



Atmospheric circulation over Patagonia from the Jurassic to present: a review through proxy data and climatic modelling scenarios

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This paper discusses the general atmospheric circulation over Patagonia on the basis of the principal palaeoclimate forcings: continental drift, orography, variations in the greenhouse gases in the Earth's atmosphere, evolution of the atmosphere and the oceans, changes in the orbit of the Earth, albedo feedbacks, and the land surface. These processes affect climate on time scales of millions to hundreds of thousands of years. Additionally, orbital forcing has had a major influence on climate during the Quaternary. The palaeo-atmospheric circulation of Patagonia is analyzed for the Early to Late Jurassic, the Cretaceous, the Late Palaeocene–Eocene Thermal Maximum, the Tortonian–Oligocene cooling, the Pliocene, the Quaternary including the Last Glacial Maximum, the Holocene Optimum, and the last millennium changes. Alternative palaeo-atmospheric circulations from climatic modelling scenarios through the ages are reviewed and compared with proxy data. Detailed and updated reference information on the topics analyzed is also provided. © 2011 The Linnean Society of London, *Biological Journal of the Linnean Society*, 2011, **103**, 229–249.

ADDITIONAL KEYWORDS: palaeoclimate – palaeoclimate forcings – palaeo-winds – westerlies.

Este trabajo discute la circulación general de la atmósfera en la Patagonia. Se han explicado los principales forzantes del palaeoclima: deriva continental, orografía, variaciones en los gases de invernadero de la atmósfera terrestre, evolución de la atmósfera y los océanos, cambios orbitales, retroalimentación debida al albedo y las condiciones generales de la superficie terrestre. Estos parámetros afectan el clima en escalas temporales de millones a cientos de miles de años, así como los forzantes orbitales dominaron los cambios climáticos durante el Cuaternario. Las palaeo-circulaciones de la atmósfera sobre Patagonia son analizadas para el Jurásico Temprano y Tardío, el Cretácico, el Máximo Termal del Palaeoceno Tardío– Eoceno, el enfriamiento del Tortoniano– Oligoceno, el Plioceno, y el Cuaternario incluyendo el Último Máximo Glacial (UMG), el Óptimo del Holoceno y los cambios del último milenio. Las palaeo-circulaciones atmosféricas alternativas a través del tiempo a partir de escenarios de modelos climáticos son revisadas y comparadas con proxy datos. Una extensiva lista de referencias provee información detallada y actualizada (Noviembre del 2010) sobre los tópicos analizados en esta publicación.

PALABRAS CLAVE: palaeoclima – forzantes palaeoclimáticos – palaeo-vientos – oestes.

INTRODUCTION

The climate of Patagonia is conditioned by the southern borders of the semipermanent anticyclons of the

southern Atlantic and Pacific Oceans, which extend southward to approximately 38°S to 40°S, or even farther south, as a result of the flux of the westerlies. Even without the presence of the continents, on a homogeneous Earth orbiting around the sun under conditions that vary in accordance with the theory of

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Milankovitch, the flux of the westerlies would be maintained at approximately the same latitude as today.

Despite past enormous tectonic changes throughout the last 250 Ma, Patagonia has generally been situated within the latitudes influenced by the westerlies (Kious *et al.*, 1996). This explains the low frequency of climatic changes, which are conditioned principally by the distribution of the continents. Glacial periods with permanent ice characterize the times when parts of continental masses were located in or near a polar region. During such periods, polar ice caps can maintain themselves and grow through snow and ice accumulation during repeated annual cycles. Long periods without permanent ice caps resulted instead when the continental mass remained removed from the polar regions. Forcing through atmospheric greenhouse gases is another important factor potentially responsible for climatic change. Two hundred and fifty million years ago, CO₂ concentrations were approximately 2000 p.p.m.v. (i.e. three- to eight-fold higher than today; Royer, 2006), producing warming and enhancing seasonal monsoon circulation, especially during the Triassic and Jurassic. Later CO₂ concentrations decreased, reaching present conditions during the last ice age. Starting during the Mid-Pleistocene transition, the glacial/interglacial cycles of temperature were accompanied by variations of CO₂ from 200 to 300 p.p.m.v., between the glacial and interglacial periods.

Starting in the latest Miocene, the Andean Cordillera reached elevations similar to those at present, causing important climatic changes in Patagonia. The flux of the westerlies, which advected moist air from the Pacific Ocean and brought precipitations was perturbed by the cordilleran 'wall', perpendicular to their flux. Through this effect, moisture began to discharge mainly on the western sector of the cordillera and precipitations diminished towards the east.

Clearly, marked climatic differences appeared between the glacial and interglacial periods during the last 900 000 years. This occurred mainly through the influence of the change of eccentricity of the terrestrial orbit and the nonlinear interaction of the climatic system with albedo produced by the ice caps. The Patagonian westerlies flux variation during these glacial maxima is still a matter of debate. During the interglacial periods, the changes of inclination of the axis of the Earth played an important role generating climatic changes like that of the mid-Holocene. Major warming at high latitudes during summer decreased the equator–pole temperature gradient and weakened the westerlies flux.

Climatic variations in time scales of hundreds and thousands of years were influenced by changes in solar radiation related to the number of sun spots and

volcanism. These were the principal forcings of the Medieval Warming and the Little Ice Age, during the Late Holocene. Higher frequency variations in atmospheric circulation occurring over decadal to annual timescales are also responsible for large scale climatic variability. Such large-scale variations as the El Niño/Southern Oscillation (Philander, 1990), the South Annular Mode (Marshall, 2003) or the Antarctic Oscillation (Thompson & Wallace, 2000), and the Pacific-Southamerican modes of oscillation (Mo & Paegle, 2001) influence the climate of Patagonia.

THE GENERAL ATMOSPHERIC CIRCULATION AND THE CONDITIONS OVER PATAGONIA

The tropics are characterized by a surplus of incoming radiation, whereas polar regions are characterized by an incoming radiation deficit. The intense incoming radiation in the equatorial region creates rising air masses. On a global scale, however, Earth is in a radiative equilibrium and the atmospheric general circulation function is to transport heat poleward. On a motionless Earth uniformly covered with water, the wind motion would be very simple (Fig. 1A). Around the equator, air warmed from below rises and flows toward the poles where it is cooled from below; it then sinks and flows back towards the equator, forming an atmospheric convection cell for each hemisphere.

The Earth's rotation gives rise to a 'virtual force' known as the Coriolis effect responsible for deflection of the air flow toward the right in the Northern Hemisphere and toward the left in the Southern Hemisphere. The pattern of atmospheric circulation is modified resulting in a triple cell, in part as a consequence of the Coriolis effect (Fig. 1B). The two cells driven by the ascendance at the equator, called the Hadley cells, close with a downward branch at approximately latitude 30°, forming the Subtropical High belt. As the wind flow reaches the surface, it returns to the equatorial zone to complete the cell movement. The northern boundary of these cells is marked by strong westerly winds in the upper troposphere, which are called the tropospheric jets. The rotation of the Earth shifts ocean and land eastward under the air and the flow coming from the mid-latitudes moves toward the equator. This gives rise to the easterly trade winds, characteristic of the tropical regions. In the real atmosphere, the convergence of surface winds and the resulting ascendance does not occur exactly at the equator, but in a band called the Intertropical Convergence Zone (ITCZ).

Between latitude 30° and 60° a second cell (called the Ferrel cell) exhibits a reversed flow pattern. Surface air in the Ferrel cells flows poleward and

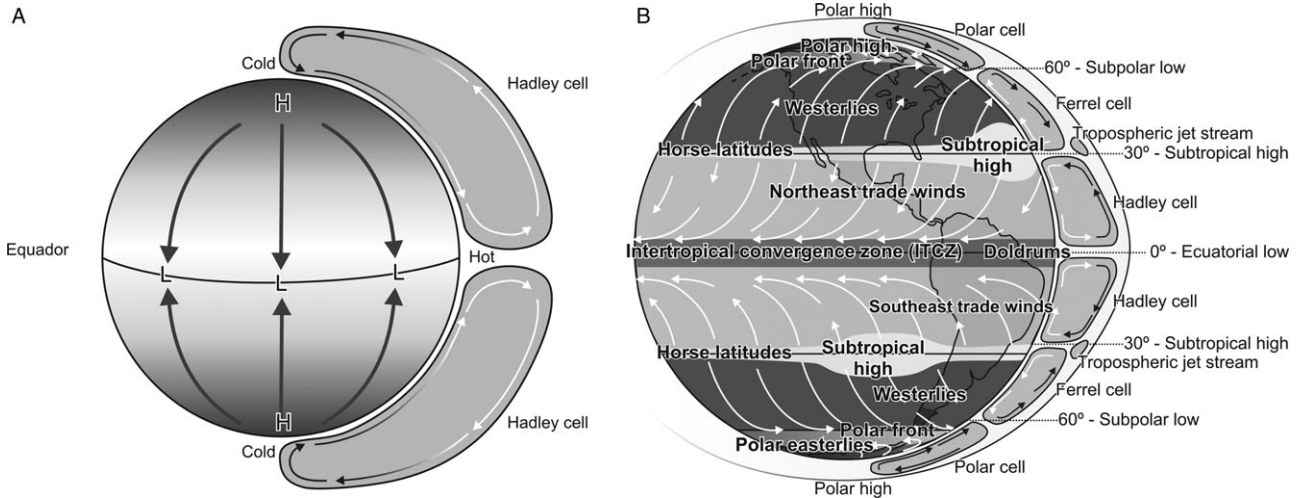


Figure 1. A, simple, single cell atmospheric convection in a nonrotating Earth. B, three cell atmospheric convection in a rotating Earth. The deflection of the winds within each cell is caused by the Coriolis force (*sensu* Lutgens & Tarbuck, 2006).

again is deflected eastward (to the right in the Northern Hemisphere and to the left in the Southern Hemisphere). The net effect is for air in this zone in each hemisphere to begin its flow from the west and move eastward. These winds are known as the westerlies; they dominate the surface extra tropical circulation. At latitude 60°, the air is still warm and moist enough to undergo convection, rising until it hits the tropopause. Then it moves toward the poles with significant cooling. It subsequently descends as a high pressure area, drifts away from the poles flowing in an easterly direction. These are the polar easterlies, which come into contact with the warmer air of the westerlies forming the Polar Front. The polar air is cold and dense, causing the westerlies to rise and flow over the Arctic and Antarctic, forming a low-pressure area. The outflow from the polar cell creates ultra-long waves, known as Rossby Waves, in the atmosphere, which are a key to determining the flow of the polar Jet Stream (Barry & Chorley, 1992)

The atmospheric circulation as a whole is strongly driven by the pole-to-equator temperature gradient (Lambeck, 1980). Wind strengths vary inversely with this temperature gradient. A low meridional temperature gradient generally reduces the zonal average wind strengths and vice versa (Holton, 1992). However, a Hadley circulation cannot provide an adequate mechanism for transporting heat polewards.

