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PALAEO
Palaeogeography, Palaeoclimatology, Palaeoecology 231 (2006) 134–157
www.elsevier.com/locate/palaeo

Antarctic climate at the Eocene/Oligocene boundary — climate model sensitivity to high latitude vegetation type and comparisons with the palaeobotanical record

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Received 18 March 2005; accepted 8 July 2005

Abstract

Simulated climate for the Antarctic continent using the GENESIS (Version 2.1) Global Climate Model with 34 Ma boundary conditions is shown to be highly sensitive to polar vegetation type. Six experiments were run using different levels of atmospheric CO_2, orbital configurations, ice sheet geometries and vegetation types to assess model sensitivity to Antarctica being covered with either a needle leaf evergreen forest or tundra. Simulations using 2× pre-industrial levels of CO_2 (combined mean annual temperature (MAT) −19 °C) are about 7 °C cooler than minimum estimates from the Antarctic Cenozoic plant record (MAT −12 to 15 °C). However, simulations using 3× CO_2 (MAT −7 °C) are in good agreement with our empirical estimates of mean annual temperature. With ice sheets and orbits set up to represent early Oligocene interglacial conditions, the tundra climate is significantly cooler than the evergreen forest climate, with local, austral summer averages up to 6 °C cooler in non-glaciated areas and continental averages ~2.5 °C cooler. In the model this is mainly due to higher albedo and decreases in net radiation and sensible and latent heat flux, especially during spring and summer. Feedbacks between coastal and continental cooling, marginal sea surface temperatures and sea ice also appear to be significant. A review of the late Palaeocene to earliest Miocene plant fossil record in the Antarctic Peninsula and Ross Sea regions shows that the vegetation was in transition ~34 Ma, from a relatively diverse, mainly evergreen forest to a tundra vegetation. The modelled sensitivity of continental temperatures to a change from forest to tundra suggests vegetation-climate feedbacks during the Eocene–Oligocene transition played a significant role in the initial rapid glaciation of the continent.

Keywords: Global climate models; Palaeoclimate; Antarctica; Cenozoic; Palaeobotany; Greenhouse-icehouse transition

1. Introduction

During the last few decades there has been increasing interest in the geological record of polar climate. Interpretations from the fossil plant record, particularly during globally warmer times in the past, are improving our understanding of the potential response of the Antarctic ice sheet to anthropogenically influenced rising temperatures. It is widely acknowledged that the

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doi:10.1016/j.palaeo.2005.07.032

global 'Greenhouse' state started in the Triassic (Frakes et al., 1992) with Antarctica remaining largely ice-free until ~34 Ma (the Eocene/Oligocene (E/O) boundary). The plausibility of the Intergovernmental Panel on Climate Change predictions of Earth system responses to increasing levels of atmospheric greenhouse gases are reliant on complex Global Climate Models (GCMs, e.g., McGuffie and Henderson-Sellers, 1997; Trenberth, 1992). GCMs are now widely utilized and accepted as a tool for providing plausible scenarios to complex natural system problems and to help us plan for possible futures in a rapidly warming world. Because these models allow the detailed simulation of the Earth system and its response to external forcing, palaeogeography, atmospheric chemistry, changing distributions of terrestrial ecosystems (e.g., Bergengren et al., 2001; Foley et al., 2000; Henderson-Sellers, 1993; Otto-Bliesner and Upchurch, 1997), and ice sheets (e.g., DeConto and Pollard, 2003a; Siegert et al., 2003), they are applicable to a wide range of palaeoclimate scenarios. Palaeoclimate simulations have become a widely accepted means for both the testing of geological data-driven hypotheses and the evaluation of model performance under a wide range of boundary conditions and forcing scenarios.

