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Review

A review of Tertiary climate changes in southern South America and the Antarctic Peninsula. Part 1: Oceanic conditions

J.P. Le Roux

Departamento de Geología, Facultad de Ciencias Físicas y Matemáticas, Universidad de Chile/Centro de Excelencia en Geotérmia de los Andes, Casilla 13518, Correo 21, Santiago, Chile

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ABSTRACT

Oceanic conditions around southern South America and the Antarctic Peninsula have a major influence on climate patterns in these subcontinents. During the Tertiary, changes in ocean water temperatures and currents also strongly affected the continental climates and seem to have been controlled in turn by global tectonic events and sea-level changes. During periods of accelerated sea-floor spreading, an increase in the mid-ocean ridge volumes and the outpouring of basaltic lavas caused a rise in sea-level and mean ocean temperature, accompanied by the large-scale release of CO_2. The precursor of the South Equatorial Current would have crossed the East Pacific Rise twice before reaching the coast of southern South America, thus heating up considerably during periods of ridge activity. The absence of the Antarctic Circumpolar Current before the opening of the Drake Passage suggests that the current flowing north along the present western seaboard of southern South American could have been temperate even during periods of ridge inactivity, which might explain the generally warm temperatures recorded in the Southeast Pacific from the early Oligocene to middle Miocene. Along the east coast of southern South America, water temperatures also fluctuated between temperate-cool and warm until the early Miocene, when the first incursion of temperate-cold to cold Antarctic waters is recorded. The cold Falkland/Malvinas Current initiated only after the middle Miocene. After the opening of the Drake Passage, the South Equatorial Current would have joined the newly developed, cold Antarctic Circumpolar Current on its way to Southern South America. During periods of increased sea-floor spreading, it would have contributed heat to the Antarctic Circumpolar Current that caused a poleward shift in climatic belts. However, periods of decreased sea-floor spreading would have been accompanied by diminishing ridge volumes and older, cooler and denser oceanic plates, causing global sea-level falls. This would have resulted in a narrowing of the Drake Passage, an intensification of the Antarctic Circumpolar Current that enhanced the isolation of Antarctica from warmer northern waters, and increased glaciation on the Antarctic Peninsula. Colder ocean surface waters would also have trapped more CO_2, enhancing climate cooling on the adjacent continents. During these periods the atmospheric belts shifted equatorward and increased the latitudinal thermal gradient, leading to higher wind velocities and enhanced oceanic upwelling along the western seaboard of Southern South America.

1. Introduction

Global warming has been a topic on the lips of prominent politicians and countless scientists over the past decade, so much so that the general public has become acutely aware of the climate problems facing the world. At the same time, there has been an intense debate concerning the causes of recent climate change, whether they are driven by human interference or can be related to geological processes that have been active over millions of years.

The present manuscript and its companion paper (Le Roux, 2012) focus on climate changes in southern South America (SSA) and the Antarctic Peninsula (AP) during the Tertiary, thus encompassing a time span of 62.9 Ma before any possible anthropological influence. Because changing conditions on these subcontinents and in their surrounding oceans also play an important role in present world climates, one of the objectives of this review is to provide basic information on the duration and possible causes of long-term climatic cycles, upon which present, anthropologically induced climate changes are superimposed. In Part 1, changes in the oceans that surrounded or invaded SSA and the AP during the Tertiary are discussed. These constitute a key factor in understanding what happened on the adjacent continents, which is the topic of Part 2.

Virtually all methods used to study paleoclimates are based on the principle of uniformitarianism: *The present is the key to the past* (Hutton, 1788). This review therefore starts with a brief summary of the main factors controlling present climate patterns in SSA and the AP. These include atmospheric systems, oceanic conditions, and tectonic evolution, all of which are intimately associated. The opening of the Drake Passage between SSA and the AP, for example, allowed

E-mail address: jpleroux@terra.cl.

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the initiation of the Antarctic Circumpolar Current, which isolated Antarctica from the influx of warm currents from the north and intensified its glaciation. This in turn caused an equatorward shift in atmospheric systems that affected ocean currents and climate patterns on the adjacent continental areas.

2. Factors that control climate patterns in southern South America and the Antarctic Peninsula

2.1. Atmospheric systems

Atmospheric systems affecting SSA and the AP (Fig. 1) are the Intertropical Convergence Zone (ITCZ), South Pacific Anticyclone, South Atlantic Anticyclone, South-westerly Wind Belt, and the Circum-Antarctic Low Pressure Belt.

The ITCZ is a continuous zone of relatively low and uniform pressure around the equator, in which air masses rise and produce tropical thunderstorms throughout the year. This belt migrates about 1000 km back and forth over the equator following the zenith point of the sun, giving rise to wet summers and dry winters in north-easternmost Chile and southern Bolivia. In particular, the presence of a large low-pressure cell located over northern Argentina, Paraguay and south-eastern Bolivia during the austral summer, may cause severe floods. The areas north and south of the ITCZ are subjected to the trade winds as air is drawn into the low pressure system, but within the zone itself winds are calm.

The South Pacific Anticyclone and South Atlantic Anticyclone form part of a broad subtropical belt of descending air masses causing high pressure and generally dry conditions around the 33°S latitude. The South Pacific Anticyclone is generally located off the west coast of central Chile and causes anticlockwise, northward air flow along the continent, which in turn generates offshore-flowing surface currents and landward oceanic upwelling. During El Niño/Southern Oscillation (ENSO) events, when the atmospheric systems shift further to the north than usual, upwelling is weakened, severely affecting the fishing industry. In summer, the South Pacific Anticyclone prevents the South-westerly Wind Belt from bringing rain to central Chile, but in winter the whole South Pacific Anticyclone-South-westerly Wind Belt system shifts equatorward and allows the northward advance of cold fronts. In Argentina, the South Pacific Anticyclone moves across Patagonia in late autumn and winter, bringing cold and dry air masses from the south known as the Sudestada. Where this system collides with warm, humid air masses from the northeast, heavy rain occurs over eastern and north-eastern Argentina.

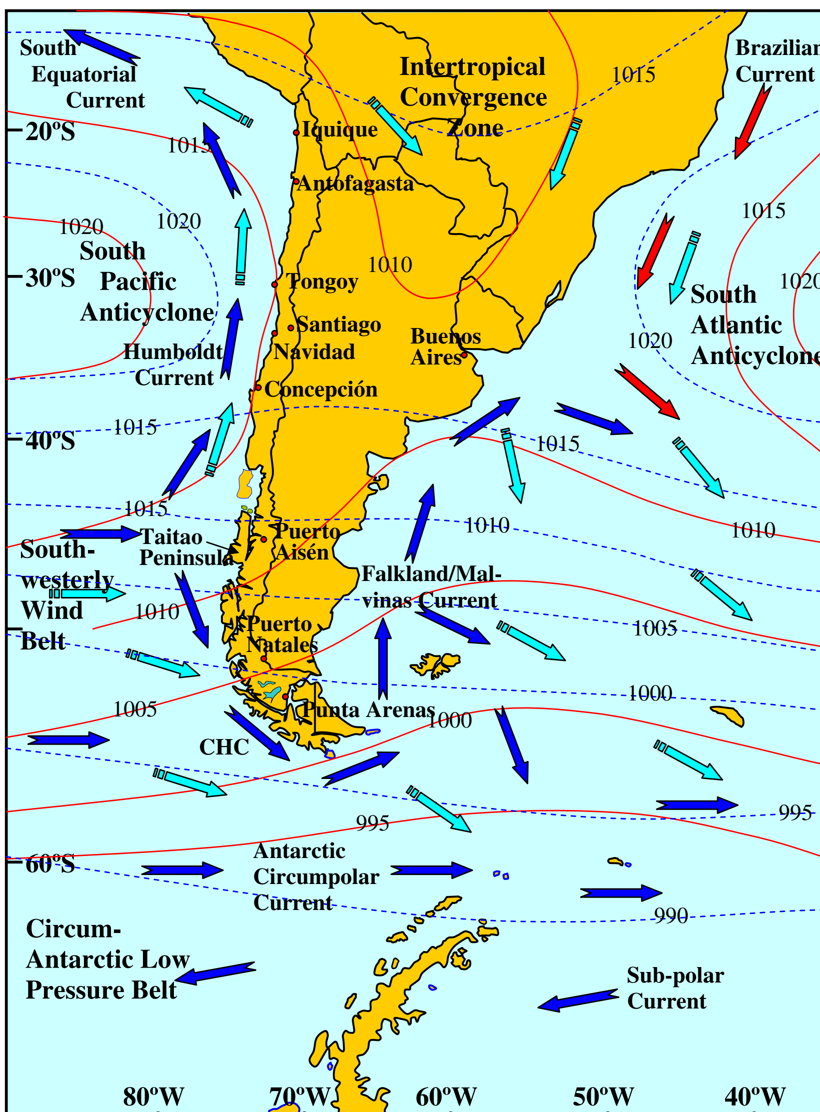

Fig. 1. Atmospheric systems and ocean currents that presently affect SSA and the AP. Dark blue arrows: cold ocean currents; red arrows: warm ocean currents; light blue arrows: atmospheric circulation; red lines: mean atmospheric pressure during summer; blue stippled lines: mean atmospheric pressure during winter (mb).

The South Atlantic Anticyclone is located in the Atlantic Ocean off southern Brazil, introducing warm and humid winds to central and north-eastern Argentina from the north and north-east (García, 1991). During the austral winter it moves closer onshore. The meridional movements of these air masses give rise to frontal rains, heavy precipitation occurring during El Niño events and extreme temperatures and droughts prevailing with the advent of blocking anticyclones (Iriondo and García, 1993).

The South-westerly Wind Belt impacts the Chilean coast most strongly at about 48°S, bringing high rainfall to this region throughout the year. Due to the orographic effect of the Patagonian Andes, precipitation increases from 7000 mm at the coast to 10,000 mm over the Southern Patagonian Ice Fields (Fig. 2), but rapidly decreases to less than 400 mm east of the latter (Carrasco et al., 2002). The cold, dry wind resulting from this system in Patagonia and central Argentina is known as the Pampero.

The location of the Circum-Antarctic Low Pressure Belt surrounding Antarctica (Fig. 1) also affects precipitation patterns in SSA. If it is located toward the Australian sector, Patagonia is relatively dry, but when it shifts toward SSA, the high latitudes receive more precipitation (Pittock, 1980).

A factor which presently has a marked effect not only on the SSA climate but also worldwide, is the El Niño/ENSO (Enfield and Mayer, 1997; Grove, 1998). El Niño is caused by the spreading of warm water from the western Pacific and Indian Ocean to the eastern equatorial Pacific, which is accompanied by high air surface pressure in the tropical western Pacific. This phenomenon is accompanied by the weakening of trade winds and oceanic upwelling, unusually high winter precipitation in central Chile and on the Bolivian Altiplano, and abnormal rainfall in northern Argentina and southern Brazil during spring and early summer. ENSO's cool phase, known as La Niña, is characterized by the development of a cold pool in the eastern Pacific, low surface air pressures in the western Pacific, and a strengthening of trade winds. Droughts or late winter rains occur in central Chile. The mechanisms that cause this oscillation remain unknown.

2.2. Ocean currents

SSA and the AP are surrounded by three oceans, namely the Southeast Pacific in the west, the Southwest Atlantic in the east, and the Southern Ocean in the south (Fig. 3). The Southern Ocean proper includes the oceanic region between Antarctica and the Subtropical Convergence Zone at approximately 40–55°S (Foldvik and Gammelsrød, 1988). However, for practical purposes it refers here specifically to the area south of Cape Horn and the Scotia Ridge, thus including the Drake Passage and Weddell Sea (Fig. 3). Some data from the Ross Sea, which lies west of the AP, are also included.

