onset of Antarctic glaciation, and coeval creation of a deep-water gap south of Tasmania, remain intact. These new uncertainties arise from developments in ancillary fields, principally in physical oceanography but also in numerical modelling of glaciation and in regional tectonics. All or most of modern ACC transport is now known to occur in continuous, deep-reaching oceanic fronts (the Sub-Antarctic Front, the Polar Front and probably a Southern ACC Front), and a biosiliceous sea-surface assemblage and the biocalcareous/biosiliceous transition cannot be taken as certain indicators of the existence and location of a continuous, deep-reaching, transporting front. In addition, numerical modelling has shown that although the ACC may have contributed to Antarctic glaciation, rapid glacial onset could have been caused by a steady decline in atmospheric pCO_2, similar to measured concentrations, without any change in ocean circulation. Thirdly, it is at least possible that a continuous deep-water circumpolar pathway, necessary for the ACC as we know it today, did not develop until long after the Eocene–Oligocene boundary. Thus, the biogenic sedimentary record of sea-surface temperature may be showing an effect of Antarctic glaciation rather than a cause and may not indicate the existence of the ACC.

In the face of this difficulty, what can be done? Of course, we hope to encourage additional work in the ancillary fields: more and better determinations of pCO_2, a clearer definition of the nature of oceanic fronts, numerical modelling of ocean circulation that explores heat transport by and around fronts, and the nature of circulation within a partially connected Southern Ocean. However, within our own field, we have identified those properties of oceanic sediment cores which will identify the presence of a deep-reaching ACC—for example, grain-size studies of bottom current strength, geochemical and mineralogical studies of clay minerals and IRD to examine palaeocirculation—but which must be accompanied by the more traditional studies of photic zone biogenic assemblages with the aim of recovering a value for the considerable effort already expended. We identify the Scotia Sea region, where the final barrier to a complete deep-water ACC pathway probably lay, as the most useful region for further study.

Acknowledgements

We are grateful for the positive and useful comments of David Rea and Neville Exon. PFB acknowledges the support of Birmingham University, the British Antarctic Survey and Natural Environment Research Council over many years, and ET's research is partially funded by NSF grant EAR-0120727.

References

Aubry, M.-P., 1992. Paleogene calcareous nannofossils from the Kerguelen Plateau, Leg 120. In: Wise, S.W., Schlich, R., et al. (Eds.), Proc. ODP, Sci. Results, vol. 120. Ocean Drilling Program, College Station, TX, pp. 471–491.

Bargelloni, L., Zane, L., Derome, N., Lecointre, G., Patarnello, T., 2000. Molecular zoogeography of Antarctic euphausiids and notothenioids: from species phylogenies to intraspecific patterns of genetic variation. Antarct. Sci. 12, 259–268.

Barker, P.F., 1992. The sedimentary record of Antarctic climate change. Phil. Trans. Roy. Soc. Lond. B. 338, 259–267.

Barker, P.F., 2001. Scotia Sea regional tectonic evolution: implications for mantle flow and palaeocirculation. Earth Sci. Rev. 55, 1–39.

Barker, P.F., Burrell, J., 1977. The opening of Drake Passage. Mar. Geol. 25, 15–34.

Barker, P.F., Burrell, J., 1982. The influence on Southern Ocean circulation, sedimentation and climate of the opening of Drake Passage. In: Craddock, C. (Ed.), Antarctic Geoscience. University Wisconsin Press, Madison, pp. 377–385.

Barker, P.F., Camerlenghi, A., 2002. Glacial history of the Antarctic Peninsula from Pacific margin sediments. In: Barker, P.F., Camerlenghi, A., Acton, G.D., Ramsay, A.T.S. (Eds.), Proc. ODP, Sci. Results, vol. 178. Ocean Drilling Program, College Station, TX, pp. 1–40.

Barker, P.F., Barrett, P.J., Cooper, A.K., Huybrechts, K., 1999.

Antarctic glacial history from numerical models and continental margin sediments. Palaeogeogr. Palaeoclimatol. Palaeoecol. 150, 247–267.