With the advent of palaeoclimate modelling it has become important to characterize past vegetation type and distribution because of its influence as a boundary condition on palaeoclimate model output (Dutton and Barron, 1996). One approach to accounting for evolving continental land cover in palaeoclimate simulations involves the coupling of predictive vegetation models (e.g., EVE or TRIFFID, Cox, 2001; DeConto et al., 1999a,b; Foley et al., 1996; Prentice et al., 1992) to a GCM. A simpler approach and the one adopted here incorporates a prescribed distribution of vegetation, which remains constant as the modelled climate evolves. Altering this fixed vegetation between equilibrated model runs provides information on the sensitivity of the simulated climates to the nature and distribution of different continental biomes. Empirical data is an essential component of such simulations providing both realistic boundary conditions and output calibration.

This paper investigates model sensitivity to different Antarctic vegetation and orbital forcing, and assesses the model's accuracy against the proxy palaeoclimate record from fossil plants. The GENESIS GCM (Version 2.1) is used to model palaeoclimate at 34 Ma around the time of the initiation of the East Antarctic Ice Sheet. Prescribed biome distributions representative of the Antarctic proxy record and preglacial, interglacial and glacial earliest Oligocene orbital and ice sheet scenarios are used in the model runs. Averaged model outputs for the Antarctic landmass, including specific regions significant for the proxy record, are then compared with annual and seasonal near-surface temperatures and precipitation values interpreted directly from the plant fossil record using modern analogue techniques.

2. The proxy vegetation and climate record from Antarctica

The fossil plant record of Antarctica is exposed today in ice-free areas and is increasingly accessible beneath sea ice and floating ice shelves from ocean sediments using state-of-the-art drilling technology e.g., the Cape Roberts Project (CRP, http://www.geo.vuw.ac.nz/croberts/index.html) and planned ANDRILL Project (http://andrill-server.unl.edu/). Information from these past floras suggests a much greater cover and diversity of vegetation over the continent during times when global climate was much warmer than at present. A significant proportion of the Antarctic flora since the mid Cretaceous consists of angiosperms, notably several species of *Nothofagus* (the Southern Beech tree) (Hill and Scriven, 1995). Cretaceous and Cenozoic macroflora including coniferous fossil forests are primarily found in the Antarctic Peninsula region (e.g., Falcon-Lang and Cantrill, 2001; Falcon-Lang et al., 2001; Francis, 1999; Poole and Cantrill, 2001). In Victoria Land, Cenozoic plant macrofossils have been found in the Sirius Group deposits, the McMurdo Sound erratics and as single *Nothofagus* leaves in the CIROS-1 and CRP-3 near-shore cores (Ashworth and Cantrill, 2004; Cantrill, 2001; Francis, 2000; Francis and Hill, 1996; Hill, 1989). Terrestrial palynomorph assemblages, which provide a more regional picture of contemporaneous onshore vegetation growth, were also found in both of these regions but can be complicated by the presence of long-ranging often recycled taxa (e.g., Askin and Raine, 2000; Mildenhall, 1989; Prebble et al., this volume; Raine and Askin, 2001). Key reviews of Antarctic vegetation and palaeoclimate have been published by Truswell (1990), Askin (1992), Hill and Scriven (1995) and Francis (1999).

A series of increases in benthic marine $\delta^{18}O$ provide the most commonly cited evidence for Palaeogene cooling and East Antarctic glaciation (e.g., Zachos et al., 2001). Oligocene isotope event 1 (Oi1) in the nomenclature of Miller et al. (1991) is the biggest (>1‰) of these increases and is generally believed to coincide with the sudden (<10^5 year) growth of the East Antarctic Ice Sheet (Flower, 1999; Zachos et al., 1996). The majority of palaeobotanical records of Antarctic margin vegetation growing during the late Palaeocene to early Miocene (bracketing the E/O boundary and of significance for assessing the proxy climate record across the Oi1 transition) are located within two distinct geographic regions: the Antarctic Peninsula and the Ross Sea. Fig. 1 and Table 1 summarize the localities of the significant deposits in these regions reviewed for this study. Fig. 2 presents a summary of palaeovegetation compositional data from fossil leaves, wood and terrestrial palynomorphs discovered at these localities along with their modern analogue interpretations.