In the Southeast Pacific, the South Equatorial Current flows west along the equator, swinging southward east of New Guinea and then turning east at about 50°S, where it joins the Antarctic Circumpolar Current or West Wind Drift (Fig. 1). As it approaches the Chilean coast between 40 and 45°S, this combined current divides into the cold, north-flowing Humboldt or Peru-Chile Current (HC) and the south-flowing Cape Horn Current (Boltovskoy, 1976). The HC

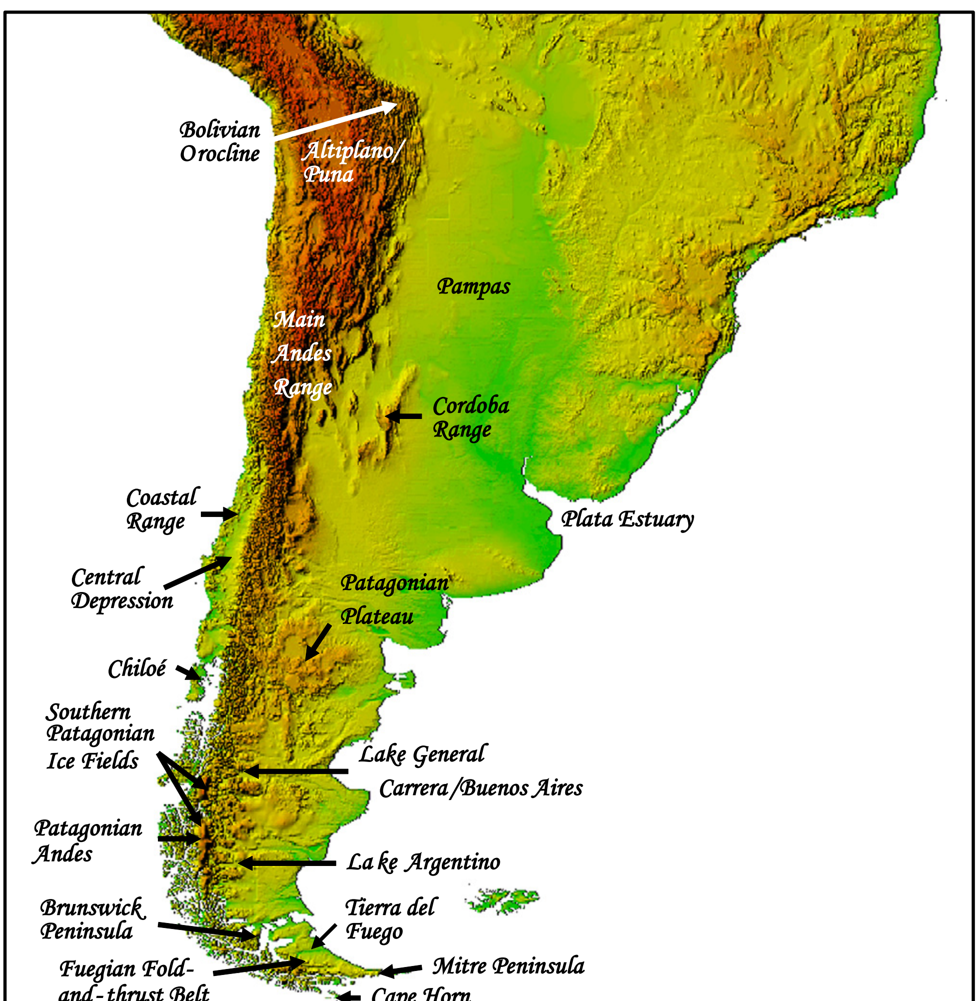

Fig. 2. Topography of southern South America, with some localities mentioned in text.

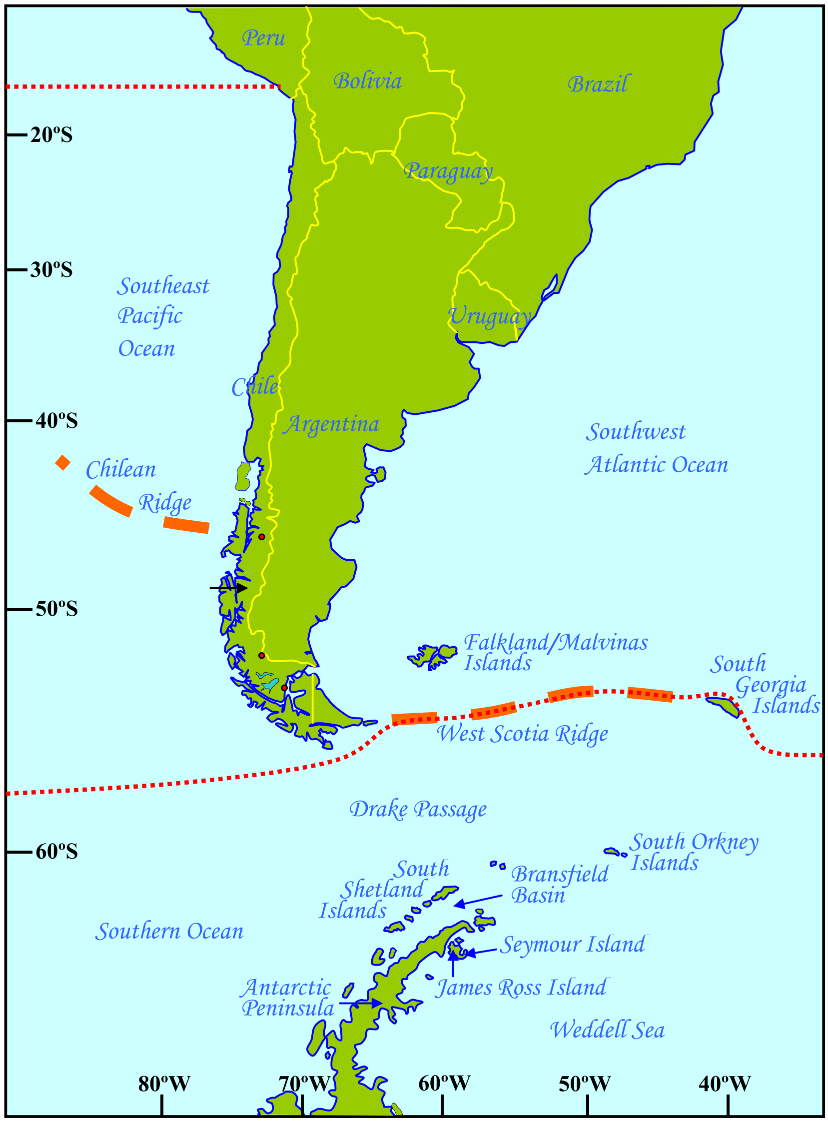

Fig. 3. Oceans and islands around southern South America and the Antarctic Peninsula.

eventually deviates to the west close to the equator, where it joins the South Equatorial Current. Occasionally, it splits into two branches, the Humboldt Coastal Current and Humbolt Oceanic Current, separated by the south-flowing Peru-Chile Countercurrent, which has warm, subtropical waters. The Humboldt Coastal Current extends for about 100 km offshore, whereas the Peru-Chile Countercurrent lies between 100 and 300 km from the coast.

Although the HC may have been active since the early Tertiary (Keller et al., 1997), it probably intensified due to a major expansion of the Antarctic Ice Sheet between 15 and 12.5 Ma (Flower and Kennett, 1993a) and especially after closure of the Central American Seaway between 4.0 and 3.0 Ma (Orgeira, 1990; Ibaraki, 1997).

The Circum-Antarctic Low Pressure Belt produces a westward-flowing coastal current, the Sub-polar Current or East Wind Drift (Fig. 1). North of the Circum-Antarctic Low Pressure Belt, the Antarctic Circumpolar Current flows clockwise around Antarctica and through the narrow Drake Passage. It keeps warm waters away from Antarctica and plays an important role in maintaining the ice cap on this continent. The Antarctic Circumpolar Current is split by the Scotia Ridge, with a shallow, warm branch flowing to the north and a deeper, cold branch passing farther to the east before also turning north. This cold Falkland/Malvinas Current flows along the Atlantic coast up to the Plata Estuary (Fig. 2).

The north-eastern part of Argentina, Uruguay and southern Brazil are affected by the south-flowing, subtropical Brazilian Current, which deviates to the east opposite the Plata Estuary. This current is the southward diversion of a portion of the South Atlantic Equatorial Current where the latter meets the South American continent.

2.3. Tectonic evolution and topography

SSA is climatologically affected by two major topographic features, namely the Andes Range, which runs along the whole length of South America and reaches elevations close to 7000 m above sea level (a.s.l.) in west-central Argentina, and the Coastal Range in central and south-central Chile (Fig. 2). The Andes Range forms a major rain-shadow, with a drastic decrease in rainfall from several thousand mm on its western flank to a few hundred mm on its eastern flank. The Coastal Range, which is separated from the Andes by the Central Depression, seldom exceeds 2000 m a.s.l., but nevertheless has its own rain-shadow effect (Moreno, 2001).

The Andean Tectonic Cycle is divided into three stages (Charrier et al., 2007), of which only the most relevant aspects are discussed here. During the First Stage, subsidence in the Andean foreland basin of the Altiplano region increased at around 89 Ma, when the tectonic regime first became compressional, and again at 73 and 58 Ma (Sempere et al., 1997; Gregory-Wodzicki, 2000). Resumed subduction during the Paleogene formed the Second Stage, which was characterized by a high plate convergence rate of more than 10 cm/yr between about 50 and 40 Ma (Pardo-Casas and Molnar, 1987). This stage coincided with a reduction in the convergence obliquity at around 45 Ma (Pilger, 1983, 1984; Pardo-Casas and Molnar, 1987) and a major deformational event controlled by the inversion of faults at about 40 Ma (Pineda and Emparan, 2006). The latter was also recorded by Hongn et al. (2007) along the eastern border of the Puna in northwest Argentina. At this stage the landscape north of 30°S was dominated by the Incaic Range, which formed by inversion and uplift of the former arc and its associated intra-arc basins (Charrier et al., 2007). Between the Incaic Range and the Coastal Range, a narrow valley (the proto-Central Depression) developed, which became the Central Depression during the late Eocene–early Oligocene when the Incaic Range was eroded. The Third Stage ranged from the late Paleogene to Recent, during which uplift established the present configuration of the Andean Range. Since the late Eocene, major extensional tectonics were recorded in central Chile, starting some time before 36 Ma and lasting until about 26 Ma (Charrier et al., 2002). This coincided with a low plate convergence rate (5–10 cm/yr) during most of the Oligocene, which was followed by a high rate of convergence (>10 cm/yr) during the early Miocene up to about 15 Ma (Pardo-Casas and Molnar, 1987). The latter was accompanied by tectonic inversion and erosion between 21 and 16 Ma in central Chile (Charrier et al., 2002) and expansion of volcanism toward the retroarc domain between 19 and 15 Ma (Ramos et al., 2002). The plate convergence rate decreased to about 6 cm/yr after 12 Ma (Le Roux et al., 2005a), but thereafter increased again to 10 cm/yr from about 8 to 7 Ma (Cande and Leslie, 1986; Le Roux et al., 2005a). Gregory-Wodzicki (2000) proposed that the Altiplano–Puna region experienced major uplift of 2300–3400 m after 10.7 Ma.

The Patagonian Andes Range developed because of a collision between the Chilean Ridge (Fig. 3) and SSA that coincided with the Quechua Phase of Andean tectonics at 18 Ma, reaching its maximum uplift and deformation at 9 Ma (Malumián and Ramos, 1984). An important rain-shadow effect can be clearly seen in the increasing $\delta^{13}C$ and decreasing $\delta^{18}O$ values in the Santa Cruz Formation of Patagonia after 16.5 Ma (Blisniuk et al., 2005), which is supported by a change in savanna-woodland fauna (Santacrucian Land Mammal Age) to pampa fauna (Colloncuran Land Mammal Age) at 15 Ma in the Río Frías Formation of the Golfo de San Jorge Basin (Fig. 4) (Marshall and Salinas, 1990).

In Tierra del Fuego, Olivero et al. (2003) identified two pulses of Andean uplift during the Tertiary on the basis of erosional unconformities, namely in the middle-late Paleocene and early-middle Eocene. The last uplift was associated with the deposition of conglomerates of the Tres Amigos Formation in alluvial fan systems. A two-phase deformation was also proposed by Ghiglione and Cristallini (2007) on the basis of analog sandbox experiments. These authors concluded that the Patagonian Orocline must have formed from two successive orthogonal indentations to match the observed structural features: first northward to form the Fuegian Fold-and-Thrust Belt and then eastward to propagate thrusting in the Patagonian Andes. This is

Fig. 4. Approximate distribution of Tertiary sedimentary basins with marine deposits in southern South America. Modified from Uliana and Biddle (1988), Aceñolaza (2004), Malumián and Nañez (2011), and other sources mentioned in text. The basin margins represent their maximum expansions during the indicated time intervals, not the basin sizes and shapes during any specific period.

consistent with a change in the convergence direction of the Farallon–Nazca Plate at 27 Ma. An important conclusion is that the orocline curvature existed at least since the Late Cretaceous and that oroclinal rotation was not significant in relation to Tertiary orogenic shortening. Paleomagnetic studies in the AP (Poblete et al., 2011) also indicate that it did not undergo significant rotation over the last 100 Ma.