Consequently, when the meridional temperature gradient increases, the flow becomes unstable in the Hadley mode, breaking down into a number of cyclonic and anticyclonic eddies (Barry & Chorley, 1992). The strength of the westerlies is named as the

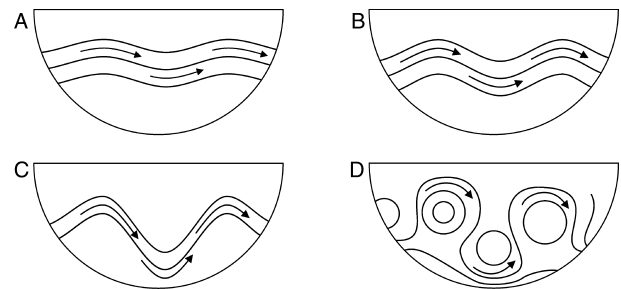


Figure 2. Westerlies circulation. A, high zonal index related strong westerlies south of their mean position. B, C, regular zonal index represents westerlies undulating with increasingly larger oscillations. D, low zonal index is associated with a complete breakup and cellular fragmentation of the zonal westerlies (*sensu* Haltiner & Martin, 1957).

zonal index. The high index (Fig. 2A), is characterized by increasing intensity of the zonal circulation at all latitudes and poleward shift of the wind intensity maxima with little north–south air mass exchange. A relatively low index (Fig. 2B, C) may also occur if the westerlies are, in the Southern Hemisphere, north of their usual latitudes. Paradoxically, such expansion of the zonal circulation pattern is associated with stronger than usual westerlies in lower latitudes. The jet expands and increases in velocity, undulating with increasingly larger oscillations. The low zonal index (Fig. 2D) is associated with a complete breakup and cellular fragmentation of the zonal westerlies, formation of stationary deep occluding cold depressions in lower mid-latitudes and deep warm blocking anticyclones at higher latitudes (Haltiner & Martin, 1957).

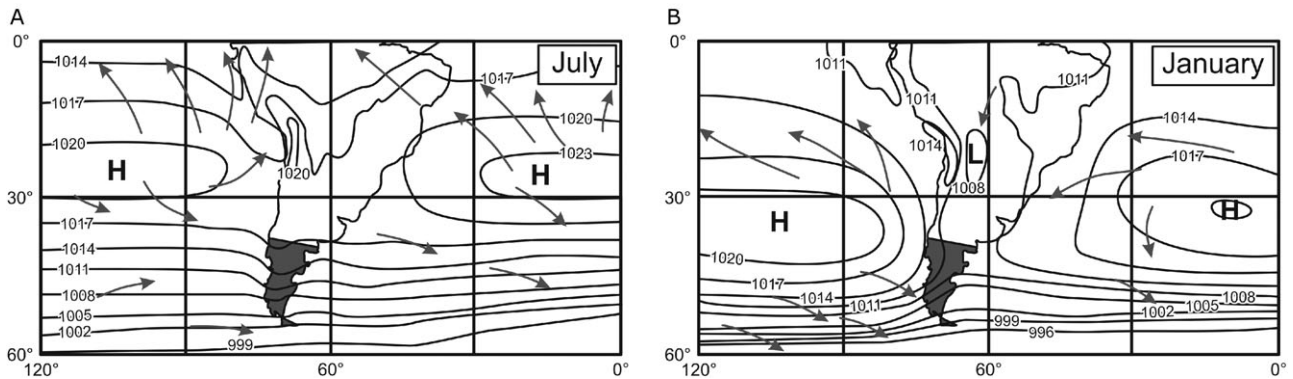


Figure 3. Middle fields surface pressure (in hPa) of the period 1961–90 for the months of (A) January (summer) and (B) July (winter) (*sensu* Lutgens & Tarbuck, 2006).

Patagonia extends from, approximately, 40° to 55°S. Its northern region is affected by the austral border of the semipermanent anticyclones of the Pacific and Atlantic oceans and its southern tip is under the effect of the low pressure belt centered at approximately 60°S (Prohaska, 1976). The climate is dominated by the westerlies, and the seasonal movement of the sun induced an atmospheric poleward circulation shift of approximately 5° during the summer (Fig. 3A, B). Furthermore, the greater annual temperature range in the subtropics and the poleward temperature decrease, results in a steeper latitudinal temperature gradient in summer than in winter. From 55°S to 65°S surface winds have maxima in the transition season.

PALAEOCLIMATE FORCINGS

Different processes are responsible for climate and atmospheric circulation forcing, and their relevance depends on the specific period analyzed and the frequency of climate change considered. Viewing the Earth's climate as a global system, Frakes (1999) describes the evolution of climate throughout the past 600 million years, from the Cambrian to the Quaternary. This study highlights the complex interactions between the carbon cycle, continental distribution, tectonics, sea level variation, ocean circulation, and temperature change, as well as other processes. Valdes (2000) provides an overview of climatic forcing mechanisms and explores their possible role in Phanerozoic climate variations.

Continental drift and orography are important for low frequency processes (i.e. those involving changes over hundreds of millions of years). Glacial periods characterized times when continental masses were located at the pole as a result of the capacity of land to support and retain ice sheets. Warm climate with the Earth free of permanent ice sheets characterized

the times when the oceans dominated the polar and sub-polar regions of the world and continental masses were located in tropical and subtropical latitudes (Crowley & North, 1999).

According to plate tectonics and continental drift theory, the supercontinent Pangaea began breaking up approximately 225 to 200 Mya, eventually fragmenting into the continents as we know them today. Although the continents have not always been in their present positions (Kious *et al.*, 1996), Patagonia has always been within in the westerlies belt area (Fig. 4). More recently Iglesia Llanos, Riccardi & Singer (2006) suggested that Patagonia shifted between the earliest and the Late Jurassic from approximately 50°S to 30°S, affecting its climate (Volkheimer *et al.*, 2008).

During the major Permo-Carboniferous glaciations (326 to 267 Mya), when Antarctica, South America and Australia drifted across or near the South Pole (Eyles, 1993; Crowell, 1999; Frakes, 1999), levels of CO₂ dropped to below 500 p.p.m. and remained at these low levels for the duration of the cold phase (Crowley & Berner, 2001; Royer, 2006). After the termination of this cold phase, CO₂ increased to ≥ 1000 p.p.m. and remained high until the Early Triassic.

The warm Mesozoic Era (230 to 65 Mya) was likely associated with high levels of CO₂ of approximately 2000 p.p.m. (Royer, 2006). The Early Jurassic to Cretaceous at 184 to 66.5 Mya (Frakes, 1999), with cool pulses each lasting only < 3 Mya, was characterized by very high CO₂ levels (6000 p.p.m.). Thereafter, CO₂ levels oscillated between very high (approximately 2000 p.p.m.) and low (500 p.p.m.) values (Royer, 2006).

A major expansion of Antarctic glaciations starting approximately 40 to 35 Mya was likely a response, in part, to declining atmospheric CO₂ levels from their peak in the Cretaceous (approximately 100 Mya)

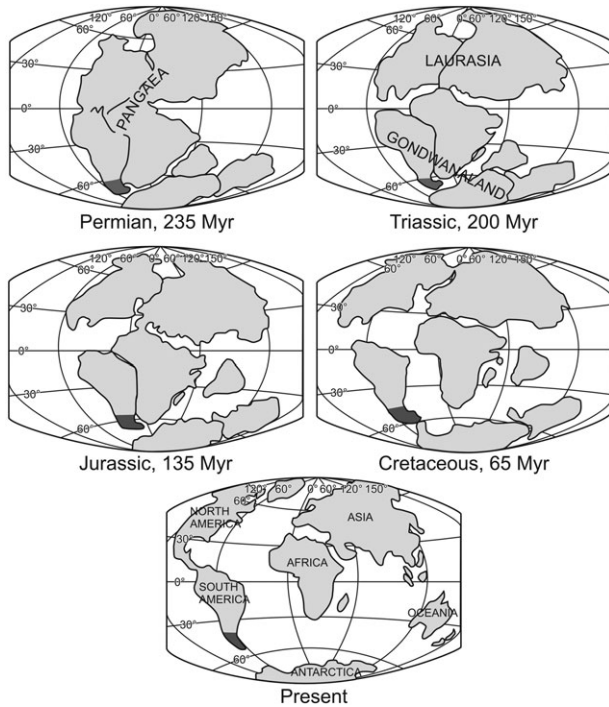


Figure 4. Patagonia: location (dark grey area) in the palaeo-continental maps according to the continental drift theory (*sensu* Kious *et al.*, 1996).

(DeConto & Pollard, 2003). During the Cenozoic Ice Age, which began in the late Eocene and continues to the present, the CO_2 concentrations declined to low values (< 500 p.p.m.) (Royer, 2006).

The processes that affect climate in time scales from millions to hundreds of thousands of years include changes in the orbit of the earth, the evolution of the atmosphere, albedo feedbacks, and changes in land surface. Orbital forcing dominates the climate changes during the Quaternary, producing approximately 100 kya oscillations. Following the temperature changes of the Quaternary, CO_2 levels varied by volume between 180–210 p.p.m. during glaciations, increasing to 280–300 p.p.m. during warmer interglacial periods (Siegenthaler *et al.*, 2005).