Continent-wide fossil evidence suggests the flora evolved in response to an overall cooling of the climate throughout the late Palaeocene to early Miocene. Fossil plants from the late Palaeocene suggest the growth of tall stature vegetation consisting of evergreen mixed broadleaf and needle leaf trees and shrubs (dominated by *Nothofagus* spp.). Through time, diversity decreased (only slightly along the Victoria Land coast, Prebble et al., this volume) along with temperature and the canopy lowered and opened to a low stature, tundra-like vegetation in the early Miocene. The E/O boundary occurs within a transition period between these two vegetation types (Fig. 2). Late Eocene to earliest Miocene floras have been compared to the Valdivian–Magellanic forest/woodland in Chile south of latitude 47° 30′ S, *Nothofagus* vegetation at Tierra del Fuego, and the fern-bush communities of the subantarctic islands, e.g., the Auckland Island group south of New Zealand. Despite the general sparseness and wide distribution of the plant fossil record from this period across the Antarctic continent we can assume that a mosaic of vegetation types occupied varied niches in the landscape from quite sheltered low altitude valleys, perhaps filled with the remnants of a temperate rainforest, to higher altitude more exposed plains covered with low shrubland and tundra.

Fossil plant assemblages can provide direct information on past terrestrial climate on local to continent-wide scales. The use of both quantitative and qualitative analytical methods for the interpretation of fossil plants allows estimates of certain palaeoclimate parameters e.g., mean annual temperature, coldest and warmest mean monthly temperature, and mean annual precipitation, depending on the chosen technique. Despite inherent limitations to each technique (reviewed by Askin, 1992; Jones and Rowe, 1999) remains of fossil plants provide rare and valuable evidence of terrestrial palaeoclimate throughout geological history.