The timing of the opening of the Drake Passage between the AP and SSA and the development of the Antarctic Circumpolar Current has been a matter of debate. Until the Late Jurassic, SSA and the AP were still locked together, but these two fragments of Gondwanaland began to move apart with the formation of the Scotia Plate. Ghiglione et al. (2008) presented evidence that an extensional basin began to develop in Tierra del Fuego during the latest Paleocene–early Eocene. This extension ended at 49 Ma, concurrent with an eightfold increase in the separation rate between SSA and the AP and the development of the Drake Passage (Livermore et al., 2005; Eagles et al., 2006). However, they contended that this initial separation did not yet allow the exchange of water between the Pacific and Atlantic Oceans, only implying continental lithospheric stretching preceding the development of embryonic basins in the West Scotia Sea. Neodinium isotope ratios from the Agulhas Ridge suggest that the influx of Pacific seawater into the Atlantic Ocean commenced at 41 Ma (Scher and Martin, 2006), considered by Lyle et al. (2007) to reflect a shallow ocean connection between the two oceans at the time. The opening of the Drake Passage has also been linked to the development of the East Antarctic Ice Sheet (Shackleton and Kennett, 1975; Barker and Burell, 1977), which commenced in the late Eocene (Ehrmann and Mackensen, 1992; Zachos et al., 2001). This might have coincided with the onset of spreading between 34 and 30 Ma in the West Scotia Sea (Livermore et al., 2005) and the end of sedimentation in the southernmost part of the Magellanes/Austral Basin (Sloggett Sub-basin; Fig. 4) of Tierra del Fuego (Olivero et al., 1998). Kennett et al. (1975) maintained that the Antarctic Circumpolar Current had already been established by the late Eocene, based on the presence of numerous hiatuses in lowermost Oligocene marine successions in the Antarctic region. These would have resulted from intensification in the ocean circulation brought about by the establishment of a deepwater current. Other authors, however, considered that the Antarctic Circumpolar Current developed later. According to Barker (2001), spreading of the West Scotia Ridge started in the late Oligocene (28 Ma), accompanied by widening of the Drake Passage and the opening of the Bransfield Basin (Fig. 3), which maintained a spreading rate of 1.1 mm/yr during the Oligocene–Miocene (Sell et al., 2004). Lyle et al. (2007), examining sediment core of late Oligocene age in the South Pacific, identified bottom scouring events that they associated with the first deep circulation of the Antarctic Circumpolar Current at 25–23 Ma. This is supported by grain-size evidence of stronger flow through the Tasman Gateway (Pfuhl and McCave, 2005) and the

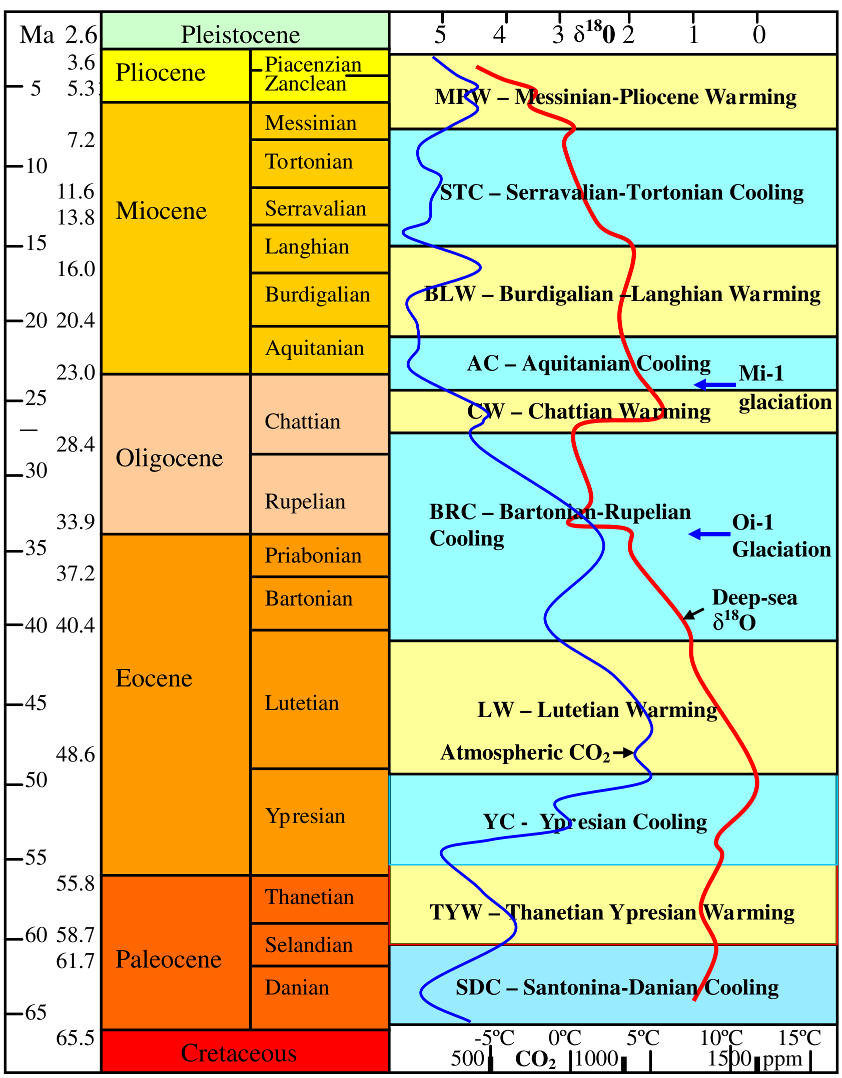

Fig. 5. Long-term global trends in ocean and air temperatures as shown by δ^{18}O and CO_2 records (after Zachos et al., 2001; Beerling and Royer, 2011). An increase in temperature is shown by a decrease in δ^{18}O and an increase in CO_2.

development of bottom scouring currents in the southern Indian Ocean (Shipboard Scientific Party, 1989) at this time. Finally, Barreda and Caccavari (1992) maintained that sufficient deepening of the Drake Passage to allow the establishment of the deepwater Antarctic Circumpolar Current occurred between 14 and 11 Ma, which caused cooling of the Antarctic landmass to its present condition. However, considering the bulk of the evidence presented above, it seems likely that the Drake Passage developed and allowed Pacific water to enter the South Atlantic at around 41 Ma, but that the deepwater Antarctic Circumpolar Current only became well established by about 24 Ma. This was followed shortly thereafter by the Mi-1 glaciation and subsequent long-term cooling leading up to the Pleistocene (Zachos et al., 2001; Fig. 5).

3. Oceanic changes during the Tertiary

The three oceans surrounding SSA and the AP not only had a considerable influence on climate changes in the region, but also at times invaded the South American subcontinent, possibly linking the Pacific and Atlantic Oceans. This temporary ocean is here referred to as the Transcontinental Sea to avoid confusion with local names (e.g., the "Paranense", "Patagoniense", and "Entrerriense" Seas, referring to distinct transgressions occurring at different times in different regions). In the description below, marine deposits along the present Atlantic seaboard of SSA and Tierra del Fuego, as well as the Falkland/Malvinas Islands are generally assigned to the Southwest Atlantic Ocean, whereas deposits occurring within the SSA interior and along the west coast of Tierra del Fuego are discussed under the heading Transcontinental Sea.

3.1. Southeast Pacific Ocean

There are very few studied marine Paleocene or Eocene deposits along the Pacific seaboard of Chile, so that oceanic conditions during this period remain unknown. However, widespread evidence exists that warm sea-water conditions prevailed along the Chilean coastline from at least the early Oligocene to middle Miocene. Some shallow, warm-water mollusks in well-dated, latest Oligocene to middle Miocene successions in southern Peru can be correlated with species in the Navidad Formation of central Chile (DeVries and Frassinetti, 2003; Nielsen et al., 2003; Finger et al., 2007; Encinas et al., 2008). This formation contains gastropod fauna that include many exclusively or predominantly warm-water genera such as *Nerita*, *Strombus*, *Sinum*, *Distorsio*, *Ficus*, *Zonaria*, *Olivancillaria*, *Terebra*, *Architectonica*, and *Heliacus* (Covacevich and Frassinetti, 1980, 1983, 1986; Frassinetti and Covacevich, 1981; Groves and Nielsen, 2003; Nielsen et al., 2004). The age of the host strata is shown by extensive ^{87}Sr/^{86}Sr dating on the macrofossil shells to range between 31.5 ± 0.6 Ma and 11.5 ± 1.0 Ma (early Oligocene–earliest late Miocene), as confirmed by the K–Ar and Ar–Ar ages of volcanic scoria and pumice clasts ranging between 26.1 ± 1.7 and 11.06 ± 0.19 Ma (Encinas, 2006; Gutiérrez et al., 2009; Gutiérrez et al., submitted for publication). Martínez-Pardo (1990) also identified an early to middle Miocene warming event that extended north from about 45°S (Forsythe et al., 1985), based on the presence of planktonic foraminifer genera such as *Globigerinoides* and *Sphaeroidinellopsis*, together with *Globoquadrina venezuelana*, *Globigerinita glutinata*, and the *Globorotalia mayeri–G. siakensis* plexus. These taxa are similar to those present in tropical areas during the early to middle Miocene (Stainforth, 1953; Bertels, 1979; Srinivasan and Kennett, 1981; Kennett et al., 1985), indicating that oceanic circulation and temperature gradients of marine waters in the Southeast Pacific were more uniform than those prevailing today. Sea surface temperatures reached about 25 °C (Sanchetta, 1978; Loutit et al., 1983).

Tsuchi (1992) recorded a warming event, which he and subsequent authors referred to as the Miocene Climatic Optimum that culminated at about 16 Ma in Japan. This event (here referred to as the Burdigalian–Langhian Warming or BLW for consistency of notation; see Table 1) is revealed by the presence of warm-water planktonic foraminifera in the Mejillones Basin (Fig. 4) north of Antofagasta (Ibaraki, 1990; Martínez-Pardo, 1990; Tsuchi, 1990) and the Coquimbo Basin (Fig. 4) near Tongoy (Martínez-Pardo, 1979; Le Roux et al., 2006). This is reinforced by the presence of typical equatorial diatoms such as *Actinoptychus senarius*, *Actinocyclus ellipticus*, and *Coscinodiscus oculoiridis* in the Coquimbo Formation (Muhina, 1971; Martínez-Pardo and Caro, 1980), as well as in the Mejillones Basin (Frenguelli, 1949; Martínez-Pardo, 1978). Sea surface temperatures were estimated to have been around 25 °C at this time.

Abrupt cooling of ocean water began at 15 Ma in Japan and Peru (Tsuchi, 1992), but seems to have been more gradual in central to northern Chile, where it only manifested clearly during the late Miocene (Gutiérrez et al., submitted for publication). This episode is here referred to as the Serravalian–Tortonian Cooling (STC). Marchant et al. (2000) reported that subpolar temperatures similar to those of today were established in the Bahía Inglesa Formation of the Caldera Basin (Fig. 4) by about 10 Ma, as indicated by the left-coiling foraminifer *Neogloboquadrina pachyderma*. Covacevich and Frassinetti (1986) attribute the low species diversity of the mollusk family Cancellariidae between the Taitao Peninsula and Navidad (Fig. 1) to temperate conditions. However, the possible early to late Miocene age range given by these authors is rather wide, with the exception of *Cancellaria crossletensis* from the Taitao Basin (Fig. 4), which was considered to date from the middle to late Miocene. They also noted a significant decrease in the total number of mollusk genera and species in the Licancheu and Rapel Formations overlying the Navidad Formation, considered by Gutiérrez et al. (submitted for publication) to be of late Miocene age. It is probable that the low species diversity of Cancellariidae and other mollusks indicates a late Miocene cooling episode. Kiel and Nielsen (2010) attribute the present-day low diversity of mollusk species north of 42°S to a late Neogene extinction and only moderate subsequent origination in this area. This extinction, although long recognized (Möricke, 1896; Herm, 1969), was only recently explained by anoxia due to upwelling, as suggested by the enhanced survival of infaunal bivalves (Rivadeneira and Marquet, 2007). Intensified upwelling normally results from a northward shift in the climatic belts and ocean currents, which would have coincided with cooler oceanic conditions during the late Miocene. In southern Chile, cooler temperatures also commenced during the late Miocene, as manifested in the Chaicayán Group of the Taitao Basin which contains planktonic foraminifera including *Globigerina apertura*, *Globorotalia* cf. *acostaensis*, *Globorotalia conoidea*, *Globorotalia conomiozea*, and *Globorotalia continuosa*, as well as *Orbulina suturalis*. These indicate temperate water masses (Forsythe et al., 1985).

Tsuchi (1992) reported two warming episodes during the late Miocene and Pliocene, at about 6 and 3 Ma, respectively. These are here considered together as being part of the Messinian–Pliocene Warming (MPW). The first event was recorded in the Navidad Formation of central Chile and also coincided with a period of maximum marine transgression reflected by the late Miocene to earliest Pliocene succession in the Mejillones Basin, where warmer water diatoms such as *Coscinodiscus asteromphalus* and *Coscinodiscus radiatus* are present (Krebs et al., 1992). This is supported by the presence of tropical mollusks and warm-water planktonic foraminifera (Finger et al., 2007). In northern Chile, a correlative succession 80 km south of Iquique (Fig. 1) was interpreted by Padilla and Elgueta (1992) to indicate marine transgression of temperate-warm water, because of similar diatom species to those in Mejillones. In the Caldera Basin (Fig. 4), Marchant et al. (2000) reported the subtropical species *Globorotalia menardii* in beds dated around 6 Ma in the Bahía Inglesa Formation. Le Roux et al. (2005b) identified a sea-level highstand at 5.6 Ma at Carrizalillo in the Coquimbo Basin, which correlates with a global highstand on both the Haq et al. (1988) and Abreu et al. (2000) sea-level curves. They interpreted this highstand to reflect a period of increased activity along the East Pacific Rise.

