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Changing Southern Ocean palaeocirculation and effects on global climate

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Abstract: Southern Ocean palaeocirculation is clearly related to the formation of a continental ice sheet on Antarctica and the opening of gateways between Antarctica and the Australian and South American continents. Palaeoenvironmental proxy records from Southern Ocean sediment cores suggest ice growth on Antarctica beginning by at least 40 million years (Ma) ago, and the opening of Tasmania–Antarctic and Drake Passages to deep-water flow around 34 and 31 ± 2 Ma, respectively. So, the Eocene/Oligocene transition appears to mark the initiation of the Antarctic Circumpolar Current and thus the onset of thermal isolation of Antarctica with a first major ice volume growth on East Antarctic. There is no evidence for a significant cooling of the deep ocean associated with this rapid (< 350 000 years) continental ice build-up. After a long phase with frequent ice sheets growing and decaying, in the middle Miocene at about 14 Ma, a re-establishment of an ice sheet on East Antarctica and the Pacific margin of West Antarctica was associated with an increased southern bottom water formation, and a slight cooling of the deep ocean, but with no permanent drop in atmospheric $p\mathrm{CO_2}$. During the late Pleistocene on orbital time scales a temporal correlation between changes in atmospheric $p\mathrm{CO_2}$ and proxy records of deep ocean temperatures, continental ice volume, sea ice extension, and deep-water nutrient contents is documented. I discuss hypotheses that call for a dominant control of glacial to interglacial atmospheric $p\mathrm{CO_2}$ variations by Southern Ocean circulation dynamics. Millennial to centennial climate variability is a global feature, but there is contrasting evidence from various palaeoclimate archives that indicate both interhemispheric synchrony and asynchrony. The role of the Southern Ocean, however, in triggering or modulating climate variability on these time scales only recently received some attention and is not yet adequately investigated.

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Key words: Antarctic Circumpolar Current, Eocence/Oligocene transition, Late Pleistocene, middle Miocene, palaeoceanography

Introduction

Ocean deepwater circulation plays an important role in climate modulation through its redistribution of heat and salt and its control of atmospheric CO_2. This paper aims to discuss the role of the Southern Ocean in global deepwater circulation and climate from a palaeoceanographic perspective as a bottom line for modern oceanographic, biogeochemical, and biological research on climate and human development. I will focus on three time periods with most pronounced changes in the Southern Ocean, the Eocene/Oligocene (E/O) boundary now at about 34 Ma, when a circumpolar current significantly influenced global circulation for the first time, the Middle Miocene around 14 Ma, when the reformation of an East Antarctic Ice Sheet (EAIS) influenced mode and level of Antarctic bottom water formation, and late Pleistocene times, when during the last 400 kyr changes in Southern Ocean circulation seem to correlate with the duration of seasonal sea ice coverage and atmospheric CO_2 draw down during glacial times. In addition, the relative timing of millennial and sub-millennial climate changes of the last glaciation between and within Hemispheres will be discussed. The overall aim of this review paper is to show that since the initiation of a circumpolar circulation at the E/O-boundary, and in particular since late middle Miocene times, after tectonic changes had developed near-modern basin configurations, the Southern Ocean is an excellent monitor for deepwater circulation changes, because of its central role as a mixing reservoir of incoming Northern Component Water and recirculated water from the Pacific and Indian oceans and as the site of major Southern Component Water formation. A second aim is to illustrate that, although major progress has been made in understanding climate variability in the past and how this is documented in ocean sediments, a still enormous and combined interdisciplinary research effort is necessary to decipher the exact timing and forcing mechanisms, the knowledge of which will be the prerequisite for predicting natural changes in the near future as a zero line for detection of and reaction to anthropogenic disturbances.

Proxies used in palaeoceanography

To provide those who are not familiar with the palaeoceanographic community with a basic background for the issues discussed below, the essentials of the most popular proxies used to infer ocean palaeocirculation and palaeoproductivity patterns are briefly summarized here. All of these proxies have their advantages and disadvantages, which cannot be discussed in detail, but which are addressed in the specific palaeoceanographic papers that develop, calibrate, and use these proxies. The reader is referred to these original contributions for proper credit and to recent reviews in Lynch-Stieglitz (2003) and Lea (2003).

Oxygen isotope ratios ($^{18}O/^{16}O$) of foraminiferal tests, which are planktonic and benthic calcium carbonate microfossils widely distributed in Cenozoic sediments, depend on temperature and isotopic composition of the water in which the tests were calcified. The water $\delta^{18}O$, in turn, reflects the amount of water bound in continental ice. Generally, $\delta^{18}O$ values of benthic foraminiferal calcite increase with decreasing water temperatures and increasing ice volume on the continents (Shackleton & Opdyke 1973).

Carbon isotope ratios ($^{13}C/^{12}C$) of benthic foraminiferal tests from ocean transects are used to compare $\delta^{13}C$ gradients in deep and bottom waters and hence their relative ages. Nutrient contents and $\delta^{13}C$ of the dissolved inorganic carbon (DIC) in seawater are negatively correlated, since in surface water masses during photosynthetic carbon fixation nutrients and preferentially the light carbon isotopes are utilized. Consequently nutrients get depleted and $\delta^{13}C_{DIC}$ values enriched. As deep water masses mix and gain remineralized carbon from surface productivity during their oceanic transit, the $\delta^{13}C_{DIC}$ values decrease as does the $\delta^{13}C$ of benthic foraminiferal calcite sequestered from these waters. Generally, benthic foraminiferal $\delta^{13}C$ values decrease with increasing nutrient contents and increasing age of a deep water mass, i.e. since the time of isolation from the atmosphere (Curry *et al.* 1988).

Cd/Ca ratios in foraminiferal tests reflect Cd water concentrations in which they were grown. Dissolved Cd in seawater, in turn, is correlated with dissolved phosphate concentrations. Thus Cd/Ca ratios of benthic and planktonic foraminifera are extensively used as a nutrient proxy (Boyle 1988). Only recently it has been found that the incorporation of Cd into planktonic foraminifera relative to seawater is temperature sensitive. Generally Cd/Ca ratios in foraminiferal tests increase with increasing nutrient contents of the seawater.

$\Delta\delta^{13}C_{DIC}$ is the $\delta^{13}C$ value the dissolved inorganic carbon of a water mass would have if phosphate were entirely removed. Thus no biologically driven fractionation would mask the sense and magnitude of the isotopic fractionation during gas exchange between the ocean surface and the atmosphere, frequently referred to as thermodynamic imprint (Broecker & Maier-Reimer 1992). The subtraction of the biological effect leaves the thermodynamic imprint as a conservative tracer after the water mass is out of contact with the atmosphere.

Sedimentary $^{231}Pa/^{230}Th$ ratios, as quoted in this paper, are used as a proxy for changes in deep-water circulation (Yu *et al.* 1996). The method is based on the fact that ^{231}Pa and ^{230}Th nuclides are produced by radioactive decay of ^{235}U and ^{234}U isotopes, respectively, in seawater. The rate of production of these two radionuclides is uniform throughout the water column, and they both are particle-reactive and rapidly scavenged to the sea floor. Since ^{230}Th is considerably more particle-reactive than ^{231}Pa, it has a much shorter residence time in the water column. Therefore,

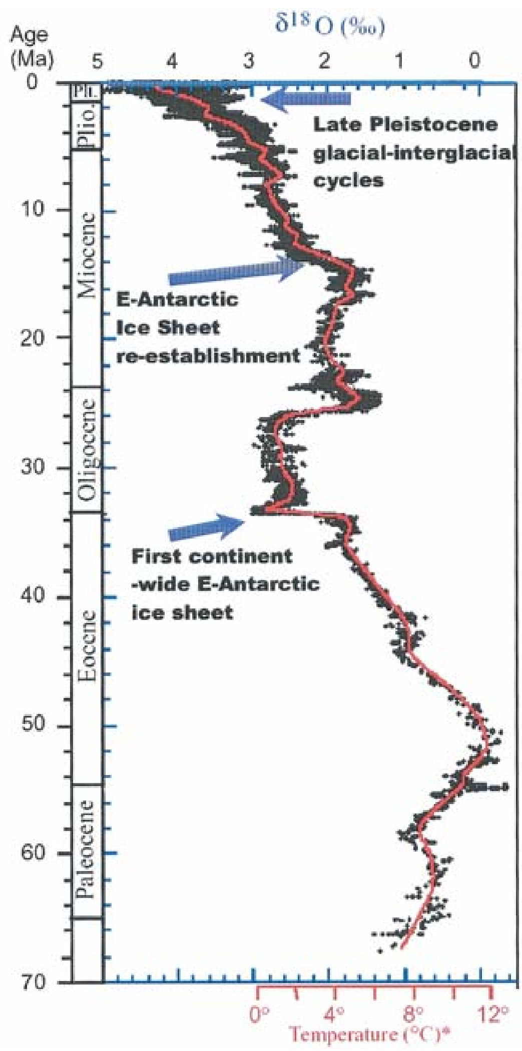

Fig. 1. Composite $\delta^{18}O$ record based on mainly two benthic foraminiferal taxa as compiled by Zachos *et al.* (2001) from more than 40 deep sea sites. The temperature scale was calculated for an ice-free ocean and thus only applies to the time prior to about 40–35 Ma. The three main time periods this paper is focussed on are indicated by bold blue arrows (modified from Zachos *et al.* 2001).