Benson, R.H., 1975. The origin of the psychrosphere as recorded in changes in deep-sea ostracode assemblages. Lethaia 8, 69–83.

Bentley, C.R., Clough, J.W., 1972. Antarctic subglacial structure from seismic refraction measurements. In: Adie, R.J. (Ed.), Antarctic Geology and Geophysics. Universitetsforlaget, Olso, pp. 683–691.

Berggren, W.A., Hollister, C.D., 1977. Plate tectonics and paleocirculation—commotion in the ocean. Tectonophysics 38, 11–48.

Berggren, W.A., Kent, D.V., Swisher III, C.C., Aubry, M.-P., 1995. A revised Cenozoic geochronology and chronostratigraphy. In: Berggren, W.A., Kent, D.V., Aubry, M.-P., Hardenbol, J. (Eds.), Geochronology, Time Scales and Global Stratigraphic Correlation. Soc. Sediment. Geol. Spec. Publ., vol. 54, pp. 129–212.

Billups, K., Schrag, D.P., 2002. Paleotemperatures and ice volume of the last 27 Myr revisited with paired Mg/Ca and $^{18}O/^{16}O$ measurements on benthic foraminifera. Paleoceanography 17, 11 pp. (10.1029/2000PA/000567).

Billups, K., Channell, J.E.T., Zachos, J., 2002. Late Oligocene to early Miocene geochronology and paleoceanography from the subantarctic South Atlantic. Paleoceanography 17, 11 pp. (10.1029/2001PA000568).

Brennecke, W., 1921. Die ozeanographische arbeiten der deutschen antarktischen expedition 1911–1912. Aus dem Archiv der Deutschen Seewarte 39, 1–216.

Cande, S.C., Kent, D.L., 1995. Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic. J. Geophys. Res. 100, 6093–6095.

Carter, R.M., Carter, L., McCave, I.N., 1996. Current controlled sediment deposition from the shelf to the deep ocean: the Cenozoic evolution of circulation through the SW Pacific gateway. Geol. Rundsch. 85, 438–451.

Carter, R.M., McCave, I.N., Richter, C., et al., 1999. Proc. ODP, Init. Rep. 181 [CD-ROM] (Available from: Ocean Drilling Program, College Station, TX 77845-9547, USA).

Chen, L., DeVries, A.L., Cheng, C.-H.C., 1997. Evolution of antifreeze glycoprotein gene from a trypsinogen gene in Antavolume-antarctic notothenioid fish. Proc. Nat. Acad. Sci. U. S. A. 94, 3811–3816.

Cooper, A.K., Barker, P.F., Brancolini, G., 1995. Geology and Seismic Stratigraphy of the Antarctic Margin. Antarct. Res. Ser., vol. 68. Atlas, CD-ROMs, AGU Washington, DC. 301 pp.

Cunningham, S.A., Alderson, S.G., King, B.A., 2003. Transport and variability of the Antarctic Circumpolar Current in Drake Passage. J. Geophys. Res. 108 (C5), 8084 (doi:10.1029/2001JC001147).

Deacon, G.E.R., 1933. A general account of the hydrology of the South Atlantic Ocean. Discov. Rep. 7, 171–238.

Deacon, G.E.R., 1937. The hydrology of the Southern Ocean. Discov. Rep. 15, 1–24.

De Conto, R.M., Pollard, D., 2003a. Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO_2. Nature 421, 245–249.

De Conto, R.M., Pollard, D., 2003b. A coupled climate-ice sheet modeling approach to the Early Cenozoic history of the Antarctic ice sheet. Palaeogeogr. Palaeoclimatol. Palaeoecol. 198, 39–52.

DeWit, M.J., 1977. The evolution of the Scotia Arc as a key to the reconstruction of southwest Gondwanaland. Tectonophysics 37, 53–81.

Diester-Haass, L., Zahn, R., 1996. Eocene–Oligocene transition in the southern ocean: history of water mass circulation and biological productivity. Geology 24, 163–166.