Table 1
Relation between tectonic and climatic events during the Tertiary.

Climatic events during Teriary period	Age (Ma)	Principal Tertiary tectonic and oceanic events	Dated ages of events (Ma)
Santonian–Danian Cooling: SDC	86–60	Decrease in South Pacific spreading rate (Conrad and Lithgow-Bertelloni, 2007)	60
Thanetian–Ypresian Warming: TYW	60–55	Start of compressional regime in central Andes (Sempere et al., 1997)	58
		1st phase of Andean uplift in Tierra del Fuego (Olivero et al., 2003)	57
		Increase in South Pacific spreading rate (Conrad and Lithgow-Bertelloni, 2007)	55
Ypresian Cooling: YC	55–49	Extension in Tierra del Fuego (Ghiglione et al., 2008)	56–49
Lutetian Warming: LW	49–41	High convergence rate of South American-Farallon Plates (Pardo-Casas and Molnar, 1987)	50–40
		Increase in spreading rate between SSA and PA (Livermore et al., 2005; Eagles et al., 2006)	49
		2nd phase of Andean uplift in Tierra del Fuego (Olivero et al., 2003)	48
		Rifting on Kerguelen Plateau and Broken Ridge (Royer and Sandwell, 1989)	46–41
		Increase in spreading rate between Australia and Antarctica (Royer and Sandwell, 1989)	41
		Uplift of Himalayas (Pearson, 2010)	40
Bartonian–Rupelian Cooling: BRC	41–28	First influx of Pacific seawater into Atlantic Ocean (Scher and Martin, 2006)	41
		Decrease in South Pacific spreading rate (Conrad and Lithgow-Bertelloni, 2007)	40
		Development of Antarctic Circum-polar Current (Barker, 2001)	40
		Extensional tectonics in central Chile (Charrier et al., 2002)	36–26
		Extensional tectonics in central and northern Chile (González, 1989)	34
		Oi-1 sea-level fall (Zachos et al., 1996)	34
Chattian Warming: CW	28–24	Breakup of Farrellon Plate (Herron and Heirtzler, 1967)	28–23
		Commencement of spreading in West Scotia Ridge (Barker, 2001)	28
		Opening of Bransfield Basin (Sell et al., 2004)	28
		Change in convergence direction of Farrallon-Nazca Plate (Ghiglione and Cristallini, 2007)	27
		Increase in South Pacific spreading rate and global sea-floor production (Somoza, 1998; Muñoz et al., 2000; Conrad and Lithgow-Bertelloni, 2007)	25
Aquitanian Cooling: AC	24–21	Decrease in Southwest Indian Ocean spreading rate (Patriat et al., 2008)	24
		Decrease in North Pacific spreading rate (Cande and Kent, 1992)	24
		Increased deep circulation of Antarctic Circum-polar Current (Lyle et al., 2007)	25–23
		Development of bottom scouring currents in Southern Indian Ocean (Shipboard Scientific Party, 1989)	25–23
		Mi-1 sea-level fall (Miller et al., 1991; Naish et al., 2008)	23–21
		Narrowing of Drake Passage (Lagabrielle et al., 2009)	23–21
Burdigalian–Langhian Warming: BLW	21–15	Tectonic inversion in central Chile (Charrier et al., 2002)	21–16
		High convergence rate between South American-Nazca Plates (Pardo-Casas and Molnar, 1987)	21–15
		Expansion of volcanism in Andean retroarc (Ramos et al., 2002)	19–15
		Quechua Phase of deformation in central Andes (Malumián and Ramos, 1984; Ramos, 2002) and Patagonian Andes (Vásquez et al., submitted for publication)	18
		Collision of Chilean Ridge with SSA (Malumián and Ramos, 1984)	18
		Nazca–Phoenix–Antarctic triple junction enters South American Trench (Breitsprecher and Thorkelson, 2009)	17–14
		Monterey Carbon Excursion (Vincent and Berger, 1985)	17–14
		Rain-shadow effect of Andes established in Patagonia (Blisniuk et al., 2005)	16.5
		Increase in global sea-floor production (Conrad and Lithgow-Bertelloni, 2007)	15
		Increase in Japan sea-floor spreading (Tsuchi, 1992)	15
Serravalian–Tortonian Cooling: STC	15–6	Subsidence in central Chile (Le Roux et al., 2005a,b, 2006)	15–11
		Activation of Sandwich spreading center (Vanneste et al., 2002)	15
		Increased deep circulation of Antarctic Circum-polar Current (Barreda and Caccavari, 1992)	14–11
		Decrease in South-Pacific spreading rate (Le Roux et al., 2005a)	12–10
Messinian–Pliocene Warming: MPW	6–2.8	Increase in South-Pacific spreading rate (Cande and Leslie, 1986; Le Roux et al., 2005a)	8–7
		Uplift of Bolivian Orocline (Ghosh et al., 2006)	8–7
		Increase in spreading rate of East Pacific Rise (Rea, 1976)	6–2
		Closure of American Seaway (Evenstar et al., 2009)	3.5–3

The second warm-water episode around 3 Ma is also represented in the southeast Pacific from northern to south-central Chile, although it apparently extended over a longer period. In northern Chile, this is indicated by the onset of diatomite deposition at 3 Ma (Tsuchi, 1992). However, in late Pliocene beds of this area, Martínez-Pardo (1990) identified *Globigerinoides conglobatus* and *Globigerinoides ruber*, together with *Globorotalia crassaformis* and *Globorotalia inflata*, indicating a warming event between 3.6 and 2.6 Ma. Marchant et al. (2000) also recorded *Globorotalia calida* and *G. crassaformis* in beds dated at around 3 Ma in the Bahía Inglesa Formation. At this time sea surface temperatures increased by 2 to 3 °C to reach about 14°–15 °C in central Chile (Martínez-Pardo and Osorio, 1968). In the Ranquil Formation of the Arauco Basin (Fig. 4) south of Concepción (Fig. 1), the gastropods *Zonaria frassinetti* and *Solatisonax bieleri* indicate sub-tropical to tropical water temperatures between 4.6 and 4.4 Ma (Groves and Nielsen, 2003; Finger et al., 2007; Nielsen and Frassinetti, 2007; Le Roux et al., 2008a, b). Such conditions are supported by the presence of the foraminifer *Pulleniatina primalis* (Kennett and Srinivasan, 1983). It thus seems that temperate-warm conditions existed more or less continuously from about 6 to 2.6 Ma in central and northern Chile, with possibly minor colder periods. However, cooler conditions were already established in southern Chile during the late Pliocene. In the Taitao Basin the late Pliocene to Pleistocene Seno Hopper succession contains *Globigerina bulloides*, *Globigerina pachyderma* (left-coiling), and *Globigerina quinqueloba*, as well as *Globorotalia* cf. *hirsuta*, *G. inflata*, *Globorotalia scitula* and *Globorotalia pseudopachyderma*, suggesting cool-temperate conditions, with the presence of left-coiling *G. pachyderma* indicating that cold subpolar waters also affected this region (Forsythe et al., 1985). On Guamblín Island (Taitao Basin), upper Pliocene deposits contain marine mollusks and foraminifers similar to that of the lower Pleistocene Tubul Formation of the Araucuo Basin, indicating temperate-cold (~10 °C) sea surface temperatures (Frassinetti and Covacevich, 1995; Rojas and Marchant, 2000).

3.2. Southern Ocean

Various studies suggest a general decreasing trend in water temperatures since the middle-Late Cretaceous around the Antarctic

Peninsula (Ditchfield et al., 1994). Relatively low water temperatures during the early Paleocene, for example, are indicated by isotopic analysis of foraminifera in the Weddell Sea (Barrera and Huber, 1990, 1991).

The worldwide Paleocene–Eocene Thermal Maximum, here referred to as the Thanetian–Ypresian Warming (TYW), peaked at about 55 Ma and was characterized by a 2–6‰ negative carbon isotope excursion, rapidly escalating global warming of 5–8 °C, and a change from a relatively well-mixed to stratified, warm ocean. Seasonal, extreme precipitation events increased in tropical areas and high latitudes but decreased in subtropical areas, implying increased humidity gradients (Dickens et al., 1995; Boersma et al., 1998; Houghton et al., 2001; Schmitz and Pujalte, 2003, 2007; Wing et al., 2005; Gibbs et al., 2006). This event is clearly detected on the Maud Rise in the eastern Weddell Sea, where Thomas et al. (2002) made a detailed study of $\delta^{18}O$ and $\delta^{13}C$ values across the Paleocene–Eocene boundary. They detected a brief period of gradual surface warming of around 2 °C, followed by a geologically instantaneous negative carbon isotope excursion of 3–4‰ and a rise of 4–8 °C to reach a sea surface temperature of about 18 °C (Fig. 6). Thomas et al. (2002) concluded that the thermal anomaly probably reached the Southern Ocean from a distal source region and via vertical diffusion.

Ivany et al. (2008) determined sea surface temperatures of about 15 °C in lower Eocene deposits on Seymour Island (Fig. 3), which is here correlated with the Ypresian Cooling (YP). However, from this time on progressive warming occurred into the middle Eocene. Cione et al. (2007) proposed that fish and invertebrate fauna in the Cucullaea I Member of the La Meseta Formation on Seymour Island indicate relatively warmer temperatures during this period, which they attributed to the warmer waters of a current reaching the region from the north. On the Maud Rise, a $\delta^{18}O$ warming anomaly was reported near 41 Ma by Diester-Haass and Zahn (1996), also detected on the Kerguelen Plateau of the southern Indian Ocean (Barrera and Huber, 1993). Bohathy and Zachos (2003) dated this event, associated with a 4 °C rise in temperature, at 41.5 Ma. This warming episode, referred to in this paper as the Lutetian Warming (LW), coincides with higher sea levels and widespread carbonate deposition in shelf areas, as well as the expansion of tropical foraminifer assemblages to middle and high latitudes in the Indo-Australasian region (McGowran, 1977).

A long-term cooling trend began shortly hereafter (Miller et al., 1987; Zachos et al., 2001), culminating with a sharp drop in temperature around the Eocene–Oligocene boundary as indicated by fauna in the uppermost part of the La Meseta Formation (Cione et al., 2007). This forms part of what is here termed the Bartonian–Rupelian Cooling (BRC). At the South Orkney Islands, the presence of *Chiasmolithus gigas* and *Nannotetrina quadrata* still indicates generally warm and equitable waters during the middle Eocene, but by the late Eocene these taxa had become scarce at those latitudes (Edwards and Perch-Nielsen, 1975; Wise, 1983). An isotopic shift in the $\delta^{18}O$ and δD values of hydrothermally altered rocks in northern Victoria Land suggests a cooling episode in the Ross Sea region during the late Eocene at about 40 Ma, with a minimum temperature decrease of 5–9 °C (Dallai et al., 2001). Calcareous nannofossils in Southern Ocean deposits show a distinct shift to cool-water taxa around 36 Ma (Wei, 1991), suggesting that cool waters continued to expand toward the north (Wise et al., 1985). The late Eocene planktonic foraminifer assemblage also began to assume a polar character with low diversity (Stott and Kennett, 1990). Furthermore, $\delta^{18}O$ data of fossil mollusks on James Ross Island show that temperate-cold to sub-polar conditions were established on the present Antarctic Peninsula during the latest Eocene–early Oligocene, where nautiloids, oysters, and bivalves in the La Meseta Formation indicate marine water temperatures between 7.9 and 11.7 °C (Ditchfield et al., 1994). Deep-sea benthic foraminifers also have $\delta^{18}O$ values indicating a decrease of about 2.5 °C in ocean water temperature at 34 Ma (Lear et al., 2008), which coincides with the global Oi-1 cooling event of