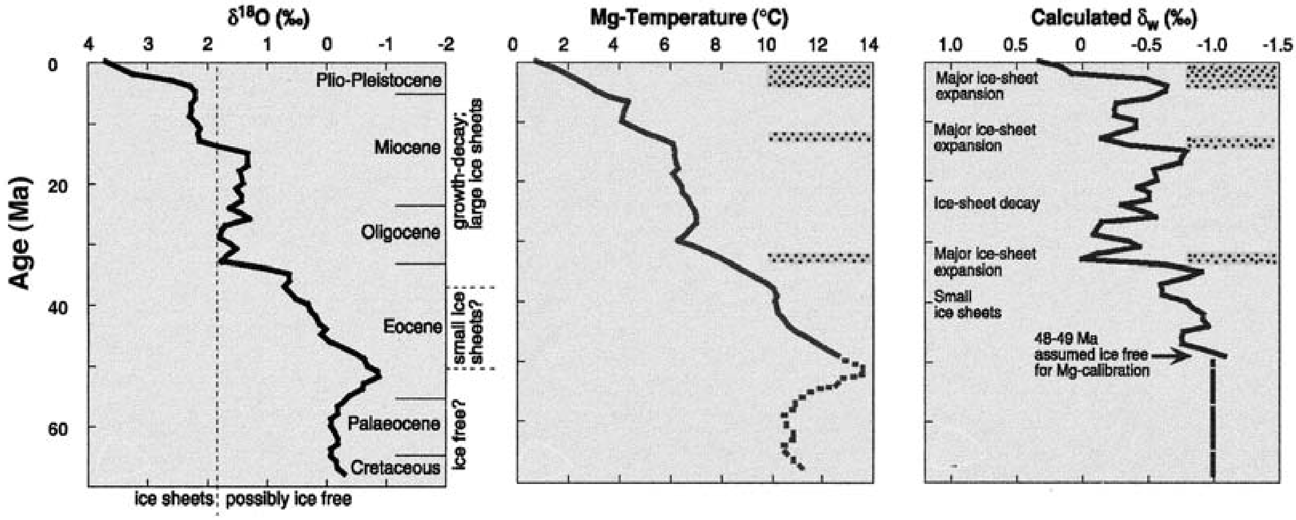

Fig. 2. (left) Composite $\delta^{18}O$ record based on Atlantic cores as compiled by Miller *et al.* (1987). (centre) Mg temperature record based on multispecies benthic foraminifera from three deep-sea sites as adjusted and calculated by Lear *et al.* (2000). Broken line indicates temperatures calculated from the $\delta^{18}O$ record assuming an ice-free world. Stipple areas indicate periods of substantial ice sheet growth determined from the benthic foraminiferal $\delta^{18}O$ record and the Mg temperature. (right) Estimated variation in $\delta^{18}O$ of seawater, a measure of global ice volume, calculated by substituting Mg temperature and benthic $\delta^{18}O$ data into the “palaeotemperature equation” (modified from Lear *et al.* 2000).

$^{231}Pa/^{230}Th$ ratios in surface sediments vary widely from their constant production ratio and depend on deep-water advection and scavenging processes (biological productivity) in the water column. Generally sedimentary $^{231}Pa/^{230}Th$ ratios can be interpreted in terms of residence times and ventilation of deep-water masses in the oceans as part of thermohaline circulation.

Nitrogen isotope ratios ($^{15}N/^{14}N$) of marine organic matter produced in the euphotic zone are used to reconstruct ocean productivity and nutrient utilization in ocean surface waters (Altabet & Francois 1994). This is based on the fact that assimilation of nitrate by phytoplankton is accompanied by nitrogen isotope fractionation, leaving organic matter depleted in ^{15}N generally depending on the isotope ratio of dissolved inorganic nitrate and the degree to which this pool is utilized.

Mg/Ca ratios of benthic foraminiferal tests have become an increasingly popular proxy to reconstruct palaeotemperatures (Nürnberg 1995). In conjunction with foraminiferal $\delta^{18}O$ values it can be used to assess changes in the $\delta^{18}O$ of seawater through time. The basis for this method lies in the observations that

i) the partition coefficient of Mg^{2+} ions into inorganic calcite strongly correlates with temperature, and

ii) of a temperature-dependent uptake of Mg in biogenic calcite .

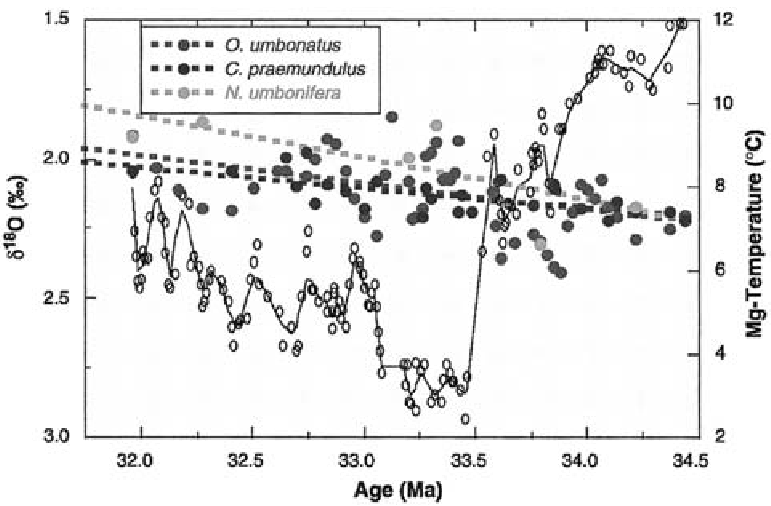

Fig. 3. Mg temperatures based on Mg/Ca ratios of three benthic foraminiferal species across the Eocene/Oligocene transition (solid symbols) plotted together with a monogeneric $\delta^{18}O$ record from Zachos *et al.* (1993) (open symbols) at DSDP Site 522. All curves represent a three-point running average of the data. This detailed view of the major climatic transition at the E/O boundary does not show any decrease in Mg derived bottom water temperature across the major $\delta^{18}O$ excursion, so the observed abrupt $\delta^{18}O$ increase within about 350 ka must reflect continental ice build-up (modified from Lear *et al.* 2000).

The *boron-isotope ($\delta^{11}B$)* approach to $p\mathrm{CO_2}$ estimation relies on the fact that a rise in the atmospheric concentration of CO_2 causes a reduction in the surface ocean pH. Because the boron isotopic composition of marine carbonates is highly dependent on pH, δ^{11}B values of fossil foraminiferal calcite are sensitive pH indicators of ancient sea water (Sanyal *et al.* 1995). For a given pH of sea water the aqueous CO_2 concentration can be calculated and thereby the atmospheric $p\mathrm{CO_2}$ can be quantitatively estimated.

The *alkenone $\delta^{13}C$* approach to $p\mathrm{CO_2}$ estimation is based on the observation that during photosynthetic carbon fixation in the ocean mixed layer, a rise in atmospheric $p\mathrm{CO_2}$ enhances the difference between $\delta^{13}C_{DIC}$ values and the organic δ^{13}C of phytoplankton. The δ^{13}C of fossil planktonic foraminiferal tests and of alkenones of prymnesiophyte algae are measured as proxies for $\delta^{13}C_{DIC}$ and δ^{13}C of phytoplankton organic carbon, respectively (Jasper & Hayes 1990).

Ice-rafted debris (IRD), i.e. terrigeneous particles in the sand and gravel size fraction of deep-sea sediment, is supposed to be derived from the calving of icebergs of continental ice sheets (e.g. Connolly & Ewing 1965). There are various simple methods in use to identify and quantify ice-drop activity, the more sophisticated ones of which include counting of grains > 2 mm having the typical sharp angular surface under a binocular, and from core x-radiographs (e.g. Grobe & Mackensen 1992).

The *clay mineral composition* of the < 2 µm fraction of deep sea sediments close to continents are mainly derived from land, and were formed as a result of various weathering processes, which ultimately were controlled by climate. The clay mineral types and the proportions of the individual clay minerals in marine sediments therefore depend on the weathering (climatic) conditions on land and on the nature of the source rocks (Biscaye 1965).

Variations in *$^{87}Sr/^{86}Sr$ ratios* in planktonic foraminifera are thought to reflect changes in the flux of Sr to seawater from mid-ocean ridge basalts and continental rocks. As used in this paper it is supposed that Sr fluxes from the continents mainly depend on the weathering conditions on land and on the nature of the source rocks (Palmer & Edmond 1992).

Al/Ti ratios calculated from the total concentrations of these metals in sediment samples can be used to identify source types of detrital material, as different rock types have different Al/Ti ratios (e.g. Latimer & Filippelli 2002). Generally, Al/Ti ratios may provide information about whether detrital sources are continental or oceanic.

The *Sortable Silt Mean Size* proxy, i.e. the mean grain size of the 10–63 µm terrigenous sediment fraction with biogenic carbonate and opal removed, is an indicator of bottom current flow speed (McCave *et al.* 1995). Generally, coarser values of the sortable silt mean size reflect stronger near-bottom flow and selective deposition and winnowing.