In shorter time scales (i.e. time scales ranging from thousands to hundreds of years), the climate is influenced by solar variability, the concentration of green house gases and by volcanic activity. The different climate forcings involved in the palaeoclimate changes are widely discussed by Huber, MacLeod & Wing (2000).

PALAEO-ATMOSPHERIC CIRCULATION

Climate models provide a framework within which existing data can be interpreted and hypotheses tested. They are particularly valuable for regions

with limited data and provide ways to help interpret local palaeoclimate records. Palaeo-atmospheric circulations are normally inferred using three-dimensional general circulation models (GCMs) of the ocean and atmosphere. These models predict the fluid flow on a rotating sphere heated by solar radiation (Crowley & North, 1999; Trenberth, 1992; Huber *et al.*, 2000; Poulsen, 2008, among others). The ages indicated for each cited period correspond to those presented in the US Geological Survey Geologic Names Committee (2010), Divisions of geological time—major chronostratigraphic and geochronologic units.

EARLY (APPROXIMATELY 99 MYA) TO LATE (APPROXIMATELY 161 TO 145 MYA) JURASSIC

Pangaea, the largest landmass in the Earth's history, was nearly bisected by the equator during the Mesozoic Era. This single continent stretched latitudinally across every part of the zonal atmospheric circulation, thereby producing an extraordinary effect on global palaeoclimate (Dubiel *et al.*, 1991; Valdes, 1993; Sellwood, Valdes & Price, 2000; Sellwood & Valdes, 2006). The supercontinent generated a mega-monsoonal atmospheric circulation in full swing during the Early and Middle Jurassic. Such atmospheric circulation led to extreme seasonality (Kutzbach & Gallimore, 1989).

In the Southern Hemisphere, latitudes north of 30°S were influenced by the ITCZ during summer, whereas southwestern South America was dominated by the movement of the Southern Panthalassa ocean semipermanent anticyclone, which shifted to the south during summer, affecting the southwest margins of Gondwana – Patagonia (Fig. 5A).

Moore *et al.* (1992a, b) used a GCM to obtain two Kimmeridgian/Tithonian (approximately 154.7 to 145.6 Mya) palaeoclimate seasonal simulations, with geologically inferred palaeotopography: one simulation used a CO_2 concentration of 280 p.p.m. (pre-industrial level) and the other used 1120 p.p.m.. Increasing the CO_2 four-fold warms virtually the entire planet. The greatest warming occurs over the higher latitude oceans and the least over the equatorial and subtropical regions.

Different GCMs simulations (Moore *et al.*, 1992a, b; Chandler, Rind & Ruedy, 1992; Valdes & Sellwood, 1992; Kutzbach *et al.*, 1989) showed strong seasonal alteration of summer monsoon lows and winter monsoon highs.

During the summer, monsoon lows were centered poleward of the western Tethys Sea, near 35°S. This location is just east of the region of summer maximum temperature. Over the oceans, in middle latitudes, the Panthalassa subtropical high, which was centered near 50°S, produced surface winds that

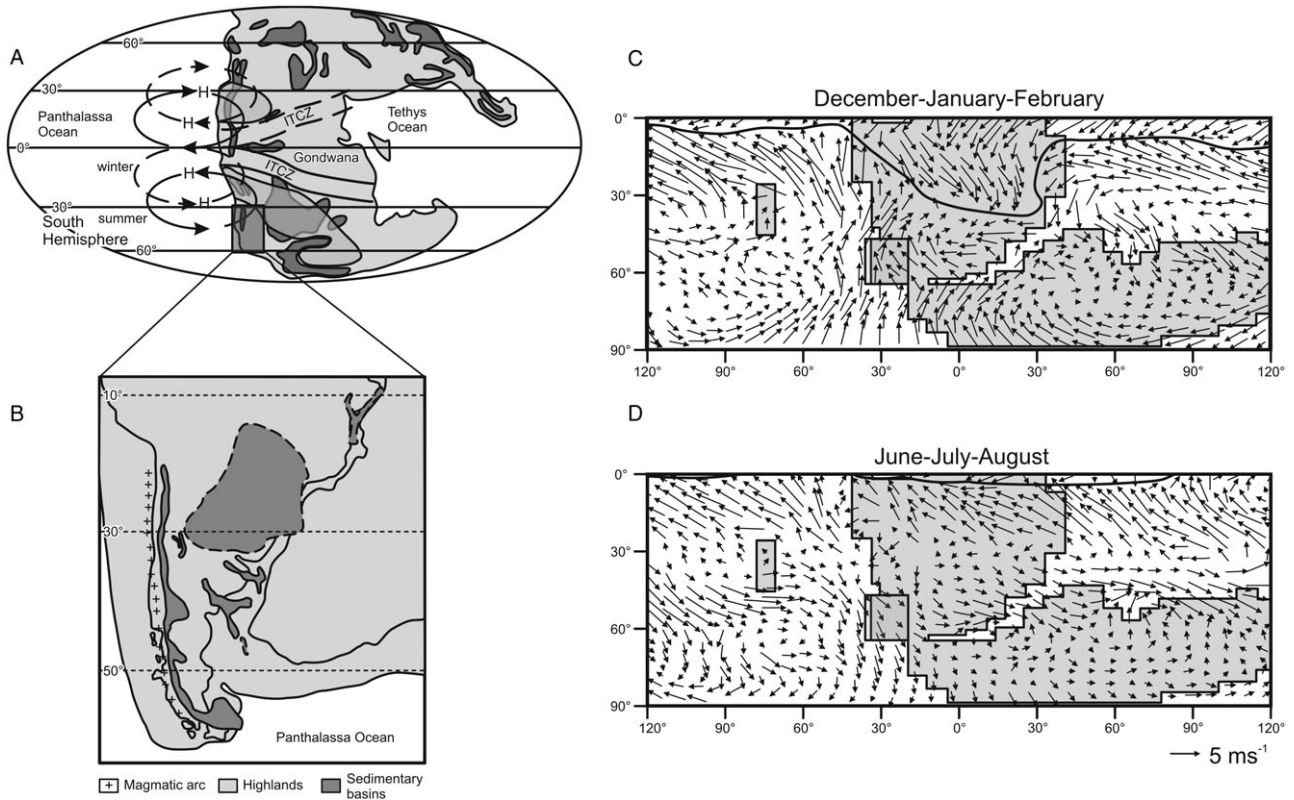


Figure 5. Latest Jurassic–earliest Cretaceous (A) Palaeolatitudes and palaeogeography, and Intertropical convergence Zone (ITCZ) with Panthalassa Ocean Semipermanent Anticyclone during Southern Hemisphere summer (solid line) and winter (dashed line) (adapted from Loope, Rowe & Joeckel, 2001). B, palaeogeographical map of the southwestern Gondwana (adapted from Scherer & Goldberg, 2007). C, D, maps of Kimmeridgian/Tithonian surface wind vectors, 1120 p.p.m. CO_2 simulation; (C) December, January, and February and (D) June, July, and August (*sensu* Moore *et al.*, 1992a).

blew from the southeast over Patagonia (Fig. 5C). These winds correlated with lower precipitation than evaporation. In autumn, the Polar highs were located at 60–70°S, poleward of the Tethys Sea (Fig. 5).

The wind pattern over Patagonia during winter was opposite to that during summer (Fig. 5C, D, arrow directions within squares). During winter the Panthalassa subtropical high was centered at near 25°S. South of it, a subpolar low, centered near 60°S, dominated the south Panthalassa Ocean. These high and low pressure systems along with the continental winter monsoon high, which was located poleward of the western Tethys Sea and was centered near 40°S, were responsible for the winds from the northwest (Fig. 5D) advecting wet and warm air over Patagonia. Furthermore, in the middle troposphere the axes of the Southern Hemisphere mid-latitude storm track is located over 60°S. These factors are related to high positive net precipitation where precipitation exceeds evaporation. The model simulated precipitation rate in the region is $\geq 5 \text{ mm day}^{-1}$ (Valdes, 1993) in accordance with the coal localities that are proxies for wet environments (Scotese, 2010).

Moore *et al.* (1992a) conclude that the Jurassic warmth evidenced by climate proxies may be explained by elevated atmospheric CO_2 . By contrast, Chandler *et al.* (1992) suggested the simulation with specified, warm SSTs was in energy balance without high atmospheric CO_2 , implying that a warm Jurassic climate could have been the product of enhanced poleward heat transport through the ocean. In other words, two feedback mechanisms are presumed to be primarily responsible for the warm climate over Patagonia during the Jurassic: the elimination of sea and land ice that resulted from the warm polar sea surface temperatures (SSTs) and the equatorward shift of Antarctica resulting in a decrease in surface albedo.

CRETACEOUS (APPROXIMATELY 145 TO 65 MYA)

The fragmentation of Pangea began during the Jurassic. In the Late Jurassic to Early Cretaceous, the separation between Gondwana and Laurasia was well in progress, and the South Atlantic Ocean had already developed between those continents. India separated from Madagascar and raced northward on

a collision course with Eurasia. North America was connected to Europe, and Australia was still joined to Antarctica. By the Late Cretaceous, the oceans had widened, and India approached the southern margin of Asia (Scotese, 2010).

The Early Cretaceous was a mild 'Ice House' world. There was snow and ice during the winter seasons, and cool temperate forests covered the Polar Regions. The Late Cretaceous was instead characterized by super greenhouse intervals of global warmth with ice-free continents. Globally averaged surface temperatures were 6–14 °C higher than at present (Barron, 1983) and the temperature gradient between the poles and the equator was lower than today, that is approximately 50 °C in the Northern Hemisphere and 90 °C in the Southern Hemisphere. The difference is largely a result of adiabatic cooling reflecting the elevation of Antarctica. Frakes (1999) summarized the data on estimates of Cretaceous sea surface and terrestrial temperatures.