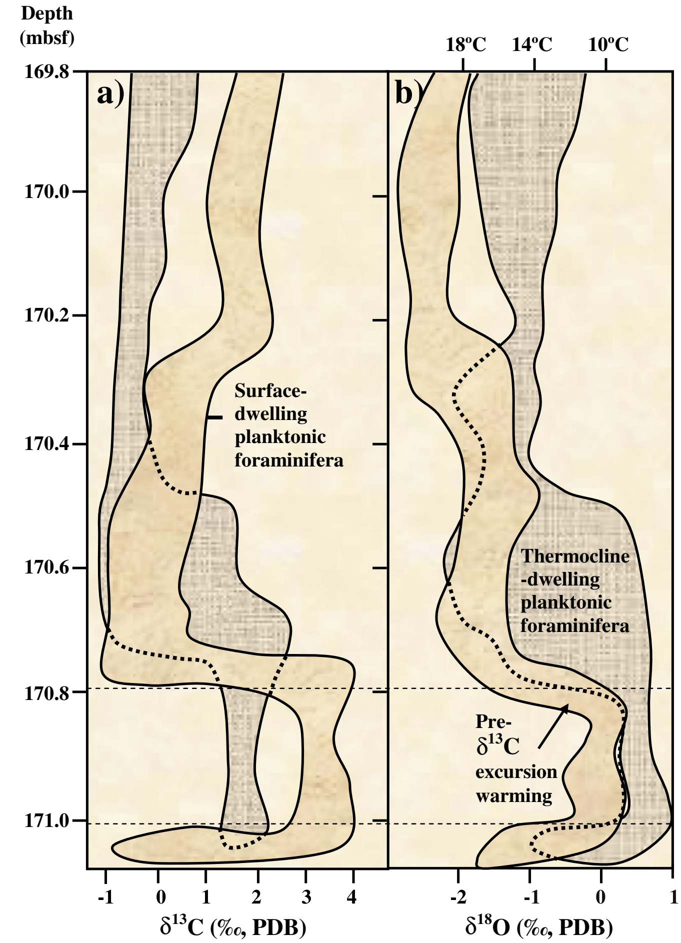

Fig. 6. $\delta^{13}C$ and $\delta^{18}O$ excursions in the eastern Weddell Sea shown by surface- and thermocline-dwelling planktonic foraminifera at the TYW (modified after Thomas et al., 2002). $\delta^{18}O$ values of surface-dwelling foraminifera (demarcated by light brown envelope in Fig. 6b) show that sea-surface temperatures cooled off drastically from an earlier thermal spike before being warmed by about 2 °C just prior to the main $\delta^{13}C$ excursions shown in Fig. 6a. The rapid cooling may be related to explosive volcanic eruptions releasing ash and sulfur into the atmosphere. The partial overlap between $\delta^{18}O$ values of surface- and thermocline-dwelling foraminifera during and shortly after the main $\delta^{13}C$ excursion could indicate the mixing of surface and thermocline waters by inversion.

Zachos et al. (1996) associated with a 70 m sea-level fall as estimated from sequence stratigraphy (Pekar et al., 2002).

Early Oligocene diatom assemblages are typically sub-Antarctic in character, with warm to temperate-warm taxa relatively rare (Baldauf and Barron, 1991). During the late Oligocene, however, the global climate warmed again (Zachos et al., 2001; De Man and Van Simaeys, 2004) as for example manifested on the Kerguelen Plateau, where paleoecological changes in the nannofossil assemblages indicate relatively stable, cool conditions from earliest to late Oligocene times. However, an increase in the abundance of temperate-water taxa is recorded at about 26 Ma, here associated with the Chattian Warming (CW) (Villa and Persico, 2006).

The oxygen isotope curve of cores from the late Oligocene–early Miocene in the western Ross Sea shows an increase and subsequent decrease in oceanic mean $\delta^{18}O$ across the Oligocene–Miocene boundary, which indicates a sharp drop in temperature and considerable expansion of the East Antarctic Ice Sheet (Naish et al., 2008). This event is here associated with the Aquitanian Cooling (AC), which was relatively short-lived.

Oxygen and carbon isotope evidence from the Ross Sea indicates that the production of warm, Tethyan-Indian Saline Water after the AC increased up to about 16 Ma, which caused meridional heat transport to the Antarctic and inhibited polar cooling and the growth of the East Antarctic Ice Sheet (Woodruff and Savin, 1989, 1991; Wright et al., 1992). Correlated with the BLW, this event might be linked to a simultaneous increase in the mean ocean $\delta^{13}C$ (the so-called Monterey Carbon Excursion) between 17 and 13.5 Ma, which caused the deposition of large volumes of organic carbon and phosphate (Vincent and Berger, 1985; Woodruff and Savin, 1991). This can probably be attributed to the large-scale release of nutrients along active mid-ocean ridges, which encourages the growth of plankton that consume oxygen and cause anoxia (Kerr, 1998).

The BLW was followed by rapid cooling during the middle Miocene (STC) (Shackleton and Kennett, 1975; Savin et al., 1985; Miller et al., 1987). Until 14.8 Ma, the Tethyan-Indian Saline Water competed with the Southern Component Water derived from a cold, high-latitude source, causing marked fluctuations in sea level as well as in the Antarctic climate and cryosphere (Flower and Kennett, 1993b). At 14.8 Ma, however, a reduction in the Tethyan–Indian Saline Water led to a definite cooling of the region, causing major growth of the East Antarctic Ice Sheet, an increase in the meridional temperature gradient and an aridification in mid-latitude continental areas (Woodruff and Savin, 1989). In the James Ross Basin (Fig. 3), late Miocene fossilliferous diamictites contain megafauna and microfauna (*Lenticulina* and *Astacolus*) suggesting normal marine salinity conditions and higher temperatures than at present at the same latitude. However, the presence of *Cassidulinoides parkerianus* also indicates cold water influences (Concheyro et al., 2007). Dingle and Lavelle (1998), based on an increase in ice-rafted debris and higher $\delta^{18}O$ values, proposed a renewed climatic deterioration in the Southern Ocean during the late Miocene (~9.3 Ma) and the beginning of glacial cycles with continental extension during the Pliocene.

The micropaleontological content of the Belén and Gage Formations on James Ross Island indicates interglacial periods during the late Miocene and late Pliocene respectively, which might correlate with the warming episodes observed in the Southeast Pacific between 6 and 2.6 Ma (MPW). Bohathy and Harwood (1998) used the biostratigraphic record of the silicoflagellate genera *Dictyocha* and *Distephanus* together with calcareous nannoplankton to examine sea surface temperature changes on the southern Kerguelen Plateau during the Pliocene. Three distinct peaks in the abundance of these microfossils reflect a warming event of at least 4 °C through either southward migration of the Antarctic Polar Frontal Zone or a weakening of the thermal gradient across this zone. Peak warming in the intervals studied occurred at about 4.3 Ma, with other warming events near 4.5, 4.2, and 3.6 Ma. A glacial advance toward the end of the Pliocene in the James Ross Basin is indicated by the lithological characteristics of the Terrapin Formation and the presence of clasts and concretions with striations (Concheyro et al., 2007). This is confirmed by the micropaleontological content of this formation (Lírio et al., 2003).

3.3. Southwest Atlantic Ocean

In Patagonia and Tierra del Fuego, warmer sea water conditions generally correspond to major Cenozoic transgression episodes (Malumián and Jannou, 2010), so that sea-level variations can be used in conjunction with paleontological evidence to determine relative water temperatures.

In the Southwest Atlantic Ocean, a warming trend similar to that detected in the Southern Ocean is manifested up to the Paleocene–Eocene boundary (TYW). Waters became relatively warm during the late Paleocene in the area around the Falkland/Malvinas Plateau, as suggested by the presence of *Discoaster multiradiatus* (Wise, 1988). In the Punta Noguera Formation of Argentinean Tierra del Fuego, the association of benthic foraminifers shows that maximum temperatures were recorded at around 55.8 Ma. The dinoflagellate *Apectodinium* also indicates an elevated temperature (Olivero et al., 2002), but not high enough to allow the incursion of large foraminifers typical of para-subtropical conditions (Malumián and Jannou, 2010). Such large foraminifers are in fact absent from the Tertiary in Patagonia, in contrast to many of the other Southern Hemisphere fragments of Gondwanaland, and only appear further north in the offshore Santos Basin of Brazil (Abreu and Viviers, 1993).

In the Fuegian Andes of Tierra del Fuego, there was a great turnover from a cosmopolitan Midway-type calcareous microfossil assemblage in the La Barca Formation to a strongly endemic early Eocene assemblage in the Punta Noguera Formation, in which the low percentage of the ostracod family Hemicytheridae suggests the commencement of a cool water period correlated with the YC (Malumián and Jannou, 2010). A decrease in water temperature is also interpreted in the lower to middle Eocene Punta Torcida Formation of Tierra del Fuego (McGowran and Beecroft, 1985), where Malumián and Jannou (2010) identified the planktonic, temperate water foraminifer *Jenkinsina triseriata*.

This cooling trend was subsequently reversed during the LW, as shown by an increase in Hemicyteridae in the overlying, middle Eocene Río Bueno Formation (Malumián and Jannou, 2010). Regional transgression was coeval with a temperature peak at 42 Ma (Leticia Formation), during which a foraminiferal assemblage with large-sized nodosarids appeared.

This foraminiferal assemblage was replaced during the late Eocene (Cerro Colorado Formation) by typical Antarctic genera that reflect falling temperatures, culminating in the *Tenuitella insolita* Zone with abundant *Chiloguembelina* (Malumián and Jannou, 2010). The low smectite content and the abundance of *Chiloguembelina ototara* also indicate falling temperatures during the deposition of the upper member of the Cerro Colorado Formation and the Glauconítico A, which correspond to the BRC. The Eocene/Oligocene boundary at 33.9 Ma in Tierra del Fuego is marked by an unconformity related to the Oi-1 sea-level fall. However, in the earliest Oligocene (María Cristina beds) deepwater foraminifers appeared, contrasting with coeval global high $\delta^{18}O$ values and a late Eocene–early Oligocene regression on the Patagonian Platform, which suggest that the deepening recorded in the Fuegian Andes was due to tectonic causes (Malumián and Jannou, 2010). The middle Oligocene Puesto Herminita beds were deposited after the maximum deepening of the basin that allowed the entrance of corrosive Antarctic waters.

The younger Desdémona Formation consists of the Cabo Ladrillero beds containing the benthic foraminifera *Spirosigmoilinella* and *Martinottiella*, and the Cabo San Pablo beds, representing late Oligocene high sea-level conditions probably coinciding with warmer sea-surface temperatures of the CW (Malumián and Jannou, 2010).

The discordant contact of the San Julián Formation with the lower Miocene Monte León Formation in the Santa Cruz Province suggests a renewed regression, probably linked to the Mi-1 glaciation of the AC (Miller et al., 1991; Malumián and Jannou, 2010). The Monte León Formation registers the influx of cold Antarctic waters, possibly related to this drop in sea-level and narrowing of the Drake Passage between 23 and 21 Ma (Lagabrielle et al., 2009). This cold-water incursion may have reached as far north as 42°S (Malumián and Náñez, 1991) and was accompanied by generalized regression in Patagonia (Malumián and Jannou, 2010). Further evidence for a climatic deterioration to temperate-cold water during the AC is furnished by the foraminifer association *Martinottiella–Spirosigmoilinella* in the Monte León Formation at about 49°S (Náñez, 1988; Barreda and Palamarczuk, 2000). The presence of *Globocassidulina* in the lower Miocene Malengüena Formation of the Mitre Peninsula (Fig. 2) also suggests temperate-cold waters (Torres Carbonell et al., 2009).

In Tiera del Fuego, the lower to middle Miocene Carmen Silva, Castillo, and Cabo Peña Formations, together with the Cabo Viamonte beds, show an increased proportion of smectite, which is a common feature in the whole subcontinent related to the BLW (Alonso et al., 2001; Malumián and Olivero, 2006). A warming episode associated with marine transgression is also suggested by the presence of *Neovenericor* and *Abertella* in the lower middle Miocene Saladar Member of the Gran Bajo de Gualicho Formation of northeastern Patagonia (Reichler, 2010). In addition, this formation contains other mollusk genera such as *Siphocypraea*, *Sconsia*, *Ficus*, *Chicoreus*, *Mitra*, and *Conus* similar to Caribbean warm-water associations, although the presence of some temperate-cold water genera such as *Glycymerita* and *Perissodonta* may indicate temperate-warm conditions (Reichler, 2010). In the lower to middle Miocene Chenque Formation of the Golfo de San Jorge Basin (Fig. 4), the endemic mollusk species *Nodipecten contenida* is associated with paratropical taxa, which also suggests the development of temperate-warm waters in this region coinciding with the BLW (Del Río, 2006).

Toward the end of the middle Miocene, *N. contenida* became extinct in Patagonia, withdrawing to tropical latitudes. This indicates an abrupt decrease in the water temperature (Del Río, 2006) associated with the STC. The appearance of *Reticulofenestra perplexa* in middle to late Miocene deposits on the Falkland/Malvinas Plateau also reflects the advent of increasingly colder water temperatures in this area, which is supported by the gradual disappearance of warmer water calcareous nannoplankton (Wise, 1988).

Martínez and Del Río (2002) concluded from an analysis of late Miocene mollusks along the southwestern Atlantic coast that these fauna did not give rise to the present assemblages of the Argentinean Province, which were characterized by a transitional association that only developed once the cold Falkland/Malvinas Current was fully operational. The southern limit of the late Miocene Valdesian Province, which includes assemblages recovered from the Puerto Madryn Formation in the Valdés Basin, was placed at 42°S, while its northern boundary with the Paranaian Province was situated between approximately 37 and 39°S. The Paranaian Province is composed mainly of paratropical warm-water fauna, including assemblages belonging to the Paraná and Camacho Formations and extending northwards along the Uruguayan and southern Brazilian coast. Considering that the water temperature in the Valdesian Province was warm, the cold Falkland/Malvinas Current probably had a minor effect on this region at the time.