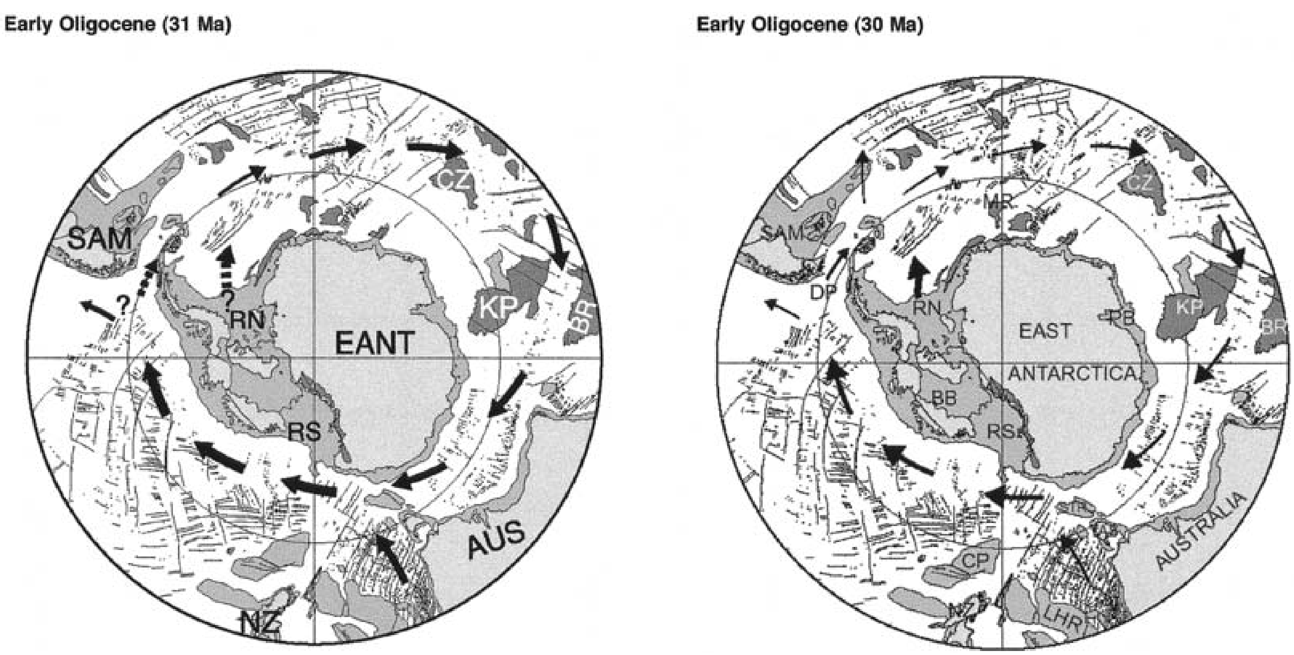

Fig. 4. Plate tectonic configuration at 31 and 30 Ma as reconstructed by Lawver & Gahagan (2003). According to these authors the final opening to deep water circulation through Drake Passage (DP) took place at 31 ± 2 Ma (compiled from Lawver & Gahagan 2003). AUS = Australia, BB = Bentley Trough–Byrd Basin, BR = Broken Ridge, CP = Campbell Plateau, CZ = Crozet Plateau, KP = Kerguelen Plateau, LHR = Lord Howe Rise, MR = Maud Rise, NZ = New Zealand, PB = Prydz Bay, RN = Ronne Embayment (Weddell Sea region), RS = Ross Sea, TAS = Tasmania.

This proxy indicates relative current speed only.

Initiation of Antarctic Circumpolar Current

Formation of continental ice on Antarctica

Since the first Cenozoic oxygen isotope records from the deep ocean indicated an increase in $\delta^{18}O$ values exceeding the plausible range that could be attributed to a temperature decrease, it is clear that huge continental ice caps varying in volume and extension, must have existed in high latitudes (Savin *et al.* 1975, Shackleton & Kennett 1975). The growing amount of data available led to ocean-wide compilations that indicate more or less progressive cooling since an Early Eocene climatic optimum and a first major glaciation of Antarctica at the Eocene/Oligocene boundary (Miller *et al.* 1987, Zachos *et al.* 2001) (Fig. 1).

Only recently the further development and better calibration of Mg/Ca ratios of benthic foraminiferal tests as an additional temperature proxy, independent from continental ice volume, enabled the $\delta^{18}O$ seawater component to be calculated, and thus a quantification of the ice caps and identification of times of rapid continental ice growth (Lear *et al.* 2000, 2002) (Fig. 2). Based on Mg/Ca-derived temperatures and the corresponding reconstructions of the oxygen isotopic composition of seawater, Billups & Schrag (2003) suggest ice growth on Antarctica at 40 Ma. There is supporting direct evidence of a regional glaciation since about that time on East Antarctica by the record of ice-rafted debris (Ehrmann 1991, Breza & Wise 1992) and the shift in the clay mineral composition of sediments from the Antarctic continental margin (Ehrmann & Mackensen 1992), as well as more indirect evidence by variations in Strontium isotope ratios ($^{87}Sr/^{86}Sr$) in planktonic foraminifera (Zachos *et al.* 1999). There is agreement, however, that the first continent-wide ice sheet on Antarctica formed at the Eocene/Oligocene boundary at about 34 Ma (e.g. Mackensen & Ehrmann 1992, Barrett 1996, 2003, Zachos *et al.* 1996). Both Lear *et al.* (2000) and Billups & Schrag (2003) do not observe a global cooling across the Eocene/Oligocene boundary, so the benthic foraminiferal $\delta^{18}0$ increase reflects only ice volume growth (Fig. 3).

Opening of ocean gateways

In a world without continent-wide glaciations, in the Southern Hemisphere an ocean circulation dominated with primarily north–south flows in the Pacific, Atlantic and Indian oceans (Kennett 1978, Stott *et al.* 1990, Barrett 2003). The separation of Antarctica from Australia and South America opened ocean gateways, i.e. Tasmanian and Drake passages, respectively, that allowed current systems to flow west–east around Antarctica. This caused thermal isolation of Antarctica through a reduction in heat transfer

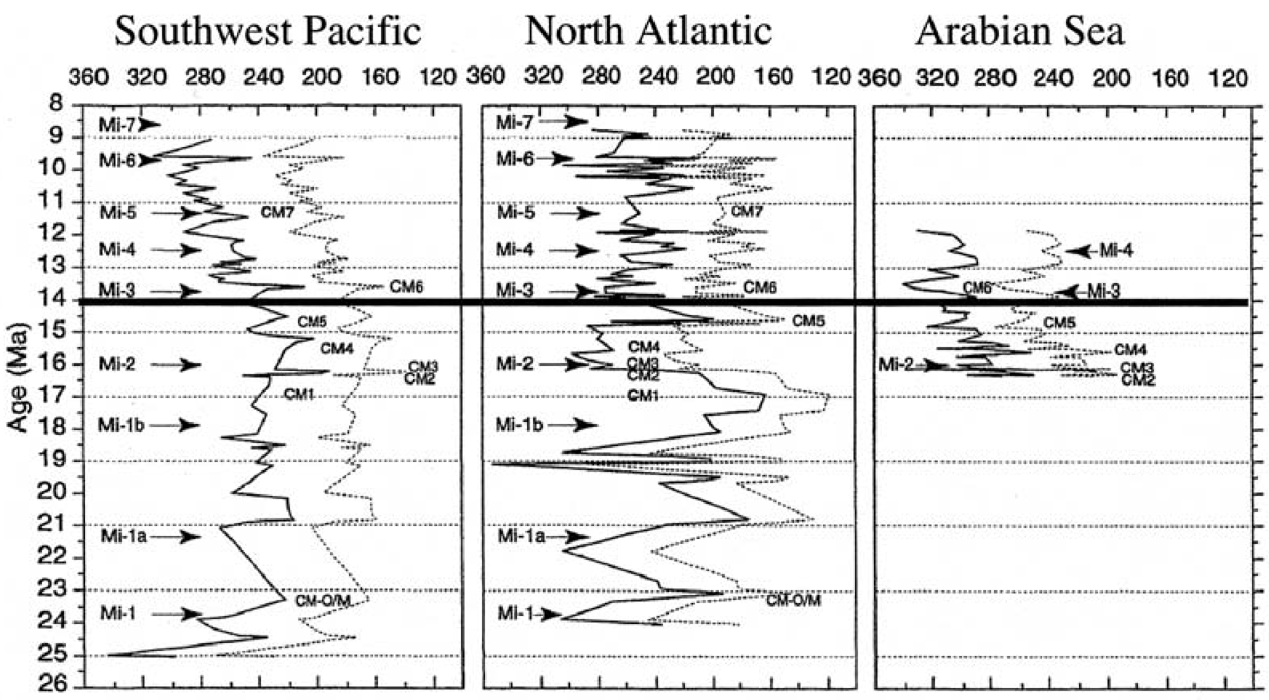

Fig. 5. Atmospheric pCO_2 estimates based on the alkenone $\delta^{13}C$ approach (Pagani *et al.* 1999). Solid and dotted lines represent maximum and minimum estimates, respectively. CM and Mi represent carbon maxima events (Woodruff & Savin 1991) and inferred glacial maxima (Miller *et al.* 1991, Wright *et al.* 1992). Horizontal line at about 14 Ma indicates $\delta^{18}O$ minimum after the middle Miocene climate optimum (modified from Pagani *et al.* 1999).

from low latitudes to the South and by this permitted a continent-wide glaciation on Antarctica and sea ice formation around it (Kennett 1977).