Diester-Haass, L., Zahn, R., 2001. Paleoproductivity increase at the Eocene–Oligocene climatic transition: ODP/DSDP Sites 763 and 592. Palaeogeogr. Palaeoclimatol. Palaeoecol. 172, 153–170.

Dingle, R.V., Lavelle, M., 1998. Antarctic Peninsular cryosphere: early Oligocene (c. 30 Ma) initiation and revised glacial chronology. J. Geol. Soc. 155, 433–437.

Dingle, R.V., Marenssi, S.A., Lavelle, M., 1998. High latitude Eocene climate deterioration: evidence from the northern Antarctic Peninsula. J. South Am. Earth Sci. 11, 571–579.

DiVenere, V.J., Kent, D.V., Dalziel, I.W.D., 1994. Mid-Cretaceous paleomagnetic results from Marie Byrd Land, West Antarctica: a test of post-100 Ma relative motion between East and West Antarctica. J. Geophys. Res. 99, 15115–15139.

Dutton, A.L., Lohmann, K.C., Zinsmeister, W.J., 2002. Stable isotope and minor element proxies for Eocene climate of Seymour Island, Antarctica. Paleoceanography 17, 14 pp. (10.1029/2000PA00593).

Eagles, G., Livermore, R.A., 2002. Opening history of Powell Basin, Antarctic Peninsula. Mar. Geol. 185, 195–205.

Exon, N.F., Kennett, J.P., Malone, M.J., 2001. Proc. ODP, Init. Rep. 189 [CD-ROM] (Available from: Ocean Drilling Program, College Station, TX 77845-9547, USA).

Foldvik, A., Gammelsrod, T., 1988. Notes on Southern Ocean hydrography, sea-ice and bottom-water formation. Palaeogeogr. Palaeoclimatol. Palaeoecol. 67, 3–17.

Fordyce, R.E., 1977. The development of the Circum-Antarctic Current and the evolution of the Mysticeti (Mammalia: Cetacea). Palaeogeogr. Palaeoclimatol. Palaeoecol. 21, 265–271.

Fordyce, R.E., 2003. Cetacean evolution and Eocene–Oligocene oceans revisited. In: Prothero, D.R., Ivany, L.C., Nesbitt, E.A. (Eds.), From Greenhouse to Icehouse: The Marine Eocene–Oligocene Transition. Columbia University, New York, pp. 154–170.

Fordyce, R.E., Barnes, L.G., 1994. The evolutionary history of whales and dolphins. Annu. Rev. Earth Planet. Sci. 22, 419–455.

Foster, F.J., 1974. Eocene echinoids and the Drake Passage. Nature 249, 751.

Foster, T.D., Middleton, J.H., 1980. Bottom water formation in the western Weddell Sea. Deep-Sea Res. 27A, 367–381.

Francois, L.M., Godderis, Y., 1998. Isotopic constraints on the Cenozoic evolution of the carbon cycle. Chem. Geol. 145, 177–212.

Frank, M., 2002. Radiogenic isotopes: tracers of past ocean circulation and erosional input. Rev. Geophys. 40 (1), 1001 (doi:10.1029/2000RG00094).

Gamboa, L.A.P., Buffler, R.L., Barker, P.F., 1983. Seismic stratigraphy of the Brasil Basin and Vema Gap, DSDP Leg 72. DSDP Init. Rep. 72, 481–498.

Ganachaud, A., Wunsch, C., 2000. Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data. Nature 408, 453–457.

Gill, A.E., Bryan, K., 1971. Effects of geometry on the circulation of a three-dimensional southern-hemisphere ocean model. Deep-Sea Res. 18, 685–721.

Gille, S.T., 1994. Mean sea surface height of the Antarctic Circumpolar Current from Geosat data: method and application. J. Geophys. Res. 99, 18255–18273.

Goodell, H.G., 1973. The sediments. In: Goodell, H.G., Houtz, R., Ewing, M., et al. (Eds.), Marine Sediments of the Southern Oceans. Antarct. Map Folio Ser., Folio, vol. 17 Pl., pp. I–IX.