There are four differing assumptions concerning Cretaceous temperatures and meridional gradients: (1) tropical SSTs were the same as today, although polar temperatures were warmer (5–8 °C) except when ice was present (0 to –5 °C); (2) the tropics were significantly cooler and mid-latitudes warmer than today; (3) tropical SSTs were 32–34 °C, with polar regions 10–18 °C; (4) tropical SSTs were approximately 42 °C and polar temperatures > 18 °C. Furthermore, different hypothesis can explain the drastic warming and equable high latitudes during super-greenhouse intervals of the Cretaceous and early Cenozoic. On the basis of coupled ocean–atmosphere model simulations of the middle Cretaceous, Poulsen, Gendaszek & Jacob (2003) hypothesized that the formation of an Atlantic gateway could have contributed to the Cretaceous thermal maximum. Kump & Pollard (2008) GCM simulation of Middle Cretaceous using $4 \times \text{CO}_2$, from preindustrial atmospheric level, failed to produce the extreme high-latitude warmth implied by temperature proxy data. However, simulations with the combined increases in cloud droplet radii, which mainly affect cloud optical depth, and precipitation efficiency, resulted in a reduction in global cloud cover from 64% to 55% with optically thinner clouds that reduced planetary albedo from 0.30 to 0.24. The ensuing warming was dramatic, both in the tropics and in high latitudes, where warming was augmented by surface albedo feedback of almost vanishing snow and sea-ice cover. Otto-Bliesner, Brady & Shields (2002) altered the models by the inclusion of high-latitude forest thus changing the palaeogeography. These low-albedo forests warmed the high-latitude continents, which then transferred more heat to the high-latitude oceans, impeding sea-ice formation and warming coastal regions.

Floegel (2002) obtained scenarios for the Turonian (approximately 93.5 to 89.3 Mya) palaeogeography by GCM simulations with different orbital configurations, and 1882 p.p.m. CO_2 ($5 \times$ year 2000 CO_2 , $7 \times$ preindustrial CO_2) concentrations. The simulated atmospheric circulation resulted in a much more complex circulation than that observed today. In this model, the tropical easterlies remain constant and strong throughout the year. However, at higher latitudes, the circulation varies with the seasons. Because of the absence of polar highs during the winter, strong westerly wind belts develop between 50°S and the high pressure zones at 30°S. Another important difference in the Floegel's scenarios lies in the strong trade winds, which developed during each hemispheric winter.

Floegel's (2002) model, in the Southern Hemisphere summer (December to February), shows the polar region under the influence of a low atmospheric pressure system and therefore the westerly winds are weak, variable, and may even reverse direction. Also, the subtropical to polar frontal systems would not exist (Fig. 6). This would result in disruption of the mid and high latitude wind systems. In the winter (June to August), the southern polar region is under the influence of a high pressure system and the westerlies are well developed (Hay, Flögel & Söding, 2005 & Hay, 2008)

The Maastrichtian (70.6 to 65.5 Mya) palaeo-wind scenario in Figure 7 was obtained by Bush (1997) using an atmospheric–oceanic GCM, dynamically and thermodynamically coupled, with four times the present-day value of atmospheric CO_2 , as indicative of Cretaceous levels. Even with these relatively new results, it is necessary to be cautious with the adopted palaeogeographical reconstruction for the Maastrichtian (Ziegler, Scotese & Barrett, 1983), used in the model, with the Drake Passage (DP) opened.

Some GCM simulations of the Cretaceous super-greenhouse also considered the DP opened as boundary conditions (Poulsen, Pollard & White, 2007; Sewall *et al.*, 2007; Kump & Pollard, 2008; Zhou *et al.*, 2008), whereas others considered DP closed (Haywood, Valdes & Markwick, 2004). For example, Bush & Philander (1997) had estimated a later age for the opening of DP, in the range 49 to 17 Mya; for more details and additional citations, see Cavalotto, Violante & Hernández-Molina (2011).

The earliest connection between the Pacific and Atlantic oceans at DP is controversial but important because the gateway opening probably had a profound effect on global circulation and climate (Cavalotto *et al.*, 2011). Sijp & England (2004) studied the DP influence on climate by means of three main model simulation experiments, set up identically with the exception of bathymetry. They kept the DP closed by

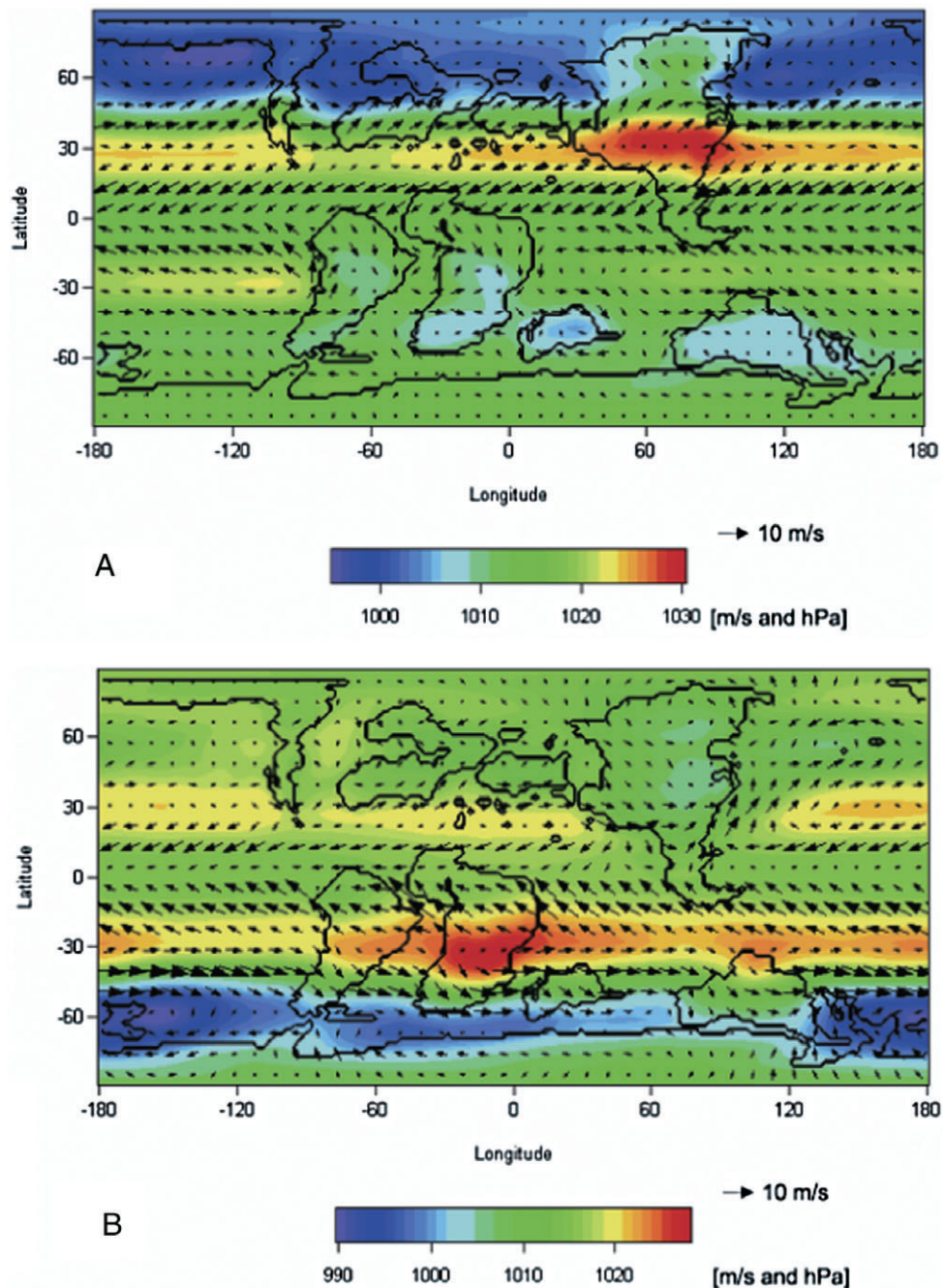


Figure 6. Turonian (93.5 Mya) wind speed and pressure at sea-level for (A) December, January, February and (B) June, July, August (m/s and hPa) (© Sascha Floegel, IFM-GEOMAR, Kiel, Germany, pers. comm., *sensu* Floegel, 2002.).

a land bridge between the Antarctic Peninsula and South America, DP open to a maximum depth of 690 m and DP open at its present day depth, which was modelled as an uninterrupted throughflow at depth of c. 2316 m.

The climate with DP closed is characterized by warmer Southern Hemisphere surface air tempera-

ture and little Antarctic ice. An increase in Antarctic sea ice and a cooling the Southern Hemisphere takes place on opening the DP to a shallow depth of 690 m. On fully opening the DP, the climate is mostly similar in the Southern Hemisphere to DP at 690 m.

The Bush's (1997) simulated pattern of annual mean wind stress (Fig. 7) resulted in a pattern similar to

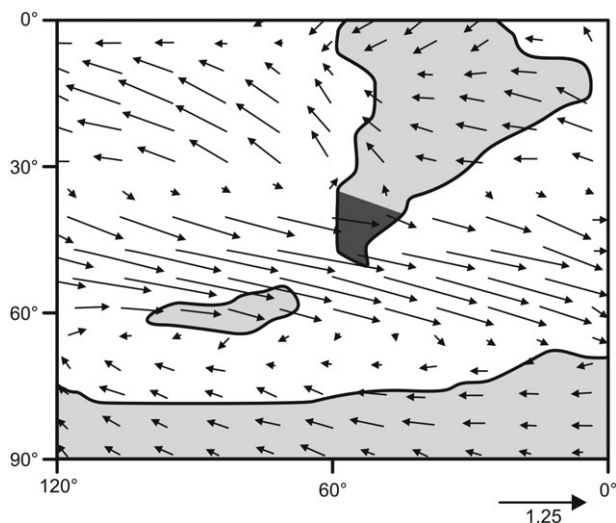


Figure 7. Annual mean wind stress vectors (dynes cm^{-2}) for the Late Cretaceous. Arrows are scaled as displayed in the bottom right (*sensu* Bush, 1997).

that of the present day with only some changes as a consequence of the modified continental–ocean distribution. The annual mean fields show the tropical–subtropical region circulation characterized by the ITCZ and the easterly winds occupying the latitudinal band from 30°S to 30°N. The Southern Hemisphere Westerlies occupy the latitudinal band from 40° to 70°S and the mid-latitude tropospheric jets dominate the atmospheric circulation over Patagonia (Fig. 7). The simulation shows no poleward shift of the mid-latitude westerlies in response to an ice-free planet.