Benthic foraminiferal and stable isotope data from several Southern Ocean drilling sites documented major productivity changes in the late Eocene and early Oligocene, which were interpreted to indicate an opening of Drake Passage for intermediate water flow and the initiation of some kind of a circum-Antarctic circulation as early as about 37 Ma (Diester-Haass & Zahn 2001). However, reconstructions of plate motions based on dated sea-floor spreading anomalies and distinct fracture zone lineations suggest that the prerequisite for an oceanic circum-Antarctic current, the opening of both gateways between the South Tasman Rise and Antarctica and between South America and Antarctica for deep water, took place at about the same time, namely at about the Eocene/Oligocene boundary (Kennett 1977, Lawver & Gahagan 1998). Although there is general agreement on the timing of opening of the Tasmanian Passage at about 34 Ma (Exon *et al.* 2001), with regard to Drake Passage, however, plate tectonic evidence is more circumstantial, since the oldest magnetic anomalies found in the western Scotia Sea are about 5 Ma younger and the exact motions of continental fragments are not known (Barker & Burrell 1982). Although it is widely accepted that opening of the Drake Passage by seafloor spreading was initiated some time prior to at least 29 Ma, it was argued that deep-water flow was inhibited by continental crust fragments along the NW–SE running Shackleton Fracture Zone until about 23 Ma (Barker & Burrell 1977). On the contrary, Lawver & Gahagan (2003), based on reasonable assumptions concerning motion of the crustal fragments in the Scotia Sea, recently concluded that Drake Passage was open to deep water circulation by about 31±2 Ma (Fig. 4). This is independently supported by a new model for the opening of Drake Passage using geophysical and geochemical evidence that suggests an opening of Drake Passage to deep water connection flow by early Oligocene times (Livermore *et al.* 2004). It is shown that the Shackleton Fracture Zone, along which continental fragments were suspected to block deep-water flow during the early Oligocene, in fact is an oceanic transverse ridge that was formed by uplift related to compression across the fracture zone only since about 8 Ma. Hence, Livermore *et al.* (2004) conclude that there was no barrier to deep circulation through Drake Passage prior to 8 Ma and a deep-water connection between the Pacific and Atlantic oceans was probably established soon after spreading began in Drake Passage, the time of which is not disputed and constrained by the age of ocean floor in Drake Passage of older than about 29 Ma (Barker & Thomas 2004). Additional sedimentological evidence was published by Latimer and & Filipelli (2002), who inferred from changes in Al/Ti ratios in sediments from the Atlantic sector of the Southern Ocean (ODP Site 1090), a metal source change from continental crust to oceanic crust at the Eocene/Oligocene transition. Together with a permanent change in the Ba concentration record, Latimer & Filipelli (2002) interpret this to reflect a change in deep-water circulation at around 33 Ma, i.e. a bathing of this eastern Atlantic site by water of Pacific origin which at this location, would point to the opening of Drake Passage. In a comprehensive investigation of a suite of biogenic and terrigenous sedimentary components of this same site Diekmann *et al.* (2004) report reduced sediment

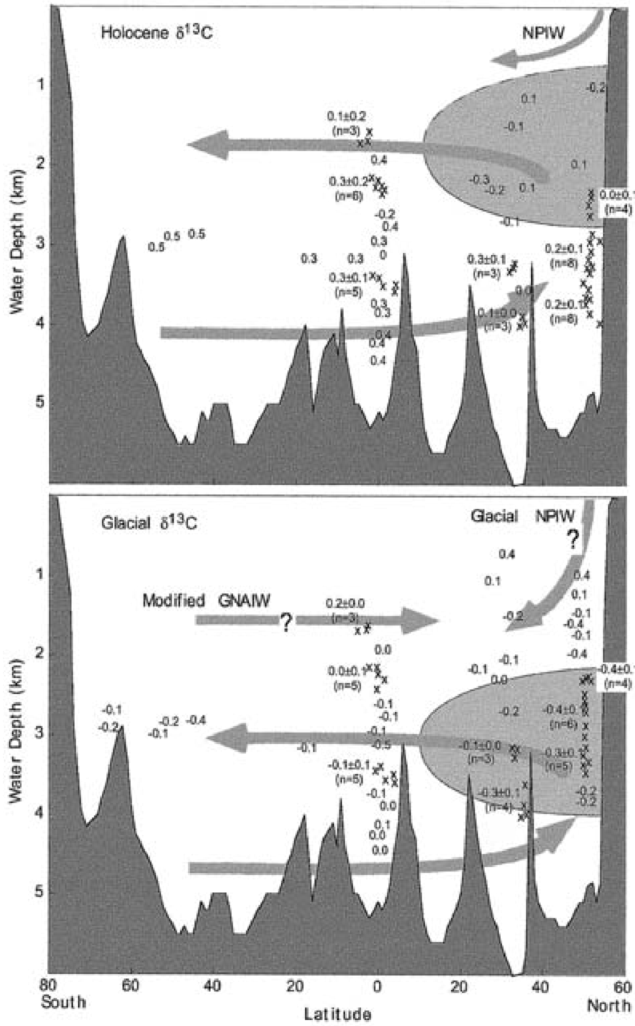

Fig. 6. Meridional distribution of benthic foraminiferal $\delta^{13}C$ in the western Pacific Ocean as compiled by Matsumoto *et al.* (2002) based on data from Duplessy *et al.* (1988), Keigwin (1998), Herguera *et al.* (1992), Matsumoto & Lynch-Stieglitz (1999) and Matsumoto *et al.* (2001). The lowest $\delta^{13}C$ indicated by shading is found in the deep North Pacific centred around 2000 m water depth during the Holocene and around 3000 m during glaciation. Arrows indicate suggested circulation. The higher $\delta^{13}C$ in the upper 2000 m in the North Pacific reflect a local source of newly ventilated glacial North Pacific Intermediate Water (NPIW), influx of modified Glacial North Atlantic Intermediate Water (GNAIW), or both.

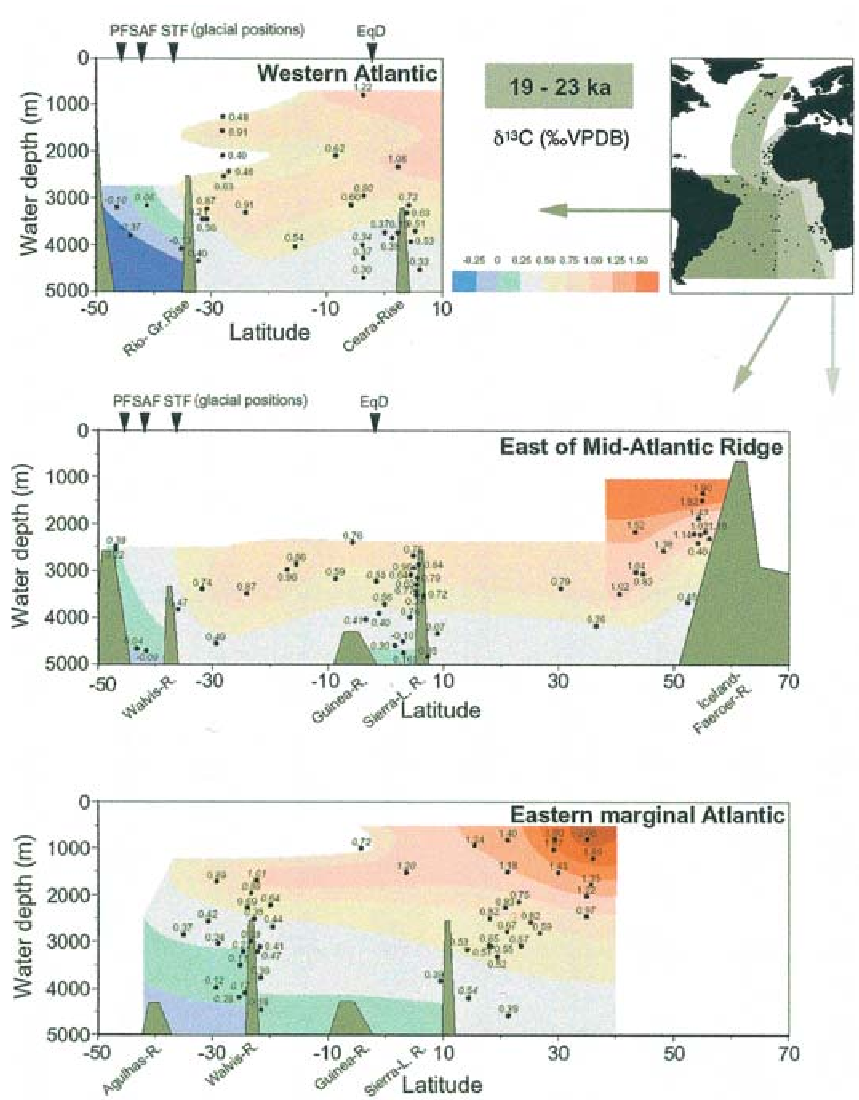

Fig. 7. Last Glacial Maximum (19–23 ka) distribution of benthic foraminiferal $\delta^{13}C$ along three vertical sections representing the western South Atlantic, the central Atlantic east of the Mid-Atlantic Ridge, and the eastern marginal Atlantic as compiled by Bickert & Mackensen (2004), based on the data from Sarnthein *et al.* (1994), Bickert & Wefer (1996) and Mackensen *et al.* (2001). Modified from Bickert & Mackensen (2004). PF = Polar Front, SAF = Sub-Antarctic Front, STF = Subtropical Front.