Gordon, A.L., 1971. Oceanography of Antarctic waters. In: Reid, J.L. (Ed.), Antarctic Oceanology. Antarct. Res. Ser., vol. 15. AGU, Washington, DC, pp. 169–203.

Gouretski, V.V., Danilov, I.A., 1994. Characteristics of warm rings in the African sector of the Antarctic Circumpolar Current. Deep-Sea Res., Part I 41, 1131–1157.

Heywood, K.J., King, B.A., 2002. Water masses and baroclinic transports in the South Atlantic and southern oceans. J. Mar. Res. 60, 639–676.

Huber, M., Sloan, L.C., 2001. Heat transport, deep waters and thermal gradients: cpoupled simulation of Eocene Greenhouse Climate. Geophys. Res. Lett. 28, 3481–3484.

Hübscher, C., Jokat, W., Miller, H., 1996. Structure and origin of southern Weddell Sea crust: results and implications. In: Storey, B.C., King, E.C., Livermore, R.A. (Eds.), Weddell Sea Tectonics and Gondwana Break-up. Geol. Soc. (Lond.) Spec. Publ., vol. 108, pp. 201–211.

Iglesias-Rodriguez, M., Armstrong, R., Feely, R., Hood, R., Kleypas, J., Milliman, J.D., Sabine, C., Sarmiento, J., 2002. Progress made in study of ocean's calcium carbonate budget. EOS 83 (365), 374–375.

IPCC, J., 2001. Climate Change 2001: The Scientific Basis. Contribution of Working Group 1 to the Third Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge Univ. Press, Cambridge. 881 pp.

Jenkins, D.G., 1974. Initiation of the proto circum-Antarctic current. Nature 252, 371–373.

Kellogg, T.B., 1987. Glacial–interglacial changes in global deepwater circulation. Paleoceanography 2, 259–271.

Kemp, E.M., Frakes, L.E., Hayes, D.E., 1975. Paleoclimatic significance of diachronous biogenic facies, Leg 28, Deep Sea Drilling Project. DSDP Init. Rep. 28, 909–918.

Kennett, J.P., 1977. Cenozoic evolution of Antarctic glaciation, the Circum-Antarctic ocean, and their impact on global paleoceanography. J. Geophys. Res. 82, 3843–3860.

Kennett, J.P., Barker, P.F., 1990. Latest Cretaceous to Cenozoic climate and oceanographic developments in the Weddell Sea, Antarctica: an ocean-drilling perspective. Proc. ODP, Sci. Results 113, 937–960.

Kennett, J.P., Houtz, R.E., et al., 1975. Initial Rept. Deep Sea Drill. Proj. 29 (1197 pp.).

Kennett, J.P., Houtz, R.E., Andrews, P.B., Edwards, A.R., Gostin, V.A., Hajos, M., Hampton, M.A., Jenkins, D.G., Margolis, S.V., Ovenshine, A.T., Perch-Nielsen, K., 1974. Development of the Circum-Antarctic current. Science 186, 144–147.

Kerrick, D.M., Caldeira, K., 1998. Metamorphic CO_2 degassing from volcanic belts. Chem. Geol. 145, 213–232.

Latimer, J.C., Filippelli, G.M., 2002. Eocene to Miocene terrigenous imports and export production: geochemical evidence from ODP Leg 177, Site 1090. Palaeogeogr. Palaeoclimatol. Palaeoecol. 182, 151–164.

Lawver, L.A., Gahagan, L.M., 1998. Opening of Drake Passage and its impact on Cenozoic ocean circulation. In: Crowley, T.J., Burke, K.C. (Eds.), Tectonic Boundary Conditions for Climate Reconstructions. Oxford Univ. Press, Oxford, pp. 212–223.

Lawver, L.A., Gahagan, L.M., 2003. Evolution of Cenozoic seaways in the circum-Antarctic region. Palaeogeogr. Palaeoclimatol. Palaeoecol. 198, 11–38.