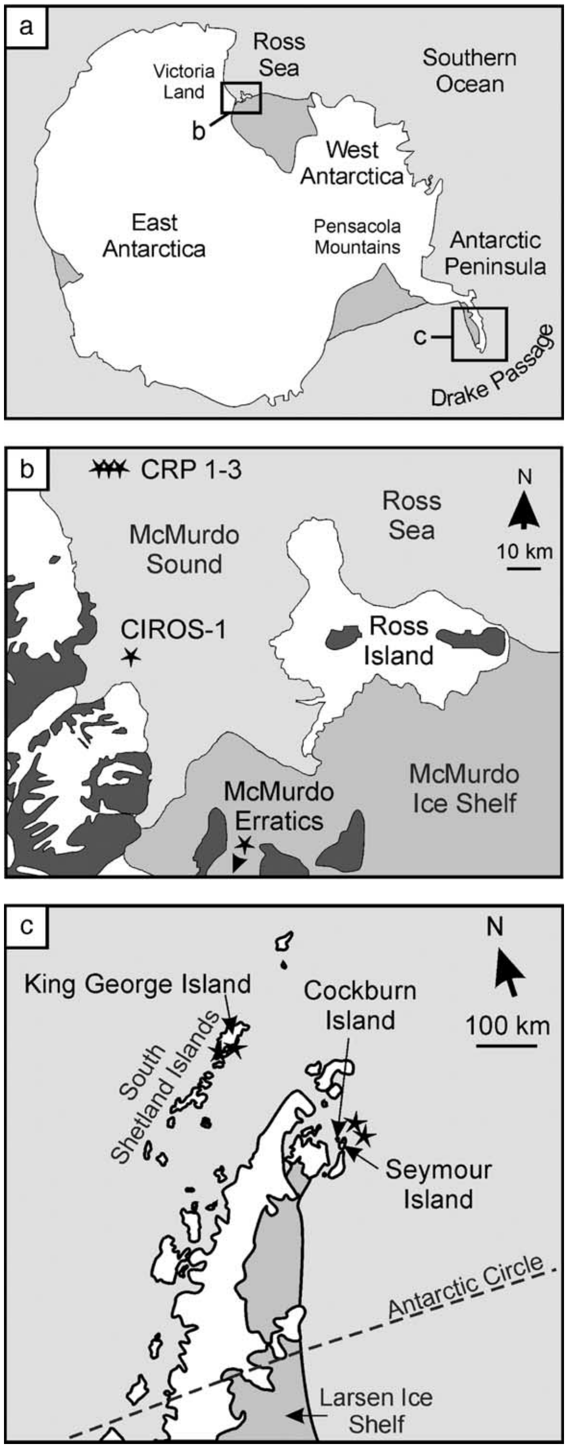

Fig. 1. Maps to illustrate localities discussed in the text and to show sites of previously described ?Palaeocene to Miocene fossil plant material reviewed for this study. Refer to Table 1 for details of each deposit. (a) Antarctica, (b) the Ross Sea region, (c) the Antarctic Peninsula region.

Table 1
Details of ?Palaeocene to Miocene fossil plant deposits from the Antarctic peninsula and Ross Sea regions reviewed for this study (refer to Fig. 1 for location maps)

Stratigraphic age and lithostratigraphy	Location	Modern analogue palaeoclimate parameters	Selected references
Antarctic Peninsula region			
Oligocene-Miocene transition: Mount Wavel Formation	Point Hennequin, South King George Island, South Shetland Islands	MAT 5–8 °C MAP 600–4300 mm	Birkenmajer and Zastawniak, 1989; Askin, 1992
Early Oligocene: Point Thomas Formation, Ezcurra Inlet Group	South King George Island, South Shetland Islands	MAT 11.7–15 °C MAP 3225–1220 mm	Birkenmajer, 1997; Birkenmajer and Zastawniak, 1989; Askin, 1992
Eocene– ?earliest Oligocene: La Meseta Formation, Seymour Island Group	Seymour and Cockburn Islands, James Ross Basin	*Dicotyledonous wood*: MAT 7–15 °C	Case, 1988; Francis, 1991; Askin, 1992, 1997; Francis and Poole, 2002
?Palaeocene: Dufayel Island Group	Dufayel Island, Barton Peninsula, King George Island, South Shetland Islands	MAT 10–12 °C MAP 1000–4000 mm → warm, wet	Birkenmajer and Zastawniak, 1986, 1989; Askin, 1992
Ross Sea region			
Early Miocene: CIROS-1 core	McMurdo Sound, Ross Sea region	Cold, disturbed interglacial MAT 5 °C (*Nothofagus gunnii* habitat in Tasmania)	Hill, 1989; Francis, 1999; Raine, 1998; Roberts et al., 2003
Late Oligocene/Miocene: CRP-2/2A core (above ~300 mbsf)	Off Cape Roberts, McMurdo Sound, Ross Sea region	Cool to cold Antarctic Zone of Greene (1964) = south of 60°, but incl. South Sandwich Islands and Bouvet Island: MST *c.* 5–7 °C Patagonia: MST 8–12 °C; Arctic (*Salix arctica* habitat): MST 4 °C Periglacial: MST 7–10 °C, MWT 1–2– 7 °C	Raine, 1998; Askin, 2000; Askin and Raine, 2000; Prebble et al., this volume (and references therein)
Early Oligocene: CRP-2/2A (below ~300mbsf) and CRP-3 core (above 410 mbsf)	Off Cape Roberts, McMurdo Sound, Ross Sea region	Cold temperate (generically MWT < −3 °C; MST > 10 °C)-periglacial; MST (at SL) ~10–12 °C Patagonia: MST 8–12 °C; Arctic (*Salix arctica* habitat): MST 4 °C *Leaf:* comparable to *Nothofagus beardmorensis* (described from the Sirius Group): growing season up to 5 °C (short and warm), winter season down to −15 to −22 °C, MAT −12 °C	Francis and Hill, 1996; Cantrill, 2001; Raine and Askin, 2001; Prebble et al., this volume (and references therein)
Middle – Upper Eocene: McMurdo Sound erratics	Moraine ridges on Minna Bluff and Mt. Discovery, McMurdo Sound, Ross Sea region	*Wood*: Cool (?warm) temperate MAT < 13 °C	Axelrod, 1984; Veblen et al., 1996; Francis, 1999; Askin, 2000; Pole et al., 2000