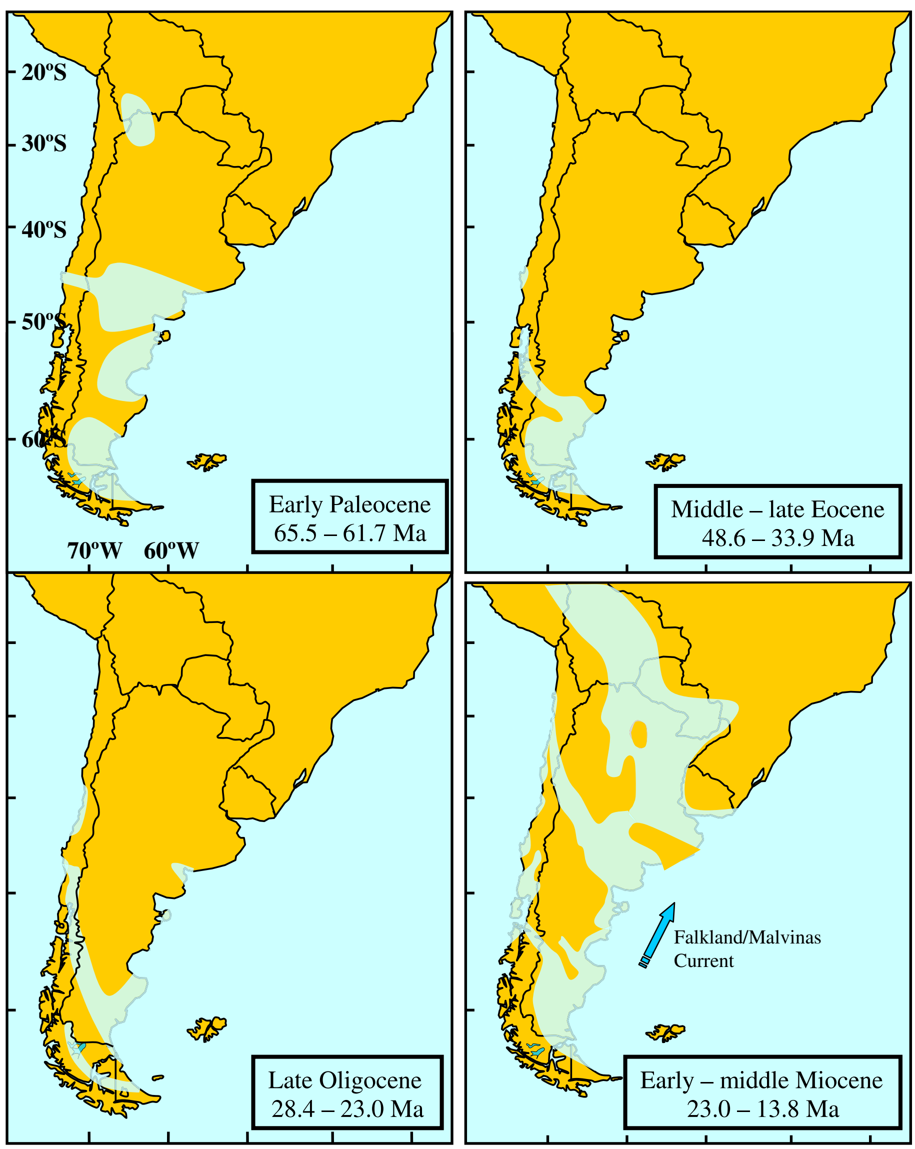

Fig. 7. Approximate incursions of the Atlantic and Pacific Oceans into SSA, here referred to as the Transcontinental Sea. Based on Uliana and Biddle (1988), Aceñolaza (2004), Malumián and Nañez (2011) and other sources mentioned in text. As in Fig. 4, the basin margins do not represent their shapes during any specific period within the indicated time intervals.

3.4. Transcontinental Sea

Tectonic processes such as plate subduction, rollback, warping and mountain building affected SSA throughout the Tertiary, generating diverse marine depocenters in central SSA that at times connected the Pacific with the Atlantic Ocean (Fig. 7).

In Patagonia, the Cretaceous–Cenozoic Magellanes/Austral Foreland Basin (Fig. 4) trends from northwest to southeast, crossing Tierra del Fuego to the Atlantic coast. Marine deposits in this basin range from the lower Cretaceous to the lower Miocene. For example, in the southeastern part of this basin, deepwater conditions existed during the late Maastrichtian to late Paleocene, probably connecting to the Southwest Atlantic. The late Paleocene Cabo Nariz Beds on the west coast of Tierra del Fuego, interpreted as deepwater turbidites by Sánchez et al. (2010), prograded toward the west-northwest (J.P. Le Roux, unpublished data), which suggests that higher ground existed toward the east.

On the Brunswick Peninsula west of Punta Arenas, the late Cretaceous Chorrillo Chico Formation has lenses of glauconitic sandstones (Charrier and Lahsen, 1968) indicating a shelf to bathial environment and probably cool water conditions (Porrenga, 1967), which may be related to the influx of cold Atlantic waters into this part of the basin during the SDC.

The earliest evidence of a Pacific-Atlantic connection may be indicated by the presence of elasmosaurid plesiosaurs in the Cerro Dorothea Formation of west-central Patagonia (Otero et al., 2009), considered to be of Maastrichtian to early Paleocene age, the latter based on the presence of Danian foraminifers (Caramés, 1996; Malumián and Caramés, 1997). Plesiosaurs of this family had a marked abundance in the Pacific at the end of the Cretaceous, having also been found in the Quiriquina Formation near Concepción (Gasparini, 1979), and further north up to Valparaíso (Suárez et al., 2003). Other fauna previously known only from the Quiriquina Formation and its equivalents in central Chile, including *Ischyrhiza chilensis*, *Centrophoroides*, and an as yet unnamed dasyatid ray, have also been recovered from the Baguales Range north of Puerto Natales (Le Roux et al., 2010). Water temperatures in the Transcontinental Sea at this time were probably temperate to warm, because elasmosaurid plesiosaurs were likely to have been cold-blooded and would have required such conditions to survive. In the Cerro Dorotea Formation, planktonic foraminifera include the early Paleocene species *Parasubbotina pseudobulloides*, which also suggests temperate waters (Arenillas and Molina, 1995; Malumián and Caramés, 1997). This implies that the Transcontinental Sea became warmer than the Southwest Atlantic during the early Paleocene, possibly because of the influence of Pacific waters from the north-west, the reduced influence of cold Antarctic waters, and its enclosure within a continental setting. Marine conditions also existed during the early to middle Paleocene in northern Argentina and south-eastern Bolivia, as indicated by marine taxa in the El Molino Formation (Sempere et al., 1997). This seaway may thus have stretched across SSA, but thereafter retreated at least from the Altiplano/Puna region, as no Eocene–Oligocene marine deposits have been recorded there (Gregory-Wodzicki, 2000).

Marine conditions persisted during the late Paleocene–early Eocene along the eastern part of the Brunswick Peninsula, as indicated by the sedimentology of the Agua Fresca Formation (Charrier and Lahsen, 1968). However, an unconformable contact developed between the Cerro Dorothea and Río Turbio Formations during a major regression throughout Patagonia in the middle to late Paleocene (Malumián and Caramés, 1997), which was probably due to tectonic uplift.

Warm conditions during the middle Eocene are suggested by a kaolinite content peak in the basal part of the Río Turbio Formation, probably reflecting the LW (Malumián and Caramés, 1997). At the end of this period, the Patagonian Andes underwent a period of uplift corresponding to the Incaic Phase (Ramos, 2002) that exposed marine deposits in the Austral/Magellanes Basin to subaerial erosion in certain areas, e.g. in the Mitre Peninsula (Torres Carbonell et al., 2009). Further north, the Río Turbio Formation was deposited in a littoral environment with shifting coastlines, probably in response to the marine regression caused by this uplift as well as regional cooling. It is unconformably overlain by thick fluvial conglomerates of the middle Eocene Río Guillermo Formation (Malumián and Caramés, 1997). In the Lake Argentino area, the upper middle Eocene Man Aike Formation overlies upper Cretaceous rocks of the Calafate Formation with a pronounced unconformity (Malumián, 2002). Sedimentation of the Man Aike Formation took place in a subtidal environment, which indicates a period of renewed marine transgression after this erosional interval, correlated here with the LW (Malumián, 1990, 2002; Marenssi et al., 2002). This transgression is also indicated by the marine Leticia Formation of Argentinian Tierra del Fuego, but not further north in the Golfo de San Jorge or Valdés Basins (Malumián, 1999; Caramés et al., 2004) (Fig. 4). Coinciding with a temperature peak at 42 Ma, this sea-level highstand probably enhanced the connection between the Pacific and Atlantic Oceans, the former contributing warmer water to the region as suggested by Ne isotopes (Scher and Martin, 2006).

On the Brunswick Peninsula, the Leña Dura Formation is assigned to the middle-late Eocene based on palynomorphs and invertebrate marine fossils (Cookson and Cranwell, 1967; Fasola, 1969). The overlying Loreto Formation was recently dated as late Eocene (Otero et al., in press) by detrital zircons and its rich diversity of fossil cartilaginous fishes, which have clear ecological affinities with Eocene Tethyan fauna previously described in the Northern Hemisphere. This assemblage also has elements in common with Eocene cartilaginous fishes from Antarctica. The depositional environment was interpreted as estuarine. The diversity of shark families, including odontaspidids, triakids such as *Galeorhinus* as well as myliobatids, suggests that temperate-warm conditions of the LW may have persisted into the late Eocene within this part of the Transcontinental Sea. North of Puerto Natales (Fig. 1), the basal portion of the Río Baguales Formation yielded a collection of sharks, rays and chimeroid fish remains including *Striatolamia macrota*, *Carcharocles* aff. *angustidens*, *Myliobatis* sp., and *Callorhinchus* aff. *regulbensis*, indicating a late Eocene age (Le Roux et al., 2010) supported by unpublished detrital zircon ages with a peak at 40.48 ± 0.37 Ma (C.M. Fanning, pers. comm., 2010). A fossil tooth of the sand tiger shark *Carcharias*, which presently occurs in warm-temperate and tropical coastal waters, was found above these beds. Marine invertebrates of Atlantic affinities in the stratigraphically equivalent Guadal Formation south of Lake General Carrera/Buenos Aires (Niemeyer et al., 1984; De la Cruz et al., 1996) also indicate a warm to subtropical sea (Frassinetti and Covacevich, 1999). This confirms the persistence of temperate-warm conditions in the central Transcontinental Sea until the late Eocene. Because the Southwest Atlantic was characterized by colder water conditions at the time, warm water could have entered the Transcontinental Sea from the Pacific Ocean, as previously proposed by Scher and Martin (2006).

In the Magellanes/Austral Basin, the Río Guillermo Formation is overlain by the uppermost Eocene–lower Oligocene, non-marine Río Leona Formation, reflecting marine regression associated with the BRC. Erosion during the rest of the middle Oligocene is indicated by the unconformable contact of the Río Leona Formation with the upper Oligocene Río Centinela Formation, which was deposited in marine conditions when a major Cenozoic transgression occurred in Patagonia during the CW (Malumián and Caramés, 1997).

The Río Foyel Formation of the Ñirihuau Basin (Fig. 4) contains the foraminifer *Transversigerina* cf. *transversa*, which suggests that the Pacific connection persisted during the Oligocene (Asensio et al., 2010). According to these authors, a Pacific connection requires the presence of Oligocene marine deposits in central and south-central Chile, which was recently confirmed by Gutiérrez et al. (submitted for publication) during a reevaluation of the age of the Navidad Formation as late Oligocene–early Miocene. Le Roux and Elgueta (2000) also mention the presence of *Panopea panis* near the base of a marine sedimentary succession in the Valdivia Basin (Fig. 4), which is of Oligocene age (Fenner and Sylvester, 1936). Furthermore, the basal deposits of the Osorno Basin (Fig. 4) were assigned to the Eocene (González, 1989; Elgueta et al., 2000) and Oligocene (Cisternas and Frutos, 1994), based on their palynological content. González (1989), in his study of Tertiary basins in Chile, detected a strong extensional phase during the Eocene–early Oligocene, which may have reinforced the Pacific connection to northern Patagonia. The same event was recorded by Charrier et al. (2002), who considered it to have lasted until about 26 Ma, i.e. overlapping with the CW.