Lawver, L.A., Gahagan, L.M., Coffin, M.F., 1992. The development of paleoseaways around Antarctica. In: Kennett, J.P., Warnke, D.A. (Eds.), The Antarctic Paleoenvironment: A Perspective on Global Change. AGU Ant. Res. Ser., vol. 56., pp. 7–30.

Lear, C.H., Elderfield, H., Wilson, P.A., 2000. Cenozoic deep-sea temperatures and global ice volumes from Mg/Ca in benthic foraminiferal calcite. Science 287, 269–272.

Legeckis, R., 1977. Oceanic polar front in the Drake Passage-Satellite observations during 1976. Deep-Sea Res. 24, 701–704.

Mantyla, A.W., Reid, J.L., 1983. Abyssal characteristics of the world's ocean waters. Deep-Sea Res. 30, 805–833.

McCave, I.N., Manighetti, B., Robinson, S.G., 1995. Sortable silt and fine sediment size/composition slicing: parameters for palaeocurent speed and palaeoceanography. Paleoceanography 10, 593–610.

Moore, J.K., Abbott, M.R., Richman, J.G., 1999. Location and dynamics of the Antarctic Polar Front from satellite sea surface temperature data. J. Geophys. Res. 104, 3059–3073.

Nelson, C.S., Cooke, P.J., 2001. History of oceanic front development in the New Zealand sector of the southern Ocean during the Cenozoic—a synthesis. N. Z. J. Geol. Geophys. 44, 535–553.

Nong, G.T., Najjar, R.G., Seidov, D., Peterson, W.H., 2000. Simulation of ocean temperature change due to the opening of Drake Passage. Geophys. Res. Lett. 27, 2689–2692.

Nowlin Jr., W.D., Klinck, J.M., 1986. The physics of the Antarctic Circumpolar Current. Rev. Geophys. 24, 469–491.

Orsi, A.H., Whitworth III, T., Nowlin Jr., W.D., 1995. On the meridional extent and fronts of the Antarctic Circumpolar Current. Deep-Sea Res. Part I 42, 641–673.

Pagani, M., Arthur, M.A., Freeman, K.H., 1999. Miocene evolution of atmospheric carbon dioxide. Paleoceanography 14, 273–292.

Pagani, M., Arthur, M.A., Freeman, K.H., 2000. Variations in Miocene phytoplankton growth rates in the southwest Atlantic: evidence for changes in ocean circulation. Paleoceanography 15, 486–496.

Paul, H., Zachos, J.C., Flower, B.P., Tripati, A., 2000. Orbitally induced climate and geochemical variability across the Oligocene/Miocene boundary. Paleoceanography 15, 471–485.

Pearson, P.N., Palmer, M.R., 2000. Atmospheric carbon dioxide over the past 60 million years. Nature 406, 695–699.

Powell, R.D., Laird, M.G., Naish, T.R, Fielding, C.R., Krissek, L.A., van der Meer, J.J.M., 2001. Depositional environments for strata cored in CRP-3 (Cape Roberts Project), Victoria Land Basin Antarctica: palaeoglaciological and palaeoclimatological inferences. Terra Antart. 8, 207–216.

Prentice, M.L., Matthews, R.K., 1991. Tertiary ice sheet dynamics: the snow gun hypothesis. J. Geophys. Res. 96, 6811–6827.

Prothero, D.R., Berggen, W.A., 1992. Eocene–Oligocene Climatic and Biotic Evolution. Princeton Univ. Press, Princeton, NJ, USA. 568 pp.

Pudsey, C.J., 1992. Late Quaternary changes in Antarctic bottom water velocity inferred from sediment grain size in the northern Weddell Sea. Mar. Geol. 107, 9–33.

Pudsey, C.J., Howe, J.A., 1998. Quaternary history of the Antarctic Circumpolar Current: evidence from the Scotia Sea. Mar. Geol. 148, 83–112.