In the Bush & Philander's (1997) simulation, using a coupled atmosphere–ocean GCM with quadrupled CO_2 , the annual mean Hadley circulation indicates a general reduction in strength of the Southern Hemisphere cells in the Cretaceous, with the south equatorial Hadley cell weakening by 20%. Subsidence in the high southern latitudes decreased dramatically in the Cretaceous in response to the elimination of the Antarctic ice sheet, which, in the present climate, induces strong subsidence poleward of 75°S. The strengths of the middle and high-latitude cells decrease as a consequence. In addition, the separation of the North and South American continents allowed stronger northeasterly trades flowing from the Tethys basin into the Pacific basin, whereas the strength of the south-easterly winds off the western coast of South America remains approximately the same.

Both Bush & Philander's (1997) and Hotinski & Toggweiler's (2003) modelling studies suggest the possibility of intensified surface circulation. The stronger upper tropospheric westerlies and stronger lower tropospheric easterlies over the tropical

Pacific suggest a 'permanent El Niño' state, as Davies (2006) found by analyzing laminated sediment. The model results showed the global precipitation approximately 10% higher than at present. The mean annual temperatures increased and the amplitude of seasonal cycle in near-surface temperatures diminished, consequently precluding the presence of year-round snow or ice in the simulation. In high latitudes, however, there are regions that seasonally drop below freezing.

LATE PALAEOCENE–EARLY EOCENE (APPROXIMATELY 57 TO 53 MYA): THERMAL MAXIMUM

By the Late Cretaceous, the oceans had widened, and India approached the southern margin of Asia. Australia, which was attached to Antarctica, began to move rapidly northward. In the Eocene, 55 to 50 Mya, India began to collide with Asia forming the Tibetan plateau and the Himalayas (Scotese, 2010). The interval from the Late Palaeocene to the Early Eocene was one of the warmest time periods in the Earth history with small annual range. Palaeontological proxy data such as coals in the Antarctic Peninsula and Patagonia, where palm trees grew, suggest a 'greenhouse' world with ice free high latitudes (Zachos *et al.*, 1993). The Late Palaeocene–Eocene Thermal Maximum (LPTM) was a dramatic, short-term (approximately 170 kya) global warming event that occurred at approximately 55 Mya (Röhl *et al.*, 2007). The LPTM proxy data predict warm (6–12 °C) deep ocean and polar temperatures, near-modern (23–27 °C) tropical SSTs (Crowley & Zachos, 2000), and higher-than-modern greenhouse gas levels (pCO_2 400–4000 p.p.m.) (Pearson & Palmer, 2000). No significant ice accumulation existed at high latitudes in either hemisphere. The ice-free state during LPTM was characterized by a much higher annual-mean temperature and a greatly reduced seasonal cycle along with a lower equator-to-pole surface temperature gradient than is known for any other period in the Cenozoic for which data exist (Abbot & Tziperman, 2008). There is substantial evidence that a weakening of the gradient occurred rapidly and synchronously with a perturbation to the global carbon cycle (Huber & Sloan, 1999). Zeebe, Zachos & Dickens (2009) also suggest that, in addition to the direct CO_2 forcing, other hitherto unknown processes and/or feedbacks that must have caused a substantial portion of the warming during the Palaeocene–Eocene Thermal Maximum. The LPTM rapid and short lived change in the pole to equator temperature gradient probably involves a rapid atmospheric and ocean circulation reorganization.

In the Huber & Sloan (1999) simulation of the LPTM climate conditions, the surface winds for

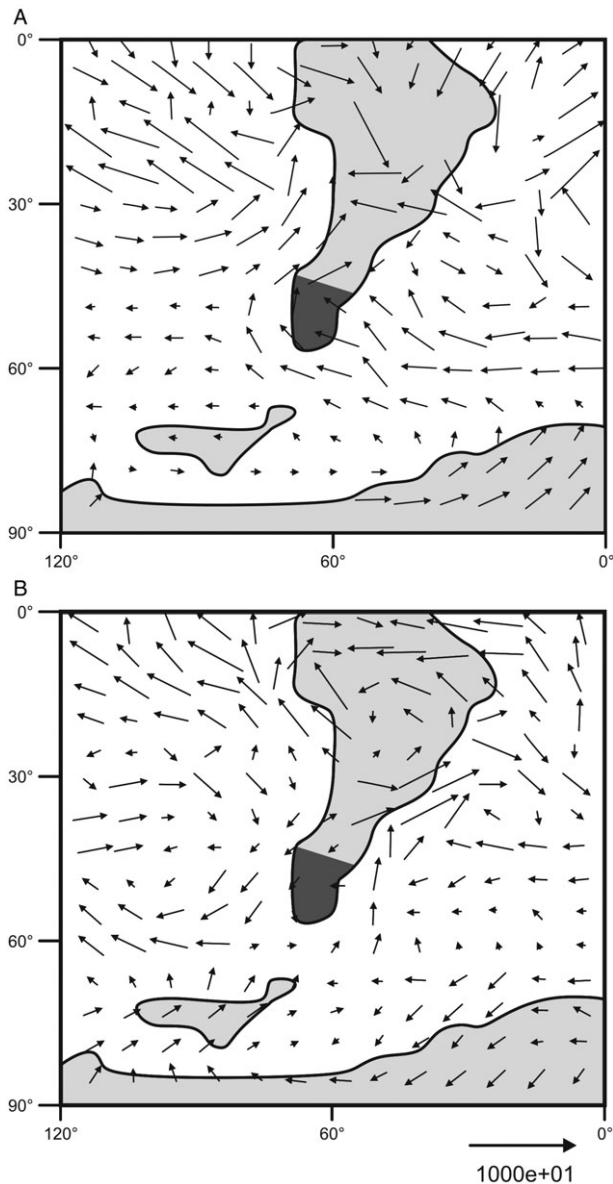


Figure 8. The seasonal maps of the surface wind vectors for the Late Palaeocene–Eocene Thermal Maximum (LPTM) case for (A) January and (B) July. Vector length indicates the magnitude of wind speed in proportion to the sample vector (*sensu* Huber & Sloan, 1999).

January (Fig. 8A) presented mid-high latitude easterlies centered on 50° to 60°S and a weak westerly wind belt around 30°S. The presence of mid-high latitude easterlies is not the result of an expansion of the tropical easterlies so much as an expansion of the polar easterlies. This wind scenario, characterized by relatively strong easterlies over Patagonia and a weakened air system over the Pacific Ocean likely generated air mass advection from the Atlantic inducing convection and precipitation over land. Such

conditions are consistent with the suggested rich subtropical vegetation that existed over a large portion of Eocene Patagonia with the presence of megathermal families such as palms, other taxa with broader climatic requirements such as conifers, cycads, and Ginkgoales (Wilf *et al.*, 2005; Iglesias, Artabe & Morel, 2011; Quattrocchio *et al.*, 2011), as well as the coals in the east Patagonia (Scotese, 2010).

The strong high latitude precipitation in the model scenarios (not shown) was a reflection of the warm high latitude SSTs, as a result of the fact that the removed seasonal sea-ice from the polar seas enhanced local evaporation and precipitation. The reversal of the ‘polar cell’ caused by the deep high latitude convection is a fundamental disruption of the normal atmospheric general circulation. During winter time (Fig. 8B), the winds display the same easterlies pattern over Patagonia but weaker than in summer.

TORTONIAN (APPROXIMATELY 11 TO 7 MYA) AFTER THE EOCENE/OLIGOCENE COOLING

The Late Miocene represents a transitional phase between the Eocene greenhouse climate and the Quaternary icehouse situation. Ice had begun accumulating at the South Pole during the Late Eocene (approximately 37.2 to 33.8 Mya) culminating in the Pleistocene ice age. The widespread glaciation of Antarctica and the associated shift toward colder temperatures near the Eocene/Oligocene boundary (approximately 34 Mya) represents one of the most fundamental reorganizations of the global climate system recognized in the geological record (DeConto & Pollard, 2003). Concordantly, the atmospheric circulation pattern also changed (Broccoli & Manabe, 1992) and the climatic conditions in the region of Patagonia were strongly affected.

Several mechanisms were connected with Oligocene cooling, including changes in continental distribution (Barron, 1985), plateau uplift (Kutzbach *et al.*, 1989; Hay *et al.*, 2002), oceanic gateways (Sijp, England & Toggweiler, 2009), and the decrease of atmospheric CO₂ (DeConto & Pollard, 2003). Scher & Martin (2006) suggest that the Drake Passage opened before the Tasmanian Gateway, implying the late Eocene establishment of a complete circum–Antarctic pathway. However, note that Cavallo *et al.* (2011) considered that the Tasman path opened before. The Antarctic Circumpolar Current thermally isolated Antarctica, leading to the growth of the Antarctic Ice Sheet. Toggweiler & Bjornsson (2000) suggest that this opening reduced the poleward heat transport in the high-latitude Southern Hemisphere cooling high-latitude surface temperatures by several degrees (in the range 0.8–4 °C).

The Late Miocene Tortonian (approximately 11 to 7 Mya) period, was characterized by intensive Antarctic glaciations and the beginning of glaciations in the North Atlantic region. Steppuhn *et al.* (2006) run an atmospheric general circulation model coupled to a simple mixed-layer ocean model representing only the uppermost layer of the ocean to calculate the palaeo-SST independently. The boundary conditions and the altitudes of the mountain chains had to be adapted to the Tortonian time interval. The height of the Tibetan Plateau was assumed to have reached approximately 50% of its present height. The Control run represents the present-day situation. Model simulations indicate that differences between the Tortonian and Control runs are less significant in the Southern than in the Northern Hemisphere. Cooling occurs over Antarctica appearing more intense during summer than during winter. Thus, the Tortonian seasonality is reduced at high latitudes. Well developed semi-permanent anticyclones occurred over oceans at subtropical latitudes affecting the equatorial boundary of Patagonia.