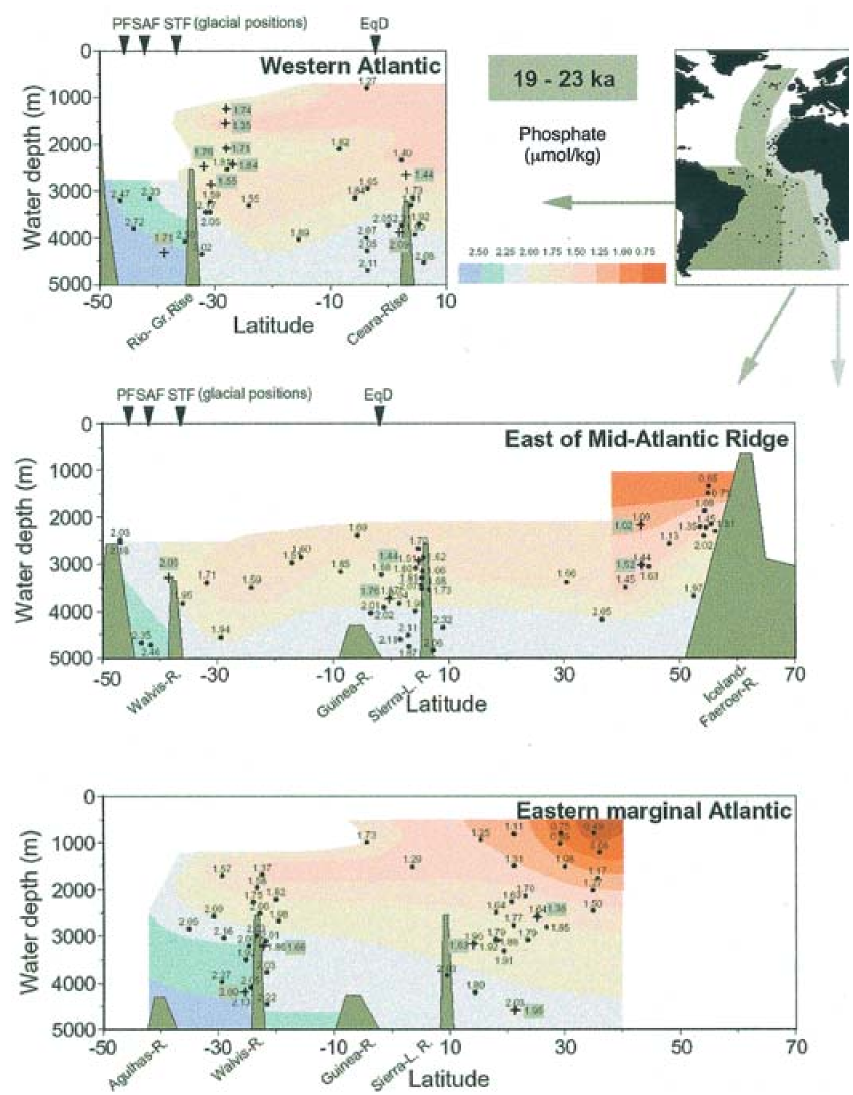

Fig. 8. Last Glacial Maximum (19–23 ka) phosphate distribution calculated from benthic $\delta^{13}C$ and from Cd_w (Boyle 1992, shaded values) along three vertical sections representing the western South Atlantic, the central Atlantic east of the Mid-Atlantic Ridge, and the eastern marginal Atlantic. Modified from Bickert & Mackensen (2004). PF = Polar Front, SAF = sub-Antarctic Front, STF = Subtropical Front.

accumulation between 33.4 and 30.2 Ma and a 1.5 Ma hiatus around 32 Ma, the latter of which was also identified as a pronounced unconformity in seismic profiles from this area (Wildeboer Schut *et al.* 2002). Unconformities in the lower Oligocene can be traced throughout the Southern Ocean and demonstrate the invigoration of bottom currents, which would be consistent with a widening and deepening of an open Drake Passage until about 30 Ma and the establishment of a deep-reaching Antarctic Circumpolar Current (Diekmann *et al.* 2004).

Atmospheric CO_2

Since Mg/Ca palaeothermometry does not indicate significant cooling associated with the rapid increase in ice volume (Lear *et al.* 2000) it is evident that more than the opening of gateways and the subsequent thermal isolation of Antarctica must be considered as forcing mechanism for the initiation of a continent-wide ice build-up (Zachos *et al.* 2001). Changes in the carbon dioxide concentration of the atmosphere is widely regarded as a likely forcing mechanism on global climate because of its large effects on temperature. Reconstructions of Cenozoic pCO_2 based on boron and alkenone isotope approaches (Pagani *et al.* 1999, Pearson & Palmer 2000) indicate that atmospheric CO_2 levels were still high at 40 Ma when the first ice sheets started growing, but were declining or already low during times of terminating ice accumulation during the late Oligocene. Unfortunately there is as yet no reliable pCO_2 record available across the E/O boundary. The fact that ice sheet growth terminated during times of low CO_2, however, may reinforce the notion that moisture supply was the critical element in maintaining large polar ice sheets (for further discussion see Zachos *et al.* 2001). Based on numerical studies using global general circulation models, it was questioned whether thermal isolation or declining atmospheric CO_2 is the primary factor for the formation of a continent-wide ice sheet on East Antarctica (Mikolajewicz *et al.* 1993, DeConto & Pollard 2003). Other models, however, suggest that even an opening to shallow depths was sufficient to bring about a significant cooling of southern high latitudes and increase in Antarctic sea ice (Sijp & England (2004) as quoted in Livermore *et al.* (2004)). A general circulation model with coupled components for atmosphere, ocean, ice sheet and sediments, suggests that a cooling due to declining CO_2 partial pressure would have gradually lowered annual snowline elevations until extensive regions of high Antarctic topography were ice-covered (DeConto & Pollard 2003). Once some threshold was reached, feedbacks related to snow/ice-albedo and ice sheet height/mass-balance could have initiated rapid ice sheet growth during orbital periods favourable for the accumulation of glacial ice. DeConto & Pollard (2003) also report a model experiment that compared effects on simulated ice sheet growth of an open and closed Drake Passage. In spite of a 20% reduction in southward heat transport resulting from Drake Passage opening, the model did not show a significant change in the pattern or magnitude of ice sheet formation at the E/O boundary. Accordingly, it was suggested that Southern Ocean gateways play only a secondary role in the ice sheet build-up on Antarctica.

In summary, it can be stated that ice sheets have been present on Antarctica since 40 Ma, the Tasmania–Antarctic Passage was open to deep waters at 34 Ma, and Drake Passage most likely at 31 ± 2 Ma. The initiation of a circum-Antarctic circulation at about the E/O boundary marks the onset of thermal isolation of Antarctica with a first continent-wide glaciation. This may explain the initial appearance of continent-wide Antarctic ice sheets, but fails to explain the subsequent termination. The role of declining atmospheric pCO_2 in the build-up of the first Antarctic ice sheet at the E/O boundary is not yet constrained.

Middle Miocene ice sheet re-establishment

After the first continental glaciation at the E/O boundary, a long phase followed with ice sheets frequently growing and decaying but in only a few places did they reach the continental margin (Barrett 1996). The second most prominent increase in benthic $\delta^{18}O$ values at about 14 Ma was associated with polar cooling, major growth of the EAIS, a large drop in global sea level and major changes in ocean circulation (Shackleton & Kennett 1975, Woodruff & Savin 1989, 1991, Miller *et al.* 1991, 1998, Flower & Kennett 1994, 1995). Based on benthic foraminiferal faunal assemblage changes and $\delta^{18}O$ as well as $\delta^{13}C$ values Woodruff & Savin (1989, 1991) suggested that during early Miocene time a warm saline plume flowed from the eastern Tethys into the Indian Ocean. As this Tethyan Indian Saline Water (TISW) flowed southward towards Antarctica it entrained large volumes of relatively shallow, warm Indian Ocean water in a manner analogous to the modern Mediterranean outflow into the Atlantic. The southward flowing Indian Ocean Water joined the circumpolar circulation, upwelled, was refrigerated, and sank to form the early Miocene analogue of Antarctic Bottom Water (AABW), the Southern Component Water (SCW). Northern Component Water (NCW), the analogue of North Atlantic Deep Water (NADW), likely existed during the early Miocene, based on the $\delta^{13}C$ difference between Atlantic and Pacific oceans, but decreased during the middle Miocene when SCW production increased in strength (Woodruff & Savin 1989, 1991, Wright *et al.* 1992). Based on mean sortable silt analyses that clearly show long-term changes in flow speed of the SCW entering the Pacific, Hall *et al.* (2003) inferred an increase in SCW production between about 15.5 and 13.5 Ma. Meridional heat transport by the TISW may have prevented large-scale growth of the EAIS through the early Miocene until about 14 Ma. The termination of TISW, and the decreased meridional heat transport by the closure of the eastern portal of the Tethys Ocean in the present day eastern Mediterranean at about 14 Ma, may have led to further cooling of Antarctic surface waters and, in conjunction with increased SCW production, fostered the rapid growth of an EAIS (Woodruff & Savin 1989, Wright *et al.* 1992, Flower & Kennett 1994).

The intensification of Antarctic bottom water formation would have been accompanied by accelerated upwelling of nutrient-rich water which might have boosted productivity and caused a temporary reduction in pCO_2. Early to middle Miocene $\delta^{13}C$ records indeed reflect large-scale changes in organic carbon deposition relative to carbonate sedimentation, and together with wide-spread, large sedimentary organic carbon and phosphatic deposits may indicate a consequent drawdown of atmospheric CO_2 (Vincent & Berger 1985, Berger & Vincent 1986). This is partly corroborated by proxy estimates of atmospheric CO_2 based on the alkenone isotope approach insofar as lower intervals of pCO_2 correspond to inferred organic carbon burial events, but there is no evidence for a sharp CO_2 decrease associated with East Antarctic ice sheet growth (Pagani *et al.* 1999) or for a large permanent drop in pCO_2 during the middle Miocene (Pearson & Palmer 2000) (Fig. 5).

However, Mg/Ca derived deep sea temperature estimates indicate that about 85% of the $\delta^{18}O$ increase can be attributed to an increase in continental ice volume (Lear *et al.* 2000, Billups & Schrag 2003), which means just a slight deep ocean cooling of at most 1°C. This might suggest that a process other than a decline in Atlantic meridional heat transport may have promoted the EAIS growth, and is in support of recent seismic evidence from the Ross Sea continental shelf indicating that a West Antarctic Ice Sheet periodically contained significant ice volume during the middle Miocene, at least on the Pacific margin of West Antarctica (Chow & Bart 2003). The number of West Antarctic Ice Sheet grounding events in the Ross Sea is consistent with that which could be deduced from $\delta^{18}O$ and eustatic records.