Quantitative modern analogue climate data are quoted from the sources listed, and where only qualitative climate terms were used in the original interpretation, e.g., 'cold temperate', relevant temperature and precipitation ranges are extrapolated from Philip (1998). MAT, mean annual temperature; MST, mean summer temperature; MWT, mean winter temperature; MAP, mean annual precipitation.

Palaeoclimate data interpreted from Antarctic fossil flora and useful for comparison with the GCM simulated data are listed in Table 1 with selected sources from the literature. Only deposits with well-constrained age control are included, which precludes the controversial Sirius Group, which may be as old as early

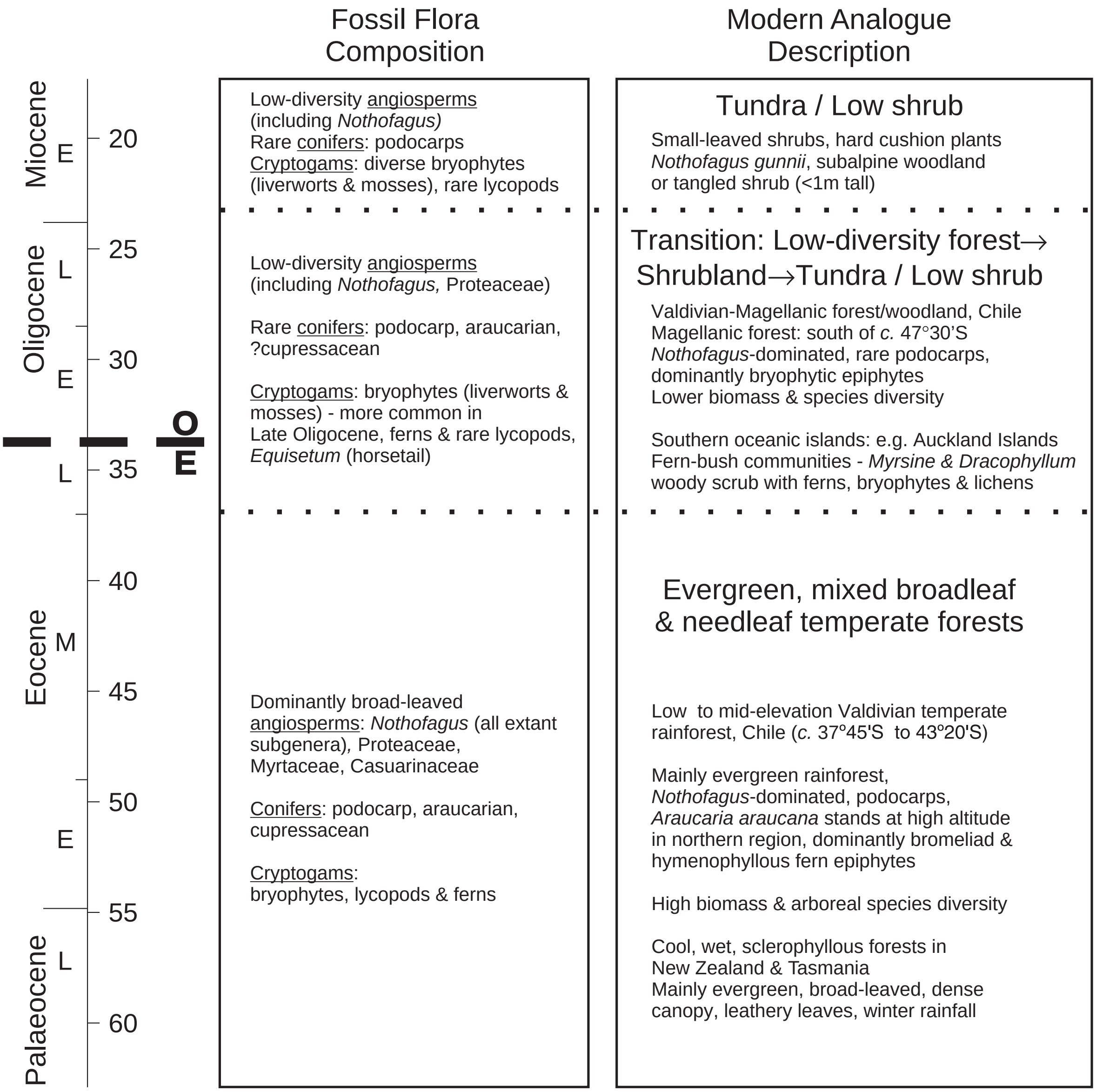

Fig. 2. Summary of the composition of the fossil plant assemblages reviewed for this study and modern analogue vegetation types. For sources refer to Table 1. The Eocene/Oligocene (E/O) boundary at 34 Ma, used for the climate model experiments, is highlighted. The modern analogue vegetation is divided into three zones describing the change from relatively diverse, mainly evergreen rainforest during the Late Palaeocene to Middle Eocene, to a tundra/low shrub community in the earliest Miocene. The E/O boundary occurs within the 'transition' zone between these two vegetation types.

Miocene (Hicock et al., 2003; Miller and Mabin, 1998). Temperature and precipitation figures are derived from the modern climate within which grows the closest analogue of the fossil vegetation assemblage. Commonly these parameters are quoted qualitatively rather than quantitatively due to the uncertainties inherent in the use of the 'Nearest Modern Relative' technique and broad climatic terms are used such as 'cool temperate' or 'periglacial'. However, to be able to compare quantitative climate model output with proxy palaeoclimate