In the south Pacific anticyclone, the Tortonian run (Fig. 9A) is stronger than the Control run, as can be seen in the differences between Tortonian and Control run (Fig. 9B) and, together with the atmospheric circulation over the Pacific Ocean, they suggest a permanent El-Niño state. The latitudes of the Patagonian region are dominated by the westerlies (Fig. 9A). The easterly arrows over the high-midlatitudes, in the difference between the Tortonian and the present Control run (Fig. 9B), means less westerlies stress than at present.

The uplift of mountain ranges and plateaus during the Neogene and the Cenozoic has been considered by some geologists to be either the direct or the indirect cause of the 'climatic deterioration' leading to the late Neogene glaciations (Ruddiman *et al.*, 1997). The South American Andes, formed mainly through crustal thickening associated with Cenozoic subduction and convergence between the Nazca Plate and the South American Plate (Folguera *et al.*, 2011).

The significant surface uplift of the northern Andean plateau, on the order of 2.5 ± 1 km, occurred between approximately 10.3 and 6.4 Mya. Bershaw *et al.* (2010) suggested that, by approximately 8 Mya in the northern Altiplano and 3.6 Mya in the southern Altiplano, both regions had reached high elevation and established a latitudinal rainfall gradient similar to modern conditions. Ramos & Ghiglione (2008) broadly described the Patagonian Andes uplift, which occurred during the Neogene.

The large influence of the Andes on regional climate has led to the speculation that surface uplift caused past climate changes. The uplift of the

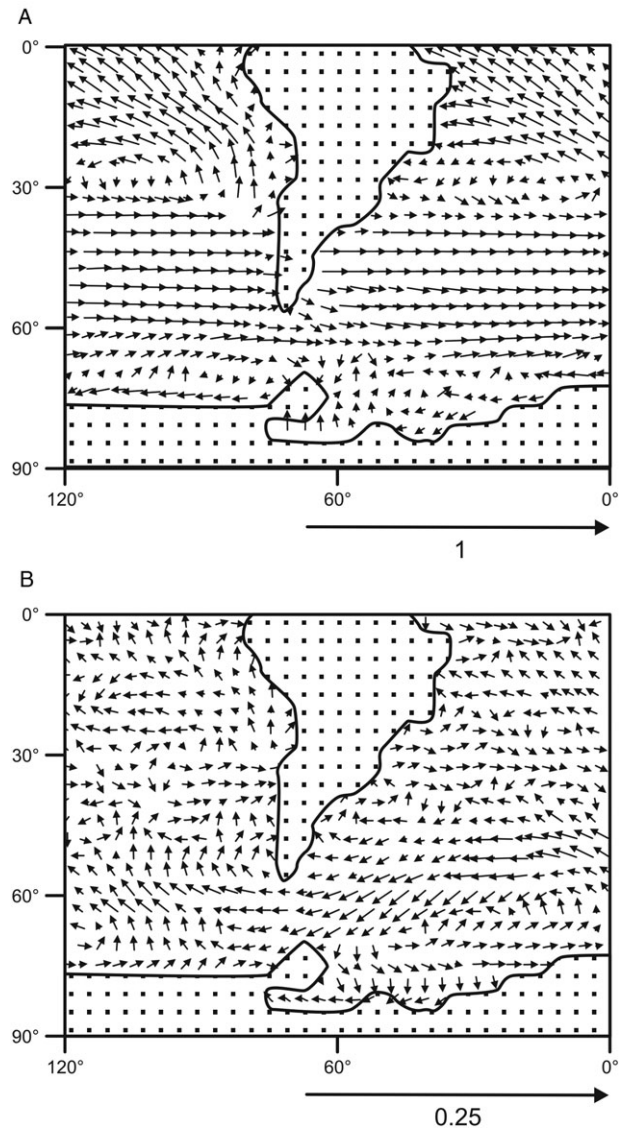


Figure 9. The mean annual wind stress field corresponding to the (A) Tortonian run and (B) the difference between the Tortonian and the present Control run. The magnitude of the reference arrow is 1 m s^{-1} for (A) and 0.25 m s^{-1} for (B) (*sensu* Steppuhn *et al.*, 2006).

southern Patagonian Andes forms a pronounced topographic barrier to atmospheric circulation in the Southern Hemisphere Westerlies, and caused one of the most drastic orographic rain shadows on Earth. Carbon isotope data from the pedogenic carbonate samples demonstrate that this led to strong desertification in the eastern foreland and, presumably, strongly increased precipitation rates on the windward western side of the mountains (Blisniuk *et al.*, 2006; Aragón *et al.*, 2011). Poulsen, Ehlers & Insel (2010) studied the influence of the Southern Andes uplift on the South American climate, precipitation

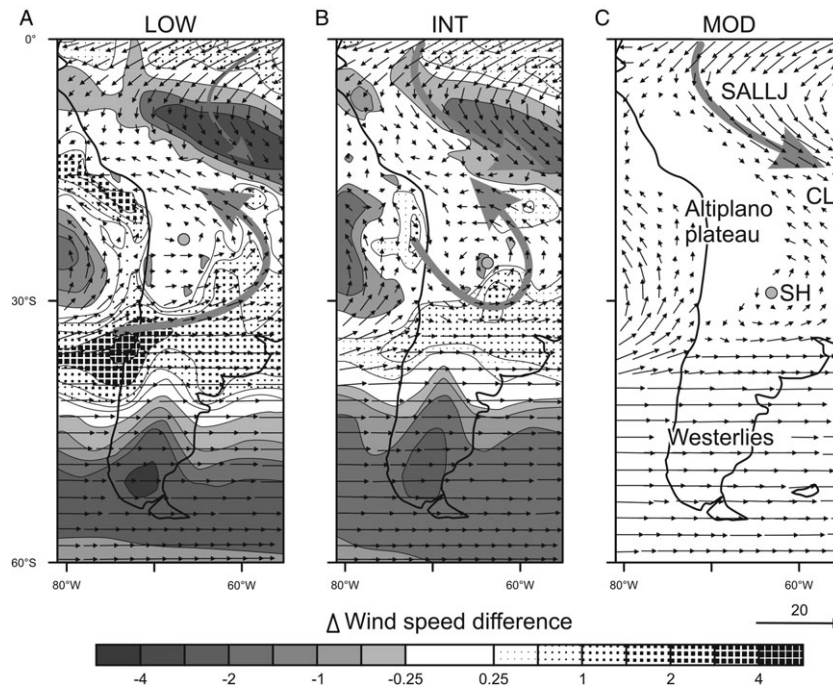


Figure 10. Simulated summer low-level (800-mbar) circulation for the three uplift scenarios: (A) low elevation (LOW), (B) intermediate elevation (INT) and (C) modern elevation (MOD). Vectors are missing in (C) where the flow intersects high topography. The difference in 800-mb wind speed (m s^{-1}) from the control (MOD) case (LOW–MOD; INT–MOD) with positive (grey shaded) values represent faster flow and negative (dotted) values represent slower flow. The circle indicates the center of the subtropical high; arrows indicate low-level circulation and American low-level jet (SALLJ) (*sensu* Poulsen *et al.*, 2010).

and atmospheric circulation, using an isotope-tracking atmospheric general circulation model. The resulting low atmospheric perturbation from three GCM experiments with prescriptions of Andean elevations corresponds to Andean elevations of 250 m (LOW) (Fig. 10A), intermediate height of 50% of contemporary altitude (Fig. 10B), and the modern elevations (Fig. 10C). Surface uplift modifies the low-level circulation. The Andes block the zonal atmospheric flow and intensify the low-level flow and vapor transport via the South American low-level jet which intensifies. Early uplift of the Andes also leads to southeastward displacement of the subtropical high, away from the growing plateau, and the mid-latitude westerlies migrate southward and become stronger.

PLIOCENE (APPROXIMATELY 5.3 TO 2.5 MYA)

The Pliocene spans a time frame from approximately 5.3 to 2.5 Mya. The epoch incorporates the time interval in which the Earth experienced a transition from relatively warm climates to the prevailing cooler climates of the Pleistocene (e.g. Lisiecki & Raymo, 2007).

The Middle Pliocene is the most recent interval of geological time during which global climate was

significantly warmer than at present took place (Crowley, 1996). By contrast to other Pre-Quaternary intervals, the boundary conditions for the Pliocene are fairly well-known. The Mid-Pliocene is recent enough that the continents and ocean basins had almost reached their present geographical configuration. The average of the warmest times during the Mid-Pliocene suggests a globally warmer world, in which atmospheric CO_2 concentrations were likely higher than the pre-industrial value (Jansen *et al.*, 2007).

The Early to Middle Pliocene temperature structure of the equatorial Pacific Ocean would have been more zonally uniform than it normally is today and more like the modern Pacific Ocean in an El Niño state (Molnar & Cane, 2002). According to Philander & Fedorov (2003), approximately 3 Mya, the thermocline was sufficiently shallow for the winds to bring cold water from below the thermocline to the surface in certain upwelling regions. Feedback mechanisms could account for the amplification of the Earth's response to periodic variations in obliquity of the Earth's axis (at a period of 41 kyr) and could involve ocean–atmosphere interactions of the type associated with El Niño and also mechanisms by which high-latitude surface conditions can influence the depth of the tropical thermocline. Therefore, the Pliocene

paradox is the persistence of El Niño conditions, which was the major factor in the warmth of the Early Pliocene. Fedorov *et al.* (2006) found that the El Niño-like permanent conditions contributed to global warming, first by causing the disappearance of stratus clouds from the eastern equatorial Pacific Ocean, thus lowering the planetary albedo, and second by increasing the atmospheric concentration of water vapor, a powerful greenhouse gas. Brierley *et al.* (2009) reconstructed the latitudinal distribution of SST approximately 4 Mya, during the early Pliocene. The reconstruction shows that the meridional temperature gradient between the equator and subtropics was greatly reduced, implying a vast poleward expansion of the oceanic tropical warm pool with enormous impacts on the Pliocene climate, including a slowdown of the atmospheric Hadley circulation.