In summary it can be stated that tectonic changes in the eastern Tethys Ocean may have led to a cooling of Antarctic surface waters and, in conjunction with an increased formation of Southern Ocean bottom waters, fostered a rapid growth of continental ice sheets, and ultimately resulted in a re-establishment of the East Antarctic Ice Sheet and the first build-up of a significant West Antarctic Ice Sheet. A slight cooling of the deep ocean of about 1°C is documented, but there is no evidence for a distinct atmospheric CO_2 decrease with the ice sheet growth or for a large permanent drop in pCO_2 during the middle Miocene.

The Southern Ocean in late Pleistocene times

The last of the rapid increases in the Cenozoic benthic foraminiferal $\delta^{18}O$ record reflects the Pleistocene onset of the Northern Hemisphere glaciation (Miller *et al.* 1987) which is testified by the massive appearance of IRD in northern high latitude oceans since about 2.7 Ma (e.g. Maslin *et al.* 1998). At about 0.9 Ma a switch from orbital obliquity (41 kyr cyclicity) to eccentricity (100 kyr cyclicity) driven global climate oscillations is documented in the benthic $\delta^{18}O$ record. It is generally agreed upon that this 100 kyr cycle represents a major component of the record of changing ice volume in the Northern Hemisphere, although it is difficult to explain this predominant cycle in terms of orbital eccentricity and direct forcing through insolation changes (Imbrie *et al.* 1993, Shackleton 2000). Recently Ruddiman (2003) suggested that the strong 100 kyr power observed in global climatic responses such as benthic $\delta^{18}O$, results from external orbital pacing of climate-system interactions initiated by orbital forcing at the periods of obliquity and precession (23 kyr cyclicity). Accordingly, greenhouse gases enhance insolation forces at the period of precession, driving the system to an interglacial state, and amplify ice sheet responses at the period of obliquity driving it towards a glacial state.

Glacial/interglacial deep and bottom water circulation changes

In the North Atlantic region Cd/Ca and $\delta^{13}C$ values of benthic foraminifera unanimously suggest that NADW during glacials was partially replaced by nutrient-rich waters of southern origin, overlain by a nutrient-depleted glacial NADW (Boyle & Keigwin 1987, Curry *et al.* 1988, Duplessy *et al.* 1988, Sarnthein *et al.* 1994).

In contrast to the North Atlantic, deep water geometry of the glacial Southern Ocean and South Atlantic is less well understood, partly because relatively few sites have been investigated south of the equator, and because of the lack of sites from the Antarctic polar frontal system not influenced by seasonal high productivity and its effect on benthic foraminiferal $\delta^{13}C$ (Mackensen *et al.* 1993) and carbonate corrosive bottom waters and its effect on Cd/Ca values (McCorkle *et al.* 1995). For similar reasons, our state of knowledge of the glacial Pacific deep water circulation is limited (Matsumoto *et al.* 2002). It is clear, however, that in the Pacific the water mass above about 2 km was distinct in its nutrient content from waters below this level (Fig. 6), but there is yet no consensus even on whether deep waters were flowing northward or southward.

As a result of a comprehensive compilation of benthic stable isotope records from the South Atlantic, including new as well as published data of Oppo & Horowitz (2000), Mackensen *et al.* (2001), and of North Atlantic data as compiled by Sarnthein *et al.* (1994), Bickert & Mackensen (2004) suggest no significant shift of NADW to intermediate depth during the Last Glacial Maximum (LGM). Instead, a core of a ^{13}C-enriched water mass

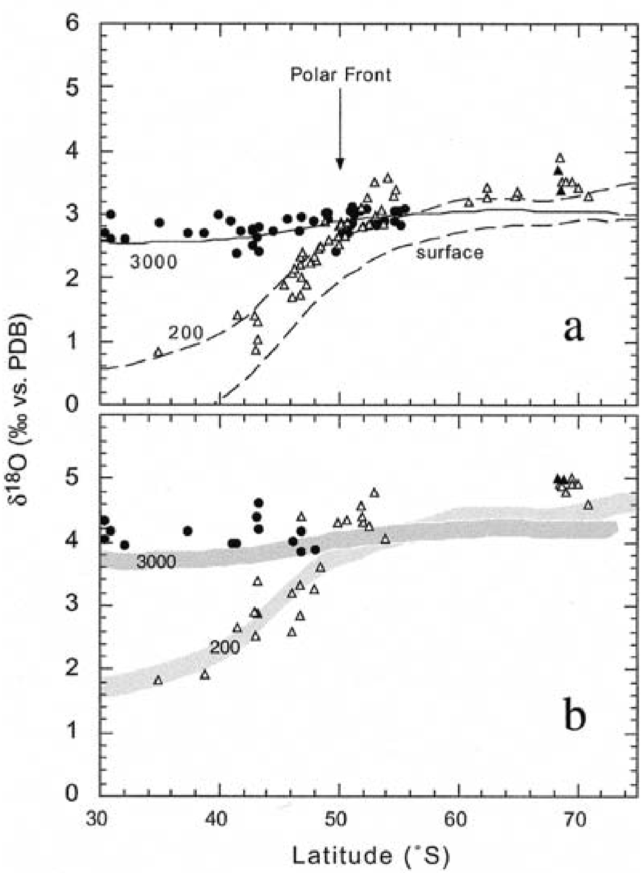

Fig. 9. South Atlantic **a.** Holocene, and **b.** LGM latitudinal distributions of planktonic (open triangles) and benthic (solid circles and triangles) foraminiferal $\delta^{18}O$ values. Calcite $\delta^{18}O$ in equilibrium with modern ambient seawater was calculated for the Holocene at the surface and at 200 m (dashed lines) using summer hydrographic data and at 3000 m (solid line) using annual mean data. Thick shaded lines in **b.** represent the 200 and 3000 m equilibrium calcite $\delta^{18}O$ lines shifted by between +1.0% and by +1.3% according to the range of the global glacial increase in $\delta^{18}O$. The approximate average position of the Polar Front (arrow) did not move significantly between the LGM and the Holocene which is indicated by a similar physical water mass stratification. Modified from Matsumoto *et al.* (2001) mainly based on data from Charles & Fairbanks (1990), Charles *et al.* (1996), and Mackensen *et al.* (1989, 1993, 1994).

spreading southward to at least 30°S between 1200 and 1900 m points to a source close to the Strait of Gibraltar, indicated by $\delta^{13}C$-values of up to 1.8‰. As proposed previously (Zahn *et al.* 1987, Zahn & Mix 1991), Bickert & Mackensen (2004) interpret this layer as an extended tongue of Mediterranean Outflow Water. Below this nutrient depleted water mass, a separate layer of glacial NADW is shown to have flowed southward at about the same depth interval or even deeper than it does today, although slightly more depleted and less extended in the water column (Fig. 7).

The admixing of NADW into the circum-Antarctic deepwater belt occurred a few degrees farther north than today, marked by a steep gradient in glacial $\delta^{13}C$ between 30° and 40°S. Based on these gradients Mackensen *et al.* (2001) suggested formation of Southern Ocean deep water in the zone of extended winter sea ice coverage south of the Polar Front by predominantly brine rejection. This is in accordance with a lowered thermodynamic imprint in the $\delta^{13}C$ signal and low temperatures near the freezing point (Mackensen *et al.* 2001) as well as a strongly increased salinity of glacial deep and bottom waters (Adkins *et al.* 2002). The spreading of this newly formed water mass, however, was restricted to the Atlantic basins south of Walvis Ridge and Rio Grande Rise, where only a small amount of nutrient-enriched deep water passed across these barriers into the northern basins (Mackensen *et al.* 2001, Bickert & Mackensen 2004).

Conversion of the new carbon isotope data set into phosphate concentrations (Fig. 8) applying the approach of Broecker & Maier-Reimer (1992), results in an only slightly increased nutrient inventory of the glacial deep Atlantic compared to present-day values (Bickert & Mackensen 2004). This is in general agreement with previously published glacial Cd/Ca data (Boyle 1992) and conclusions based on south-east Pacific and Southern Ocean data (Matsumoto & Lynch-Stieglitz 1999). This reconciling of the two deep-water nutrient proxies, Cd/Ca and $^{13}C/^{12}C$ ratios, in the Atlantic sector of the Southern Ocean has important implications for the evaluation of the Southern Ocean's role in glacial/interglacial CO_2 cycles. The new Atlantic Southern Ocean data (Mackensen *et al.* 2001, Bickert & Mackensen 2004) are in support of Boyle's (1992) original suggestion of only little change in the global nutrient inventory, hence minimizing the atmospheric pCO_2 response. An increase of nutrients in the deep ocean below 3000 m water depth of more than 50% is needed to lower atmospheric pCO_2 by 80 ppm, and still 30 to 40% nutrient increase is necessary, if a coincident decrease in calcium carbonate export is assumed (Sigman & Boyle 2000). In contrast, if balancing the Atlantic data of Bickert & Mackensen (2004) on a global scale, the calculated increase amounts to only about 10% in the LGM deep ocean (Matthies *et al.* 2004).