Pudsey, C.J., Howe, J.A., 2002. Mixed biosiliceous-terrigenous sedimentation under the Antarctic Circumpolar Current, Scotia Sea. In: Stow, D.A.V., Pudsey, C.J., Howe, J.A., Faugeres, J.-C., Viana, A.R. (Eds.), Deep-Water Contourites: Modern Drifts and Ancient Series, Seismic and Sedimentary Characteristics. Geol. Soc. Lond. Mem., vol. 22, pp. 325–336.

Pudsey, C.J., Barker, P.F., Hamilton, N., 1988. Weddell Sea abyssal sediments: a record of Antarctic bottom water flow. Mar. Geol. 81, 289–314.

Roberts, A.P., Wilson, G.S., Harwood, D.M., Versosub, K.L., 2003. Glaciation across the Oligocene–Miocene boundary in southern McMurdo Sound, Antarctica: new chronology from the CIROS-1 drill hole. Palaeogeogr. Palaeoclimatol. Palaeoecol. 198, 113–130.

Raine, J.I., Askin, R.A., 2001. Terrestrial palynology of Cape Roberts Project Drillhole CRP-3, Victoria Land Basin, Antarctica. Terra Antart. 8, 389–400.

Ramsay, A.T.S., Baldauf, J.G., 1999. A Reassessment of the Southern Ocean Biochronology. Geol. Soc. Lond. Mem., vol. 18, 122 pp.

Raymo, M., 1997. Carbon cycle models: how strong are the constraints? In: Ruddiman, W.F. (Ed.), Tectonic Uplift and Climate Change. Plenum, NY, pp. 367–380.

Smetacek, V., de Baar, H.J.W., Bathmann, U.V., Lochte, K., Rutgers van der Loeff, M.M. (Eds.), 1997. Ecology and Biogeochemistry of the Antarctic Circumpolar Current during Austral spring: Southern OceanJGOFS Cruise Ant X/6 of R.V. Polarstern. Deep-Sea Res. Part II, Top. Stud. Oceanogr., vol. 44 (1–2). 519 pp.

Stilwell, J.D., Feldmann, R.M., 2000. Paleobiology and Paleoenvironments of Eocene Rocks, McMurdo Sound, East Antarctica Ant. Res. Ser. American Geophysical Union, Washington, DC.

Troedson, A.L., Smellie, J.L., 2002. The Polonez Cove Formation of King George Island, West Antarctica; stratigraphy, facies and palaeoenvironmental implications. Sedimentology 49, 277–301.

Toggweiler, J.R., Bjornsson, H., 1999. Drake Passage and paleoclimate. J. Quat. Sci. 15, 319–328.

Tolstikov, E.I., 1966. Atlas Antarktiki I. Moscow, Glavnoe Upravlenie Geodezii I Kartografii, plate 103.

Webb, P.-N., 1979. Paleogeographic evolution of the Ross sector during the Cenozoic. Mem. NIPR, Spec. Issue 13, 206–212.

Wei, W., Wise Jr., S.W., 1992. Selected Neogene calcareous nannofossil index taxa of the Southern Ocean: biochronology, biometrics and paleoceanography. In: Wise Jr., S.W., Schlich, R., et al. (Eds.), Proc. ODP, Sci. Results, vol. 120. Ocean Drilling Program, College Station, TX, pp. 523–537.

Weissel, J.K., Hayes, D.E., 1972. Magnetic anomalies in the southeast Indian Ocean. Antarctic Oceanology II: the Australian–New Zealand Sector. AGU Antarct. Res. Ser., vol. 19, pp. 165–196.

Whitworth III, T., Peterson, R.G., 1985. Volume transport of the Antarctic Circumpolar Current from bottom pressure measurements. J. Phys. Oceanogr. 15, 810–816.

Zachos, J., Pagani, M., Sloan, L., Thomas, E., Billups, K., 2001. Trends, rhythms and aberrations in global climate, 65 Ma to present. Science 292, 686–693.

Zenk, W., 1981. Detection of overflow events in the Shag Rocks Passage, Scotia ridge. Science 213, 1113–1114.