Simulated 850 hPa wind fields (Jiang *et al.* (2005) show weakening Middle Pliocene Westerlies during summer, whereas, over the Atlantic and Pacific oceans, semi-permanent anticyclones appears to have been strengthened and poleward shifted. During winter in the south Pacific Ocean, the westerlies are strengthened and only slightly reinforced over Patagonia, compared with today.

THE QUATERNARY SINCE 2.5 MYA AND THE LAST GLACIAL MAXIMUM (LGM; APPROXIMATELY 20 KYR BP)

The warm equable climates were, after the Pliocene, replaced by recurring ice ages. Raymo *et al.* (1996) attempted to explain this replacement appealing to changes in CO₂ in the atmosphere. Estimates of partial pressures of CO₂ in the atmosphere of the past, however, show only modest changes during the past 50 Mya (Pearson & Palmer, 2000). The additional cause, suggested by Sloan, Crowley & Pollard (1996), is the reduced heat transport by the Atlantic Ocean. Other forcing mechanisms included the fact that the obliquity cycle became more prominent approximately 3 Mya. Haug & Tiedemann (1998) propose that the progressive increase in obliquity amplitudes was the final trigger for amplification and continuation of the long-term expansion of the Northern Hemisphere ice sheets after the necessary preconditions were met 4.6 to 3.6 Mya by formation of the Isthmus of Panama. Cane & Molnar (2001) instead, proposed that the closing of the Indonesian Seaway, with New Guinea approaching the equator and with much of Halmahera emerging in the last 5 Myr, had more effect on the climates of the Pacific and Indian Oceans than did the closing of the Panamanian isthmus or changes in thermohaline circulation.

As a consequence of these forcing effects and the cyclic variations in eccentricity, axial tilt, and preces-

sion of the Earth's orbit, mathematically theorized by Milankovitch (Crowley & North, 1999), an alternating sequence of glacial and interglacial conditions in approximately 100 kyr cycles characterized the second half of the Quaternary. The strongly affected atmospheric circulation displayed patterns that differed significantly between glacial and interglacial periods. These differences may have been similar to those that exist between the LGM at approximately 20 kyr BP and interglacial warm periods represented by the extremely Holocene Optimum.

The LGM is conventionally defined from sea-level records as the most recent interval in Earth history when global ice sheets reached their maximum integrated volume (Mix, Bard & Schneider, 2001; Clark *et al.*, 2009). Information on precipitation and surface temperature distributions over South America during the LGM, as reconstructed from pollen, lake levels, ocean cores, speleothems, and plant microfossils, indicates levels of aridity that may be interpreted as plant available moisture, precipitation minus evaporation, or some other related variable (Cusminsky *et al.*, 2011). Changes that occurred through the LGM and the Holocene over South America are widely described in Vimeux, Sylvestre & Khodri (2009).

Patagonia was evidently affected by changes in the westerlies and in the position of the southern border of the semi permanent Atlantic and Pacific anticyclones. These changes, which took place during the LGM and the glacial-interglacial transition, are however still poorly understood. The hypothesis that the westerlies shifted north during the LGM has been inferred from terrestrial palaeoecological and marine records off Chile (Heusser, 1989; Stuut & Lamy, 2004). In response to Heusser (1989); Markgraf (1989) proposed that the westerlies shifted to high latitude, giving rise to a long-standing controversy regarding the interpretation of pollen records in Patagonia and the westerlies behavior (Markgraf *et al.*, 1992). However, most of the analyses by proxy data suggest an equatorward shift and intensification of this atmospheric circulation system during the LGM (Lamy *et al.*, 1998, 1999, 2001; Moreno & León, 2003; Valero-Garcés *et al.*, 2005).

The climate modelling results are ambiguous and somewhat contradictory. Although simulations by Kitoh, Murakami & Kiode (2001) and Shin *et al.* (2003) indicate a poleward shift in the surface westerlies, those by Kim, Flato & Boer (2003) indicate a shift towards the equator, and experiments by Otto-Bliesner *et al.* (2006) show no shift in the position of the maximum westerlies. Recently, Rojas *et al.* (2009) analyzed four coupled ocean-atmosphere simulations carried out by the Palaeoclimate Modelling Intercomparison Project Phase 2. They did not find a definite

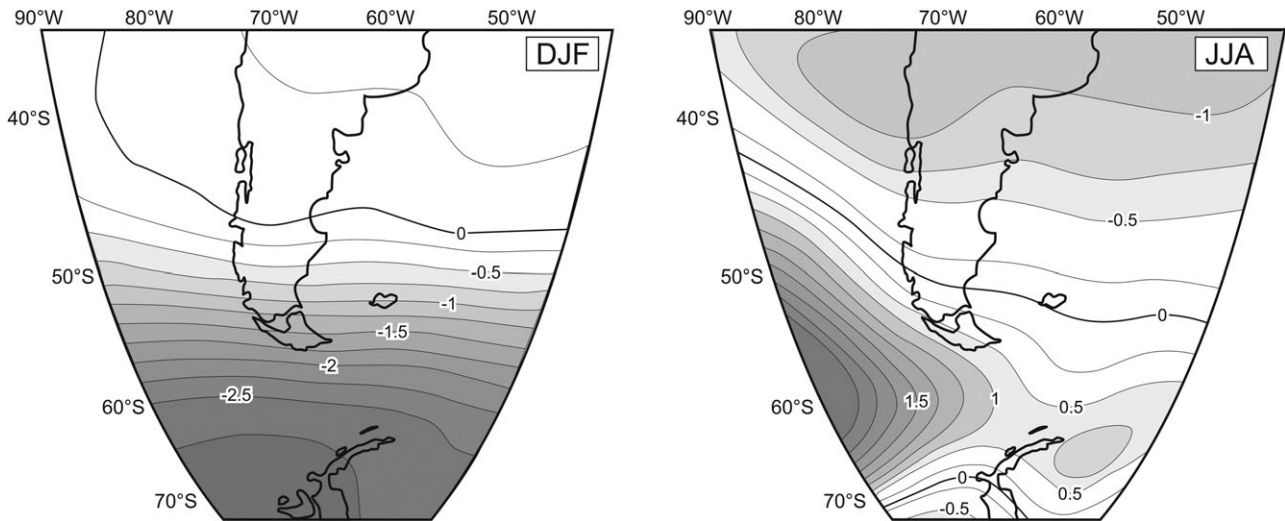


Figure 11. The sea level pressure (SLP, in hPa) difference between the mid-Holocene and the pre-industrial climate for December, January, February (DJF, left) and June, July, August (JJA, right) (*sensu* Wagner *et al.*, 2007).

'shift' in the westerly circulation, but suggest a general decrease in surface windspeeds in the Southern Ocean and sub-Antarctic sectors. This decline, in practice, could induce a similar effect as the hypothesized equatorward shift of the southern margin of the southern westerlies.

THE HOLOCENE: THE OPTIMUM APPROXIMATELY 9000 TO 6000 YEARS BP AND THE LAST MILLENNIUM
The Holocene climate probably resulted from predictable changes in the Earth's orbit and from those which caused the end of the LGM. These orbital changes would have had a maximum heating effect in the Northern Hemisphere climate at 9000 years BP when axial tilt was 24° and the nearest approach to the Sun took place during the boreal summer. The changes in the climatic forcings affected the global atmospheric circulation (Crowley & North, 1999; Issar, 2003). In Patagonia, the westerlies shifted poleward toward their modern position as the climate warmed following the LGM (Toggweiler, Russell & Carson, 2006). This atmospheric change produced dried conditions (i.e. decreased precipitations and lake levels) in northern Patagonia and central Chile (Galloway, Markgraf & Bradbury, 1988; Heusser, 1990; Villagrán & Varela, 1990; Villa-Martínez & Villagrán, 1997; Lamy *et al.*, 2001; Jenny, Wilhelm & Valero-Garcés, 2003).

Wagner *et al.* (2007) model simulation for the Middle-Holocene period between 7.0 and 4.5 kyr BP suggest negative pressure anomalies increased with latitude during summer leading to stronger westerlies over Patagonia than the pre-industrial climate (Fig. 11, left). Winter conditions were very different

from those in the summer during this period (Fig. 11, right). These were characterized by a weakening of the zonal winds and reduced gradients in the polar frontal jet as indicated by the positive pressure anomalies located at 60°S over the southwestern part of the South American southern tip.

Renssen *et al.* (2005) examined potential climate changes through the Holocene using a coupled atmosphere, sea ice, ocean, and vegetation model to simulate atmospheric circulation variables for Southern Hemisphere mid-high latitudes from 9000 years BP to the present conditions.

Their results indicate that, over the Southern Ocean, the Early and Middle-Holocene temperatures were higher than at present in all seasons (Renssen *et al.*, 2005). The wind speed average of over $45\text{--}60^\circ\text{S}$ and the meridional temperature gradient between $23\text{--}30^\circ\text{S}$ and $50\text{--}60^\circ\text{S}$ (Fig. 12) were closely linked, showing stronger westerlies with an increased meridional temperature gradient. The westerlies weakened in winter from 9 to 5 kyr BP, and then remained nearly constant up to the last 2 kyr and increased weakly in the last millennium. In the other seasons, and especially for summer, the wind had opposite behavior.