North of Puerto Natales (Fig. 1), Le Roux et al. (2010) interpreted the depositional environment of the upper part of the Río Baguales Formation as a northwestward-facing estuary developing during the late Oligocene. This suggests that the Transcontinental Sea was bordered by a land mass to the southeast or east. Similarly, Cuitino and Scasso (2010) considered the Centinela or Estancia 25 de Mayo Formation south of Lake Argentino (Fig. 2) to have been deposited in a transgressive, shallow sea with oyster-rich biogenic accumulations, which evolved into a coastal plain with a progradational estuarine system. Marine fauna in the Río Baguales Formation are also recorded in the upper part of the Río Turbio Formation (Griffin, 1991) and the Guadal Formation of the Puerto Aisén region (Fig. 1) (Frassinetti and Covacevich, 1999). These formations therefore indicate the continued presence of the Transcontinental Sea in southeastern Patagonia during the Oligocene.

Le Roux et al. (2010) interpreted the sedimentology of the uppermost part of the Río Baguales Formation to reflect a sea-level fall commencing at about 23.3 Ma, which coincides with a global marine regression across the Oligocene–Miocene boundary and the Mi-1 glaciation. This indicates that the AC also affected the Transcontinental Sea.

Marine sedimentation apparently ended near the Oligocene/Miocene transition in west-central Patagonia, possibly related to marine regression combined with tectonic uplift. For example, the El Salto Formation overlying the Loreto Formation north of Punta Arenas (Fig. 1) lacks marine fossils. Considered to be of at least early Miocene age (Hoffstetter et al., 1957), it is absent north of Puerto Natales, where the Burdigalian Palomares Formation overlies the Río Baguales Formation directly (Le Roux et al., 2011). This indicates erosion in west-central Patagonia during the early Miocene. The Palomares Formation can be correlated with the Santa Cruz Formation of Argentina, as both contain Colhuehuapian-basal Santacrucian vertebrate fauna and are represented by meandering river deposits.

During the early Miocene, the Transcontinental Sea apparently shifted to the northeast. The Gaiman Formation of the Golfo de San Jorge Basin contains elasmobranchs, including *Carcharoides totuserratus*, which are typical temperate to warm water species, thus reflecting the BLW (Cione et al., 2011). In the same area, marine conditions persisted during the middle to late Miocene, where calcareous, siliciclastic and evaporitic sedimentary rocks of the Colorado Basin were deposited in a shallow coastal setting. This thin marine level underlies fluvial sandstones of the Río Negro Formation. Visconti et al. (2003) considered this unit to be contemporaneous with the Barranca Final Formation of the Colorado Basin, which is the depositional record of the "Entrerriense" marine transgression that affected the provinces of Entre Ríos, Buenos Aires and Patagonia during the BLW. The mollusks of this episode are typical of tropical and subtropical warm water masses and indicate higher sea surface temperatures than today (Aguirre and Farinati, 1999). The Transcontinental Sea hereafter invaded central, western and northern Argentina, reaching at least the Paraguayan Chaco and the southern border of Bolivia. Known locally as the Paranense Sea, it formed a biogeographic barrier between the Andean and pampas regions during the middle Miocene (Aceñolaza, 2004). Marine fauna in the middle–late Miocene Paraná Formation of northeastern Argentina also indicate an important marine incursion into this area, but having stronger links with Brazilian than Patagonian fauna (Aceñolaza, 2004). The presence of Pontoporiidae cetaceans suggests a biogeographic connection with the northern South Pacific, as similar taxa have been reported from northern Chile and southern Peru (Cione et al., 2000). This marine tongue also allowed the incursion of foraminifer species such as *Lippsina tisburyensis* and *Ammonia beccari parkinsoniana* from the Brazilian coast, which subsequently prospered in brackish and continental fresh water conditions. Some mollusks presently inhabiting the Brazilian coastline, such as *Crassiostrea rhizophorae* and *Nucula*, *Glycimeris*, *Corbula* and *Erodona* species, are also present in deposits of the so-called Querandinense incursion (Aceñolaza, 2004). The overlying Ituzaingó Formation, deposited in a fluvial environment, indicates regression from this area during the middle Miocene, possibly related to the STC. However, the Transcontinental Sea retreated from central, northeastern and northwestern Argentina, as well as from southeastern Bolivia, only during the late Miocene–Pliocene (Jordan and Alonso, 1987; Marshall et al., 1993; Aceñolaza, 2004; Urrutia et al., 2008).

During the early Pliocene, warmer sea water temperatures accompanied another marine transgression in Tierra del Fuego related to the MPW. The presence here of species presently endemic to offshore Peru, such as the foraminifer *Nonion hancocki* identified in the Irigoyen Formation of Tierra del Fuego (Malumián and Olivero, 2005; Malumián and Scarpa, 2005; Malumián and Jannou, 2010), may represent vestiges of the Transcontinental Sea temporarily linked by this event.

4. Discussion and conclusions

4.1. Summary of events

Fig. 5 shows how the climatic cycles discussed above relate to global deep-sea $\delta^{18}O$ records (red curve) based on data compiled from more than 40 Deep Sea Drilling Project and Ocean Drilling Project sites (Zachos et al., 2001) and the change in atmospheric CO_2 (blue curve) through time (Beerling and Royer, 2011). Although the $\delta^{18}O$ and CO_2 curves coincide to a large extent there are some discrepancies, for example during the TYW and BRC. Whereas the CO_2 curve confirms an increase in temperature leading up to the TYW, the $\delta^{18}O$ curve seems to indicate the opposite. However, during the BRC a warming period is suggested by the $\delta^{18}O$ curve, while the CO_2 curve indicates cooling. Some of the warming cycles discussed above coincide with both CO_2 and $\delta^{18}O$ records. For example, during the BLW, CO_2 levels were similar to those of today, and decreases in CO_2 were synchronous with major episodes of glacial expansion during the STC and PC (Tripati et al., 2009). The MPW or Messinian–Pliocene Warming is also clearly shown by the CO_2 curve, but not by $\delta^{18}O$. That atmospheric CO_2 played a large role in continental climatic variations around SSA and the AP is clear, as discussed in some detail in the companion paper (Le Roux, 2012). DeConto and Pollard (2003), Pollard and DeConto (2005) and Pearson et al. (2009) considered the sudden build-up and subsequent variations of Antarctic ice at the Oi-1 glaciation to be a result of a decline in atmospheric CO_2, orbital forcing, and ice-climate feedbacks. However, an alternative explanation could be that colder ocean surface waters simply trapped more CO_2 (see Section 4.4), and that the decline in CO_2 since the middle Eocene was a result and not the cause of global cooling.

In the Southeast Pacific, warm oceanic conditions seem to have prevailed from southern Peru to southern Chile throughout the Oligocene to middle Miocene, culminating near the end of the BLW (21–15 Ma). This was followed by cooling of the STC (15–6 Ma) starting during the late Miocene, which was succeeded by the MPW interval between about 6 and 2.6 Ma. The Pleistocene Cooling (PC: 2.6–0.01 Ma) in central and northern Chile seems to have started during the late Pliocene in southern Chile.

The Southern Ocean experienced a decrease in temperature since the late Cretaceous, corresponding to the SDC (86–60 Ma), but appeared to have warmed rather significantly and abruptly at the TYW (60–55 Ma). After an initial period of cooling during the YC (55–49 Ma), gradual warming continued until the LW (49–41 Ma), when the long-term cooling trend of the BRC (41–28 Ma) began. This was interrupted by a slightly warmer period in the CW (28–24 Ma). Cooling during the AC (24–21 Ma) was followed by renewed warming culminating in the BLW. Colder temperatures during the STC were succeeded by warming with fluctuating temperatures from about 6 to 2.6 Ma.

In the Southwest Atlantic Ocean, the Paleocene warming trend culminating at the TYW was followed by cooling up to the YC. Subsequent warming peaked during the LW, after which temperatures declined once more into the BRC. Tectonic subsidence in the Austral/Magellanes Basin is probably responsible for deepwater conditions around Tierra del Fuego during the Oligocene, but there is no direct evidence for warmer sea water temperatures. However, widespread regression during the AC probably corresponds to colder temperatures associated with the Mi-1 glaciation and the expansion of the East Antarctic Ice Sheet. This caused an incursion of cold Antarctic waters to about 42°S. Warming during the BLW was followed by cooling during the STC. North of 42°S along the Atlantic coastline, temperate-warm water conditions graded into paratropical conditions at about 38°S, suggesting that the cold Falkland/Malvinas Current did not extend this far north until at least the middle Miocene.

The Transcontinental Sea possibly connected the Pacific and Atlantic Oceans as early as the late Cretaceous. Cold Atlantic waters may have influenced the present Brunswick Peninsula during the SDC, possibly coinciding with a marine regression. Waters thereafter warmed up again until the TYW. It is not clear whether the subsequent regression registered in this area was caused by Andean uplift, colder water temperatures, or both, but it led to continental conditions and the development of unconformities. Renewed transgression corresponding to the LW coincided with a larger influx of water from the Southeast Pacific, which may have been responsible for the extension of warmer water conditions into the late Eocene. During the BRC, which lasted here until the middle Oligocene, marine regression exposed some of the stratigraphic units to subaerial erosion, but this was succeeded by a transgression during the CW. The Transcontinental Sea migrated to the northeast during the late Oligocene–early Miocene, followed by a sea-level fall during the AC. The BLW is also well represented by marine transgressions and warm water fauna in the Colorado Basin and further north, but continental conditions resumed during the regression of the STC. Finally, warmer temperatures are recorded in Tierra del Fuego during the MPW.

4.2. Relationship between paleoclimate changes and plate tectonics

A possible cause for these alternating warming and cooling events may be sought in cycles of tectonic activity in the mid-ocean ridges. Ridge activity accounted for major sea-level rises during the Eocene and Miocene (Pitman, 1978). For example, a ridge spreading at a rate of 6 cm/yr will have 3 times the volume of one spreading at 2 cm/yr, and considering that the mid-ocean ridges have a total length of 45,000 km (Hays and Pitman, 1973), their effect on the world's oceans should be considerable. Increasing ridge volumes accompanied by the outpouring of ocean floor basalt resulting in a hot, less dense oceanic crust, would cause sea-level highstands (Hallam, 1963) and also an increase in sea-water temperatures.

An association between ocean temperatures and ocean ridge activity seems to have characterized the Tertiary. Conrad and Lithgow-Bertelloni (2007) examined the spatial gradients of present-day sea-floor ages and inferred ages for the subducted Farallon Plate, which they used to construct a history of spreading rates since 140 Ma. The average global spreading rate for the Late Cretaceous–early Cenozoic was 4.2 cm/yr, with increasing activity at about 55 Ma (4.5 cm/yr) and 25 Ma (5.4 cm/yr) and decreased spreading rates at around 60 Ma (4 cm/yr) and 40 Ma (4.2 cm/yr).

The TYW (60–55 Ma) not only coincided with increased activity in the East Pacific Rise at 55 Ma, but also with North-Atlantic rifting and volcanism, as well as uplift of the Himalaya Range (Zachos et al., 2001). The fact that water temperatures began to rise in the Southern Ocean before the abrupt negative carbon isotope excursion of the TYW led Thomas et al. (2002) to postulate that warm water first reached this area from a distal source, which would be consistent with the idea that it was heated over a mid-ocean ridge or rise.

The LW (49–41 Ma) coincided with an increase in the spreading rate between SSA and the AP (Livermore et al., 2005; Eagles et al., 2006), between Australia and Antarctica (Royer and Sandwell, 1989), rifting on the Kerguelen Plateau and Broken Ridge (Royer and Sandwell, 1989), and the second phase of Andean uplift in Tierra del Fuego (Olivero et al., 2003).

The CW (28–24 Ma) occurred when the South Pacific spreading rate and global sea-floor production increased (Conrad and Lithgow-Bertelloni, 2007), the convergence direction of the Farrallon–Nazca Plate changed (Ghiglione and Cristallini, 2007), the Farrellon Plate broke up (Herron and Heirtzler, 1967), spreading in the West Scotia Ridge commenced (Barker, 2001), and the Bransfield Basin opened (Sell et al., 2004).

The BLW (21–15 Ma) coincided with Japan Sea spreading (Tsuchi, 1992, Fig. 4), the Columbian River volcanism (Zachos et al., 2001), and activation of the Sandwich Spreading Center east of the Scotia Plate (Vanneste et al., 2002). At this time, the Nazca–Phoenix–Antarctica triple junction also entered the South American Trench, which according to Breitsprecher and Thorkelson (2009) would have led to super-elevated thermal conditions because of mantle upwelling.

Inactivity in the mid-ocean ridges, on the other hand, probably led to lower sea surface temperatures, older, denser oceanic crust, and falling sea-levels. A decrease in sea level and increased thermohaline circulation due to expansion of the East Antarctic Ice Sheet would have enhanced the velocity of the Antarctic Circumpolar Current, causing erosion of bottom sediments and increased isolation of the AP from warm Indian Ocean water. Increased glaciation in Antarctica would also cause the ice cap to expand into the Drake Passage, where much of the ice would be grounded because of low sea levels, enhancing the strength of the Antarctic Circumpolar Current even more. Under these conditions, large areas covered under snow or ice and increased dust in the atmosphere because of compressed climate belts and higher wind velocities would reflect sunlight back to space, thus lowering temperatures further.