Water mass structure reconstructions independent from nutrients, based on $\delta^{18}O$ values of planktonic and benthic foraminifera (Fig. 9), used as an essentially conservative tracer that reflects seawater density, are in general agreement with the inferred circulation and the presence of a glacial NADW (Mackensen *et al.* 2001, Matsumoto *et al.* 2001). Similarly, measurements of the $^{231}Pa/^{230}Th$ ratios in glacial and modern sediments indicate that the export of ^{231}Pa from the Atlantic into the Southern Ocean continued at roughly the modern rate during the LGM (Yu *et al.* 1996). It is inferred from these results that the export of deep water formed in the North Atlantic into the Southern Ocean continued at a comparable rate during the LGM. This is in

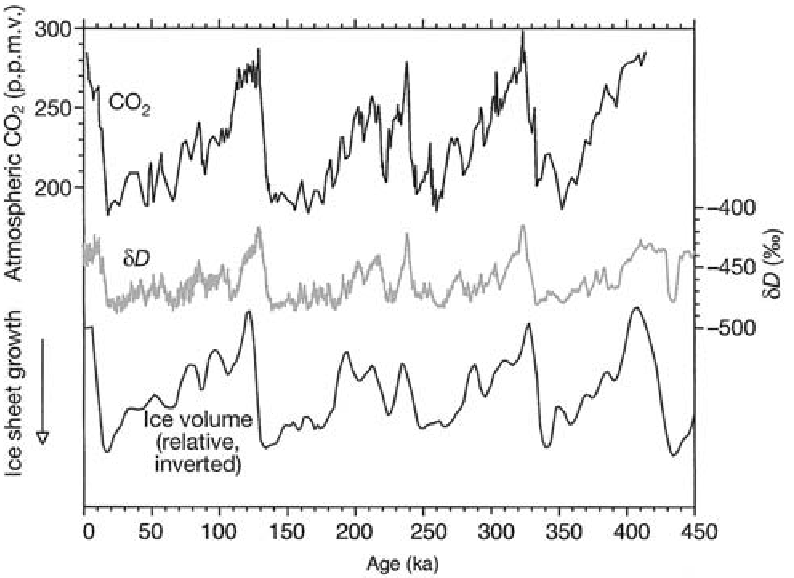

Fig. 10. Atmospheric CO_2 and air temperature variation as recorded by the gas in the ice and by the δ^2H values of the ice, respectively, at Vostok (Petit *et al.* 1999), and global ice volume change as recorded by the $\delta^{18}O$ of deep-sea benthic foraminifera (Bassinot *et al.* 1994) over the last 450 kyr (modified from Sigman & Boyle 2000).

agreement with a study based on the detrital Nd isotopic composition of south-east Atlantic sediments that infers a decreased influence of NADW during glacial periods, which could be due to an increased flow of CDW rather than a major weakening of NADW itself (Bayon *et al.* 2003). Further support for a continuous generation of NADW comes from Oppo & Horowitz (2000) who suggest that the upper Pacific was bathed by a "sub-Antarctic water" which partly received its characteristic nutrient signal by a strong export of deep water out of the Atlantic and its subsequent modification in the Southern Ocean during its long transit via the Antarctic circumpolar circulation.

Glacial/interglacial atmospheric pCO_2 changes

Here focus is on the role of the Southern Ocean circulation in the 100 kyr glacial–interglacial cyclicity of atmospheric pCO_2 during the Late Pleistocene. It is generally accepted that changes in global oceanic circulation patterns and carbon system chemistry have occurred during glacial/interglacial global climate cycles (see above and Schnitker 1974, Streeter & Shackleton 1979, Curry *et al.* 1988, Oppo & Fairbanks 1990, Boyle 1992). In contrast to the reconstruction of Eocene to Miocene/Pliocene atmospheric CO_2 concentrations, which necessarily were based on proxy records, for the last 420 kyr, direct measurements of the gas content in ice cores are available (Barnola *et al.* 1987, Petit *et al.* 1999). These measurements show that the pCO_2 in the atmosphere was approximately 80 ppm lower during glacials and on timescales of 10–100 kyr has varied in phase between the hemispheres and in step with glacial/interglacial cycles of temperature, precipitation, vegetation and ice volume (Fig. 10). These climate variations are due to orbitally driven insolation changes and Earth's radiative balance (Bender 2003, Ruddiman 2003). Radiative forcing from CO_2 may account for half of the glacial/interglacial climate change (Webb *et al.* 1997).

As yet there is no broadly accepted understanding of the processes involved. Many theories for explaining low glacial atmospheric CO_2 concentrations focus either on an increase in the strength of the Southern Ocean's biological pump as a result of changes in the supply or utilization of nutrients or light (e.g. Anderson *et al.* 2002), an increase in the ocean's alkalinity due to coral reef dissolution or carbonate-sediment interactions (for recent summaries see Archer *et al.* 2000, Sigman & Boyle 2000), or on inhibiting gas exchange, between the deep sea and the atmosphere in the Southern Ocean either by increasing Antarctic sea ice (Elderfield & Rickaby 2000, Stephens & Keeling 2000, Keeling & Stephens 2001) or by midwater stratification (Toggweiler 1999). The latter models, however, may be considered as to be variants of a sea surface nutrient drawdown scenario as well (Archer *et al.* 2003).

Models that invoke biological pump changes are generally inconsistent with the magnitude or direction of glacial/interglacial changes in $^{13}C/^{12}C$ ratios and nutrients (Charles & Fairbanks 1990, Boyle 1992), and with the lack of widespread glacial deepwater anoxia (Archer *et al.* 2000). On the other hand, models invoking alkalinity changes predict large increases in the depth of the glacial lysocline, much more extended than observed in sediments (e.g. Bickert & Wefer 1996, Catubig *et al.* 1998). Similarly, models relying on barrier mechanisms to communication between the deep sea and the atmosphere in the Southern Ocean applied to questions of glacial/interglacial CO_2 cycles, suffer from the uncertainty about the high-latitude sensitivity of the real ocean (Archer *et al.* 2003).

Diatom based reconstructions of the LGM sea ice distribution in the Southern Ocean revealed an extension of the winter sea ice distribution as far north as the Polar Front

(Crosta *et al.* 1998, Gersonde *et al.* 2003). Stephens & Keeling (2000) and Elderfield & Rickaby (2000) proposed a mechanism, that is based on changes in the rate of air-sea gas exchange as a result of increased winter sea ice cover at high southern latitudes. By significantly limiting the sea-to-air CO_2 flux in the primary region for deepwater ventilation, expanded Antarctic sea ice during glacial times may trap relatively more carbon in the deep ocean, thereby reducing atmospheric CO_2 concentrations. However, the sea ice-driven regulation of CO_2 outgassing is effective only if the wintertime sea ice coverage south of the Antarctic Polar Front rises to 99–100%. At present, strong circumpolar winds pull the ice apart, and intense oceanic heat fluxes from deep vertical mixing hinder ice growth to a point that sea ice covers no more than 80–90% of the total winter pack ice area. Morales Maqueda & Rahmstorf (2002) presented simulations with a coupled sea ice/upper ocean model indicating that the CO_2 sequestration under glacial ice cover could account for at most 15–50% of the total glacial CO_2 decline. However, there is increasing evidence from proxies that glacial deep and bottom water masses in the Antarctic have been close to the potential freezing point (Mackensen *et al.* 2001, Adkins *et al.* 2002, Schrag *et al.* 2002), which might have allowed for an almost complete winter sea ice cover south of the Polar Front. Also, low $\delta^{13}C$ values of Antarctic benthic foraminifera are partly (by ≈ -0.4‰) caused by a reduced thermodynamic imprint (Mackensen *et al.* 2001), which is in accordance with low air-sea gas exchange ratios at the site of deep water formation and the sea ice formation scenario.

Sigman & Boyle (2000) presented a hypothesis that focuses on both the biology and physics of the open ocean surrounding Antarctica. In this hypothesis, a cooler climate caused a northward shift in the maximum westerly winds that drive upwelling and northward surface flow in the modern Antarctic. This shift caused a decrease in the upwelling of deep water into the Antarctic surface, replacing it with upwelling of intermediate-depth water into the Subantarctic surface. Subsequently, a stable, fresh, frequently ice-covered surface layer developed, further reducing the deep ocean ventilation in the open Antarctic. The increased stratification that resulted from these changes lowered the rate of nutrient supply to the Antarctic surface and reduced CO_2 outgassing. On the other hand, the hypothetical Subantarctic was more productive due to a combination of higher export production and a reduction in nutrients from the Antarctic surface. By increased drawdown of CO_2 in the ocean interior, this change could explain lower glacial atmospheric CO_2 levels.

This hypothesis is supported by nitrogen isotope data which suggest that during the last ice age nitrate utilization was twice or more its current value (Francois *et al.* 1997). A preliminary box model calculation predicts that considerably higher nitrate utilization (that is, 50–65% during the last ice age compared to 25% during the present interglacial) could lower atmospheric CO_2 by the full glacial–interglacial amplitude (Sigman *et al.* 1999). In contrast, Elderfield & Rickaby (2000) infer from Cd/P ratios in planktonic foraminifera that phosphate utilization during the LGM was much smaller in the Antarctic, but similar to Holocene values in the sub-Antarctic Ocean. However, taking those palaeoceanographic proxy data at face value that suggest that during the LGM Antarctic export production south of the Polar Front was lower but north of it in the sub-Antarctic higher (cf. Bopp *et al.* 2003), Sigman & Boyle (2000) inferred that more complete nitrate utilization in the Antarctic was due to a lower rate of nitrate supply from the subsurface, implying that the fundamental driver of the CO_2 change was a glacial decrease in the ventilation of deep waters at the surface in the Antarctic.