THE LAST MILLENNIUM

Two opposing climate anomalies took place during the last millennium, the so-called Medieval Warm Period (MWP, approximately 9th to 13th Centuries AD, Lamb, 1982; Jansen *et al.*, 2007) and the Little Ice Age (LIA) from approximately 1450 AD to 1890 AD. The LIA comprised three main cold stages, the first beginning at approximately 1650 AD, the second at 1770 AD, and

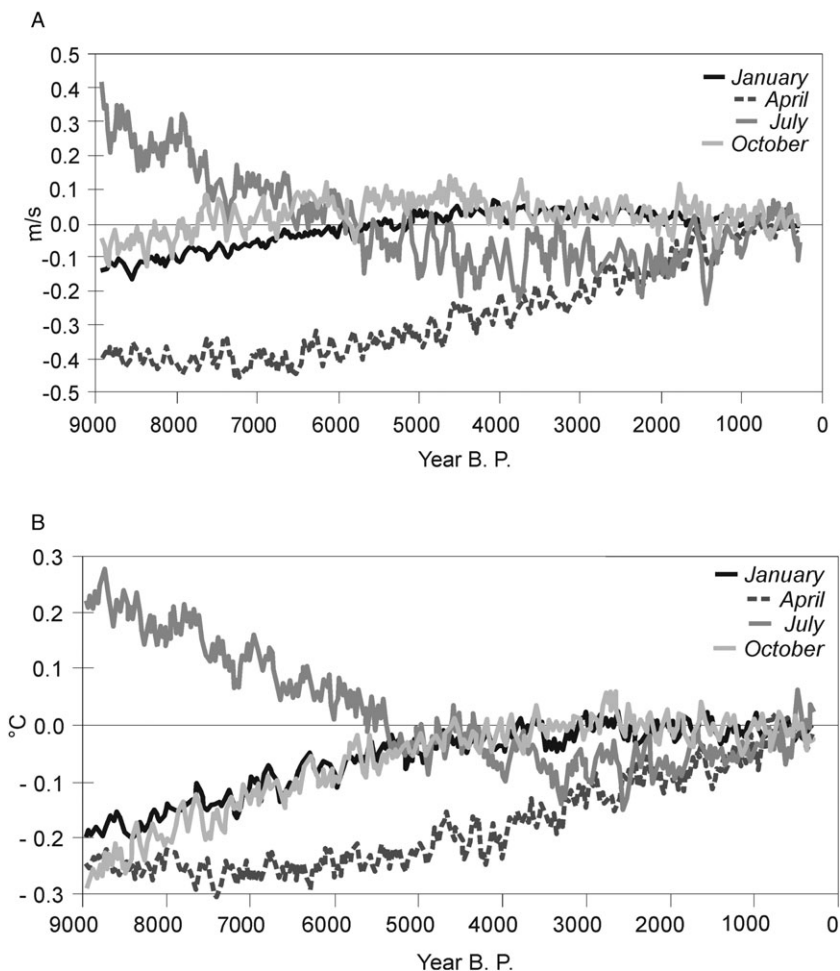


Figure 12. Simulated anomalies time series, through the last 9000 years, as deviations from the preindustrial mean (1000/250 years BP) for (A) wind speed (m s^{-1}) over $45\text{--}60^\circ\text{S}$ and (B) meridional gradient ($^\circ\text{C}$), $20\text{--}30^\circ\text{S}$ minus $50\text{--}60^\circ\text{S}$ (*sensu* Renssen *et al.*, 2005).

the third at 1850 AD. They were separated by intervals of slight warming. The cooler temperatures were caused by a combination of large volcanic eruptions, less solar activity during the Spörer (1460–1550), Maunder (1645–1715), and Dalton (1790–1830) minima and the inherent climate variability (Crowley & North, 1999; Jansen *et al.*, 2007).

Evidence of MWP and LIA effects over Patagonia has been recovered from different localities, including Laguna Aculeo in Central Chile (Jenny *et al.*, 2002; von Gunten *et al.*, 2009); the Jacaf Channel of Chilean Northern Patagonia (Rebolledo *et al.*, 2008; Sepulveda *et al.*, 2009) Laguna Potrok Aike (Zolitschka *et al.*, 2004; Haberzettl *et al.*, 2005); and Valle de Andorra, Ushuaia (Mauquoy *et al.*, 2004). Glacier retractions following the LIA have been documented for the Laguna San Rafael and the glacier of the same name (Araneda *et al.*, 2007), for the North Patagonian Icefields (Hielo Patagonico

Norte; Harrison, Winchester & Glasser, 2007), for the Proglacial Lago Frías at Monte Tronador in northern Patagonia (Rabassa, Rubulis & Suárez, 1979; Villalba *et al.*, 1990), and for the Glaciar Río Manso, also at Monte Tronador (Masiokas *et al.*, 2010), among others.

The atmospheric circulation at this time appears to have been characterized by an increase of the meridional component at high latitudes of the Southern Hemisphere (Kreutz *et al.*, 1997) and by a cooling at the South Pole, resulting in stronger zonal westerlies (Mosley-Thompson *et al.*, 1990). Palaeoecological reconstruction and model simulation for LIA in southern Patagonia show that the precipitation fluctuation can be attributed to latitudinal northward shift of the westerlies (Mayr *et al.*, 2007; Meyer & Wagner, 2008). Recently, Maenza & Compagnucci (2010) obtained scenarios of the anomalies between the Maunder Minimum and the last two decades of the 20th

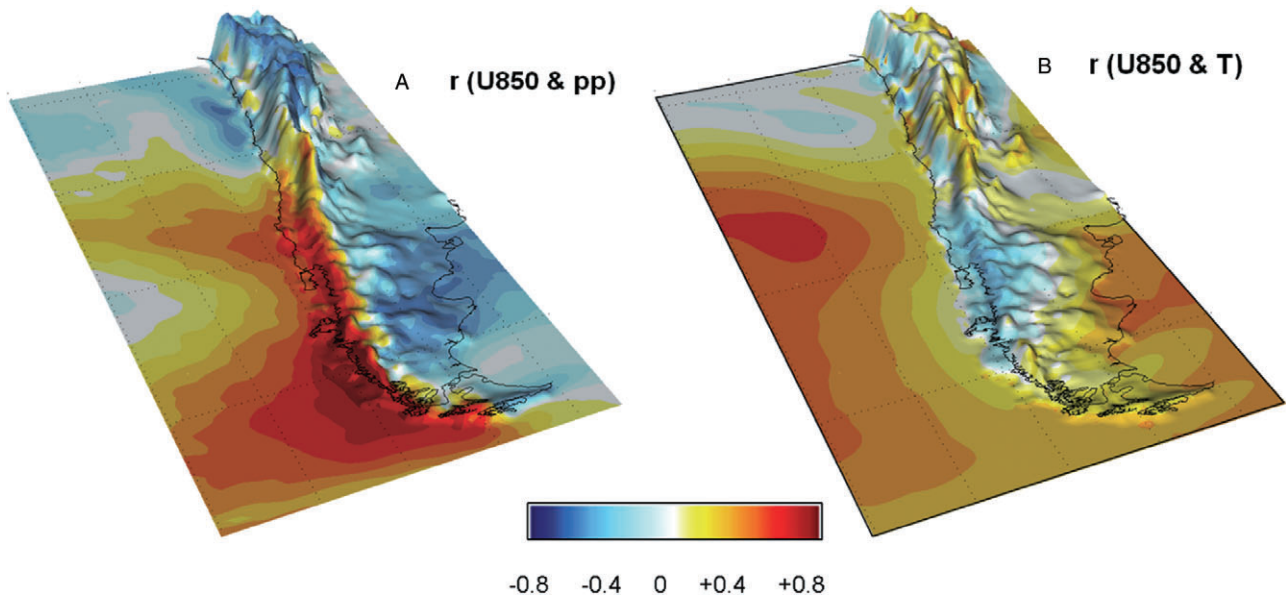


Figure 13. Mean annual values of the correlation coefficients between the zonal wind component at 850 hPa and (A) the total precipitation and (B) the surface air temperature for the second half of the 20th Century [obtained from R. D. Garreaud, pers. comm.; *sensu* Garreaud *et al.* (2010)].

Century. The results show the zonal wind component increased over Central and Northern Patagonia and decreased in the southern tip during the warm semester, whereas, during the colder semester, the wind increased all over Patagonia, reaching a maximum over the southern tip. Varma *et al.* (2010) used a global climate model and proxies reflecting changes in the position of the westerlies, to infer that, centennial-scale periods of lower solar activity caused equatorward shifts of the westerlies during the past 3000 years.

At present time, the climatic anomalies in northern Patagonia could be related to El Niño (Garreaud *et al.*, 2009) and the whole region appears to be affected by the Southern Annular Mode. This mode is the most important pattern of the atmospheric circulation variability for the middle and high latitudes in the Southern Hemisphere.

The atmospheric circulation at surface over the southern cone of South America corresponds to the behavior synthetically described above (Fig. 3). Among the daily atmospheric circulation types obtained by Compagnucci & Salles (1997) the mean field or basic flow, with the westerlies prevailing over Patagonia, explains more than 50% of the total variance. The remaining variance is explained by different synoptic features representing an increased meridional component (e.g. passage of low and high pressure systems) relative to the basic flow. The variance explained by the basic flow during winters of El Niño years is lower than during winters of La Niña years (Compagnucci & Vargas, 1998).

The westerly component of the low-level flow ahead of the cold front is largely blocked by the Andes, producing a northerly jet along the coast of south-central Chile. This results in a drastic precipitation change between the sides of the Andean Cordillera, affecting temperature as well (Garreaud, 2009; Garreaud *et al.*, 2009). On the basis of instrumental observations along with downscale large-scale signals and up-scale local environmental changes, Garreaud, Lopez & Rojas (2010) described how the lower level (850 hPa) zonal flow (U) strongly modulates precipitation (pp) and surface air temperature (T) (Fig. 13). The mean annual value of zonal wind at 850 hPa correlation coefficients with precipitation (Fig. 13A) and temperature (Fig. 13B) show that the year-to-year changes in the zonal flow are strongly and positively (negatively) correlated with annual mean precipitation changes to the west (east) to the Andes. The seasonal variation of the precipitation dependence to the zonal wind appears to be weak. Furthermore, the surface air temperature relationship exhibits seasonal and geographical dependence. Stronger than normal westerlies flow are associated with milder than normal winters in western Patagonia but colder than normal summers in eastern Patagonia.

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