The period of decreased East Pacific Rise activity at 60 Ma (Conrad and Lithgow-Bertelloni, 2007) coincides with the SDC (86–60 Ma), while the YC (55–49 Ma) seems to have coincided with extension in Tierra del Fuego (Ghiglione et al., 2008).

Slowdown in the South Pacific spreading rate during the latest middle Eocene (Conrad and Lithgow-Bertelloni, 2007) may be linked to the BRC (41–28 Ma), which was also when the first influx of Pacific seawater into Atlantic Ocean was recorded (Scher and Martin, 2006). This signals the opening of the Drake Passage and initial development of the Antarctic Circumpolar Current (Barker, 2001), isolating Antarctica from warmer northern waters. Extension in central and northern Chile at 34 Ma (González, 1989; Charrier et al., 2002) can be related to oceanic plate rollback and inactivity in the East Pacific Rise. This was accompanied by a further, rapid decrease in temperatures, enhancing glaciation in Antarctica and causing the Oi-1 sea-level fall (Zachos et al., 1996).

Patriat et al. (2008), using magnetic data to calculate rotation poles from the Southwest Indian Ridge, detected a 50% decrease in spreading rate at 24.2 Ma. At this time (coinciding with the AC between 24 and 21 Ma) the Farallones Plate split to form the Nazca and Cocos Plates (Handschumacher, 1976), the spreading rate of the North Pacific Ridge decreased dramatically (Cande and Kent, 1992), and the spreading direction of the Macquarie Ridge in the Southwest Pacific also changed (Cande and Stock, 2004). This coincided with the deepening of the Drake Passage at 24 Ma (Lyle et al., 2007), as indicated by the development of bottom scouring currents in the Southern Indian Ocean (Shipboard Scientific Party, 1989). The lower sea-levels of the Mi-1 glaciation event (Miller et al., 1991; Naish et al., 2008) caused a further narrowing of the Drake Passage and higher velocities in the Antarctic Circumpolar Current (Lagabrielle et al., 2009).

Le Roux et al. (2005a,b, 2006) recorded a period of coastal subsidence in central Chile between about 15 Ma and 11 Ma, which was interrupted by uplift caused by the arrival of the Juan Fernández Ridge. If this subsidence was the result of plate rollback, it would imply ridge inactivity at least during the initial stage of the STC (15–6 Ma). Le Roux et al. (2005a) also estimated a reduction in plate velocity from about 13 cm/yr between 16 and 12 Ma (Somoza, 1998) to 6 cm/yr between 12 and 10 Ma. In the Southern Ocean, the STC was apparently accompanied by activation of the Sandwich Spreading Center (Vanneste et al., 2002), which seems to contradict a direct link between active sea-floor spreading and warming episodes. However, in this case sea-floor spreading was caused by a relatively small, new spreading center, which in fact would have enhanced the isolation of Antarctica from northern, warmer water influx by intensifying the Antarctic Circumpolar Current in a deeper Drake Passage (Barreda and Caccavari, 1992).

Renewed acceleration of the South Pacific spreading rate to about 10 cm/yr between 8 and 7 Ma (Le Roux et al., 2005a) apparently continued in the East Pacific Rise, where a spreading rate of about 9 cm/yr was recorded between 6 and 2 Ma (Rea, 1976). This was accompanied by uplift of the Bolivian Orocline (Ghosh et al., 2006) and also coincided with the MPW (6–2.8 Ma).

4.3. Role of the Drake Passage in ocean circulation

The opening of the Drake Passage played an important role in the climatic history of SSA and the AP by controlling the ocean current circulation (Katz et al., 2011). The gradual development of the Antarctic Circumpolar Current isolated Antarctica from warmer northern waters and led to continental-scale glaciation from the late middle Eocene to early Oligocene (~38–28 Ma).

The South Equatorial Current would cross the East Pacific Rise twice, once on its way to New Guinea and again on its return to SSA. During periods of increased ridge activity therefore, it could pick up enough heat to cause significant, at least subcontinental-scale changes in climate patterns. Higher water temperatures would reduce the latitudinal thermal gradient, making the South-westerly Wind Belt less intense and reducing ocean circulation. Before the opening of the Drake Passage at around 41 Ma, however, ocean current circulation in the South Pacific would have been different from that of today in that the Antarctic Circumpolar Current would not have existed. Nevertheless, circulation would still have been anti-clockwise around the Pacific Ocean. It is thus likely that the precursor of the South Equatorial Current, not coming into contact with a cold Antarctic Circumpolar Current in the southern part of its circulation pattern, would have flowed north along the western SSA seaboard as a temperate current even during periods of ridge inactivity. This might explain why temperate-warm conditions are recorded in the Southeast Pacific from at least the Oligocene until the middle Miocene. Although the effects of the BRC (41–28 Ma), in particular a large temperature drop at about 34 Ma, were felt in the Southern Ocean and Southwest Atlantic, the cold Humboldt Current only initiated at around 15 Ma (Flower and Kennett, 1993a) and intensified after the closure of the American Seaway between 4-3 Ma (Orgeira, 1990). Tectonic subsidence recorded in Tierra del Fuego could have accompanied an initial deepening of the Drake Passage during the early Oligocene, but this was only manifested clearly at around 24 Ma when the Antarctic Circumpolar Current intensified enough to lead to the cooling of Antarctica. At this time the Drake Passage would have been too wide to be closed by the Mi-1 glaciation. However, the AC (24–21 Ma) was followed soon afterwards by the BLW (21–15 Ma) associated with increased sea-floor spreading, so that the effects of the cold Humboldt Current and Falkland/Malvinas Current were not really felt along the western and eastern seaboards of SSA until the STC (15–6 Ma). Cione et al. (2011) also concluded that an equivalent of the Falkland/Malvinas Current was probably present on the southwest Atlantic shelf during the early–middle Miocene, but its temperature would have been higher than that of the present current because global temperatures were higher than today. In addition, the shelf was much wider than at present and the coastal waters were probably located further away from the Falkland/Malvinas Current.

4.4. Role of volcanic activity

In the Weddell Sea, Thomas et al. (2002) recorded a sharp negative $\delta^{13}C$ excursion during the TYW (60–55 Ma), but did not comment on the high temperature spike and cooling shown by their data prior to this event (Fig. 6). Their high-resolution oxygen isotope data show that sea-surface temperatures rose to around 16 °C, cooled to about 11 °C, and thereafter warmed rapidly to 18 °C. This was accompanied by the rapid transfer of methane (CH_4) to the ocean surface and atmosphere.

Methane, which is continuously produced in marine sediments, rises as bubbles diminishing in size and finally disappears as it dissolves in the seawater (Hovland et al., 1993). In a well-oxygenated water column, dissolved methane and sinking organic matter are oxidized by microbes. In stagnant conditions, such as might be produced by a seal of cold surface water, the Eh of the water column can be overwhelmed by the flux of sinking organic matter plus the escape of methane bubbles. The dissolved methane may thus escape oxidation in the water column and accumulate (Thomas et al., 2002). This would be enhanced in cold surface water by the solubility of CH_4 and CO_2, which at 0 °C is about double that at 20 °C. Cold surface water would thus capture these gasses as they escape upward. The subsequent rapid release of gas may take place when the water heats up, or by overturning of the water column when the surface water reaches a certain critical density and sinks into the thermocline. Heating of the descending water will decrease the solubility of CH_4 and CO_2 and cause large-scale exsolution. The rapid release of methane, being isotopically light, will result in a negative carbon isotope excursion in the geological record (Krull et al., 2000; De Wit et al., 2002).

The possibility of melting glaciers forming a cold, freshwater surface seal on the Weddell Sea during the late Paleocene can be ruled out in this case, as this would have resulted in a negative $\delta^{18}O$ excursion. Furthermore, the earliest evidence of mountain glaciers in the AP region dates from 45 to 41 Ma (Birkenmajer et al., 2005), which coexisted with *Nothofagus* forests at lower elevations. Therefore, the only other possibility seems to be rapid atmospheric cooling, which might have been brought about by explosive volcanic eruptions related to accelerated sea-floor spreading. There would have been some delay between the onset of ridge activity and subduction-related volcanism, which would explain the high temperature spike at the start of the TYW (60–55 Ma) and the rapid cooling shortly thereafter as a result of volcanic dust and sulfur compounds reflecting sunlight back into space. Modern examples of significant global cooling occurred after the eruptions of Mounts Tambora in 1815 and Pinatubo in 1991 (Stothers, 1984; Martí and Ernst, 2005). Although the individual effect of these two events lasted only a few years, accelerated subduction would probably be accompanied by many volcanic eruptions over a longer period of time.

Volcanic activity is directly related to the subduction angle, which in turn depends on the density of the oceanic crust. At the start of mid-ocean ridge activity, the oceanic crust arriving at the continental margin will be old, cold and dense, with a normal, 30–45° subduction angle. In this case partial melting of the descending plate will generate andesitic magma, which is richer in silica than basalt and more viscous, producing explosive eruptions. However, as sea-floor spreading proceeds, the oceanic crust at the subduction zone will become younger, hotter and less dense, thus descending at a lower angle. This would reduce volcanic activity (Pilger, 1981). Presently, for example, there are about 20 active volcanoes in Chile south of 33°S, where the subduction angle is normal, and only 4 in the flat-plate sector north of 28°S. Opposite the Juan Fernández Ridge between 28 and 33°S, where the subduction angle is low because of the buoyancy effect of the latter, there are currently no active volcanoes (Morales, 1984). The temperature spike in the Weddell Sea at the start of the TYW could thus indicate the first arrival of water heated over a nearby mid-ocean ridge, followed soon after by explosive volcanism and an abrupt decrease in mean air temperature. A subsequent decrease in explosive volcanic activity accompanied by continued sea-floor spreading would have allowed atmospheric temperatures to rise again because of the large-scale release of CO_2 and CH_4. Subduction activity around the AP during the early Tertiary has been recorded by several authors (e.g., McCarron and Larter, 1998), at which time an ocean spreading center collided with the trench (Scarrow et al., 1997).

4.5. Does ocean ridge activity prevent a Snowball Earth?

There is no doubt that glaciation in Antarctica played an important role in global cooling. Both the CO_2 and $\delta^{18}O$ curves in Fig. 5 show that temperatures reached a peak at about 49 Ma and from there steadily decreased into the Pleistocene. According to Livermore et al. (2005) and Eagles et al. (2006), 49 Ma was concurrent with an eightfold increase in the separation rate between SSA and the AP and the initial development of the Drake Passage. From about 34 Ma the latter was sufficiently well developed to allow intensification of the Antarctic Circumpolar Current, which isolated Antarctica from warm northerly waters and led to the formation of the East Antarctic Ice Sheet (Shackleton and Kennett, 1975; Barker and Burell, 1977; Ehrmann and Mackensen, 1992; Zachos et al., 2001). This led to accelerated global cooling, as is particularly well reflected in the CO_2 curve of Beerling and Royer (2011).

Expanding glaciation in Antarctica was accompanied by a gradual drop in sea-level, probably enhancing circulation through the Drake Passage and the further isolation of the continent. Colder surface water would have trapped more CO_2, thus preventing its release into the atmosphere, while the volume of carbonate precipitation and associated CO_2 liberation on the continental shelves would also have diminished significantly due to cooler water temperatures and less accommodation space. Furthermore, the higher pressure gradient between Antarctica and the equatorial regions would have intensified wind velocities, so that increased dust volumes in the atmosphere reflected more sunlight back to space. The latter would have been aggravated by the expansion of mountain glaciers and ice caps, in particular along the Patagonian Andes where uplift intensified after about 19 Ma (Malumián and Ramos, 1984; Vásquez et al., submitted for publication).

All of these interrelated effects and their feedbacks would tend to intensify global cooling, with apparently no intrinsic atmospheric or hydrospheric mechanisms to counteract this trend. However, accepting that ocean ridge activity may have a fundamental influence on global climates, this would indeed be able to prevent a Snowball Earth. Within the context of long-term climate deterioration since about 49 Ma, the Earth is presently in an interglacial stage, in which modern CO_2 levels (about 400 ppm) match those during the BLW (21–15 Ma) (Beerling and Royer, 2011) and slightly exceed those during the MPW (6–2.8 Ma). Interesting enough, the modern rate of sea-floor spreading on certain sectors of the East Pacific Rise, which is about 15 cm/yr (Sinton et al., 1991) also matches that during periods of increased ridge activity during the Tertiary. This suggests that the present climatic conditions might simply reflect another warming event superimposed upon a general cooling trend, and that it may be accompanied by increased subduction with more earthquakes, volcanic eruptions and sudden switches to cold periods.

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