Moore *et al.* (2000) came up with a hybrid hypothesis that invokes, during glacial times south of the Antarctic Polar Front, intense surface stratification and nutrient consumption during the summer, followed by prevention of gas exchange by ice cover during winter. North of the Polar Front altered phytoplankton species composition and increased export production resulted in a more efficient atmospheric CO_2 drawdown. These processes, averaged over the Southern Ocean as a whole and encompassing the area from the Subtropical Front to the Antarctic continent, resulted in an increased net flux of CO_2 into the ocean during glacial times. Increased primary production during glacials in this scenario is driven by elevated fluxes of iron from continental dust sources to surface waters (Martin *et al.* 1990). To make this scenario consistent with the known sediment record that *per se* does not confirm a net increase of Southern Ocean palaeoproductivity, Moore *et al.* (2000) call upon the effect of iron limitation on uptake ratios of macronutrients by diatoms (Takeda 1998). Under iron limitation during interglacial periods phytoplankton silica/carbon ratios are 2–3 times higher than under supposed iron fertilization during glacials. This means that glacial organic carbon export calculated from biogenic opal accumulation rates may be severely underestimated (Martin *et al.* 1990). In addition, it is argued that *Phaeocystis* spp., a common phytoplankton group in Antarctic coastal waters, may have increased their abundance and range at the LGM (Martin *et al.* 1990, Moore *et al.* 2000). This is important because these phytoplankton contain no hard parts and would leave little signal in the sediments, but sinking *Phaeocystis* colonies can be efficient exporters of carbon. However, a recent model derived estimate of productivity changes and its effect on atmospheric pCO_2 at the LGM using high dust deposition rates from the LGM as input shows a small increase in the relative abundance of diatoms in today's iron-limited regions, causing a global increase in export production by only 6% and an atmospheric CO_2 drawdown of just 15 ppm (Bopp *et al.* 2003).

Recently, Shemesh *et al.* (2002) reconstructed a sequence of events during the last deglaciation from a sediment core south of the present day Polar Front indicating that sea ice and nutrient changes lead the increase in atmospheric pCO_2 by approximately 2 kyr. These authors assume that their data from the Atlantic sector are representative for the circumpolar Ocean south of the Polar Front. Hence, they conclude that the delayed CO_2 response to sea ice and nutrient cycling implies that the latter cannot be responsible for the observed atmospheric pCO_2 variations. If this is correct, the pattern of glacial to interglacial pCO_2 variations cannot be dominantly controlled by Southern Ocean circulation dynamics but must be triggered elsewhere.

Although the above review of recent, in part controversial Southern Ocean hypotheses for explaining the observed glacial CO_2 drawdown is far from complete, in summary appears that we are not yet able to decide whether and at which times the Southern Ocean played a dominant role.

Rapid climate changes

To address further the Southern Ocean's role in climate variability, the relative timing of millennial climatic changes between and within Hemispheres is important. Millennial climate variability was first noticed in Greenland ice cores (Dansgaard *et al.* 1993) and the rapid changes are now known as Dansgaard-Oeschger (D/O) events. Although this paper aims to focus on the major changes in Southern Ocean circulation during the last 34 Ma, I here need to mention that changes in the North Atlantic thermohaline circulation are the most commonly cited forcing or feedback mechanisms for climate variability on centennial and millennial time scales. Three main modes of the Atlantic thermohaline circulation have been identified from palaeoceanographic data and in numerical palaeomodels:

i) a warm, interstadial mode with strong convection in the Nordic Seas,

ii) a cold, stadial mode with weaker convection occurring south of the Greenland–Scotland Ridge, and

iii) an "off" mode with virtually no deep water formation in the North Atlantic and with waters of Antarctic origin filling the deep Atlantic basins (e.g. Sarnthein *et al.* 1994, Rahmstorf 1996, Seidov *et al.* 1996).

However, the trigger mechanism for the past mode switches remains unknown. Millennial circulation anomalies may be forced in the Atlantic itself, by some coupled ocean-atmosphere instability. Alternatively they may be controlled remotely from the tropics, by shifts in the atmospheric planetary-wave pattern (for recent discussion see Cane 1998, Stocker 1998, 2003, Seidov & Maslin 2001, Rahmstorf 2002). The role of the Southern Ocean in modulating these processes has received less attention, although increasing evidence shows that millennial to submillenial variability is a global feature of the climate (Clark *et al.* 1999, Seidov *et al.* 2001, Voelker 2002) and has interhemispheric significance (Ninnemann *et al.* 1999, Kanfoush *et al.* 2000, Knorr & Lohmann 2003).

There is contrasting evidence from various palaeoclimate archives that indicate both interhemispheric synchrony and asynchrony (e.g. Charles *et al.* 1996, Bard *et al.* 1997, Blunier & Brook 2001, Sarnthein *et al.* 2002, Pahnke *et al.* 2003). Blunier & Brook (2001) for example, documented that Antarctic temperature changes precede northern Hemisphere ice surges by a few millennia, whereas Pahnke *et al.* (2003) suggested direct climate linking and synchrony between the hemispheres, based on the similarity of centennial-scale records of planktonic foraminiferal $\delta^{18}O$ and Mg/Ca from the southern mid-latitudes with D/O cycles in Greenland ice cores. Kanfoush *et al.* (2000) analysed the IRD content of two cores in the eastern South Atlantic situated well south and north of the Polar Front, to be not affected by meridional shifts of the frontal system during past climatic changes. Distinct IRD events in both cores are more or less coeval across 12° latitude between 30 and 60 kyr and correlate with an increase in benthic foraminiferal $\delta^{13}C$ suggesting that South Atlantic IRD events are associated with strong NADW production and inferred warming in the North Atlantic during the strongest interstadials in Greenland (cf. Charles *et al.* 1996). The timing of IRD events, coinciding in the South Atlantic with the strongest warm phases of D/O cycles and in the North Atlantic with cold phases (Heinrich events), may be a manifestation of antiphase climate behaviour between these regions (Kanfoush *et al.* 2000). However, a robust assessment of interhemispheric temporal pattern of millennial and submillennial climate change is difficult due to the uncertainties related to radiocarbon dating and the associated problems of variations in local reservoir ages as well as errors attended to the correlation of different age/depth models.

Summary and conclusions

Within the given frame of this paper only a brief overview of the most apparent times of palaeoceanographic changes in the Southern Ocean could be given. This necessarily remains incomplete and as such is biased by the author's personal view and limited experience. In the following I nevertheless will summarize the most important conclusions:

> The initiation of a circum-Antarctic circulation at the Eocene/Oligocene boundary coincided with the first continent-wide glaciation on East Antarctica, but the role of declining atmospheric CO_2 in both, the initial build-up as well as the subsequent decaying and growth of the ice sheet during Oligocene and early Miocene times is not yet constrained. Furthermore, strong evidence exists that there was no significant deep ocean cooling associated with the continental ice

build-up at the E/O boundary.

In the Middle Miocene the tectonic changes in the Tethys Ocean in the present day eastern Mediterranean, may have led to a cooling of Antarctic surface waters and in conjunction with an increased formation of Southern Ocean bottom waters fostered a rapid growth of continental ice sheets, and ultimately resulted in a re-establishment of the East Antarctic Ice Sheet and the first build-up of a significant West Antarctic Ice Sheet. A slight cooling of the deep ocean of about 1°C is documented, but there is no evidence for a distinct atmospheric CO_2 decrease with the ice sheet growth or for a large permanent drop in $p\mathrm{CO}_2$ during the middle Miocene.

During the Last Glacial Maximum, NADW was partially replaced by nutrient-rich bottom waters of southern origin, but not as extensively as previously suggested. Rather it seems that below a nutrient depleted water mass of probably partly Mediterranean origin, a layer of glacial NADW flowed southward at about the same depth interval as today, just slightly more depleted and less extended in the water column. This might point to an increased flow of glacial CDW rather than a major weakening of NADW itself. The reconstructed nutrient distribution based on this deep water mass geometry would further suggest a comparatively small change of 10% in the oceans nutrient inventory and thus could not explain the lower atmospheric CO_2 concentration by 80 ppm as observed in ice cores.

There is contrasting evidence for a dominant control of glacial to interglacial $p\mathrm{CO}_2$ variations by Southern Ocean circulation dynamics. Within those hypotheses that are in favour of a significant role of the Southern Ocean, two groups of widely spread hypotheses invoke either changes in the strength of the Southern Ocean's export productivity as a result of varying supply and utilization of nutrients, or barriers in the gas exchange between the deep sea and the atmosphere by sea ice or mid-depth water stratification. However, uncertainties associated with the proxies for surface ocean nutrient status and export production do not yet allow us to decide whether the Southern Ocean played a dominant role or whether the pattern of glacial to interglacial $p\mathrm{CO}_2$ variability was triggered elsewhere.

To further elucidate the linking between atmospheric CO_2 and ocean circulation changes, a substantial portion of funded research nowadays is dedicated to the relative timing of millennial and submillennial climatic changes between and within Hemispheres. Changes in the North Atlantic thermohaline circulation are most commonly called for as forcing or feedback mechanisms for so-called rapid climate changes. However, increasing evidence shows that millennial to centennial variability is a global feature of the climate. The role of the Southern Ocean in triggering or modulating these processes has only recently received some attention, but contrasting evidence from various palaeoclimate archives indicate both interhemispheric synchrony and asynchrony. Synchrony between the Hemispheres call for a direct climate linking via the atmosphere or a forcing from the low latitudes, whereas an antiphase climate behaviour can suggest a trigger mechanism effective on either Hemisphere, transmitted and modulated by the thermohaline circulation. However, a robust assessment of interhemispheric temporal pattern of millennial and submillennial climate change is difficult due to the uncertainties related to dating and errors attended to the correlation of different climate archives.

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