information from fossil plants such qualitative information is assigned generic temperature and precipitation estimates for this study based on atlas definitions of climatic zones (Philip, 1998). Where quantitative temperature data is quoted commonly mean annual temperature is used. However, when interested in the initiation of ice sheet growth the mean summer temperatures are more relevant due to their influence on the survivability of snow throughout successive years.

3. Model description

We use the latest version of the GENESIS GCM (Version 2.1, Pollard and Thompson, 1997; Thompson and Pollard, 1997) to test the sensitivity of a partially glaciated Palaeogene Antarctica to a changing land cover. GENESIS has been used extensively in past, present, and future climate studies including simulations of the E/O boundary and long time-continuous simulations of the Oi1 glaciation using a coupled dynamical ice sheet model component (DeConto and Pollard, 2003a,b; Pollard and DeConto, 2003). The atmospheric component of GENESIS has 18 vertical layers and a spectral resolution of T31 (~3.75° latitude × 3.75° longitude) coupled to 2° × 2° surface model components including a non-dynamical 50 m slab ocean, a dynamic–thermodynamic sea ice model and multi-layer models of soil, snow and vegetation. Ocean heat transport is not prescribed, but is calculated as linear diffusion down the local ocean temperature gradient with the diffusion coefficient depending on latitude and the zonal fraction of land versus sea. The use of a computationally efficient slab ocean model allows for multiple sensitivity experiments and is likely a reasonable approach given recent palaeo-AOGCM (coupled atmosphere–ocean-GCM) simulations resulting in near-modern poleward ocean heat transports over a wide range of ancient boundary conditions and forcings e.g., Late Cretaceous (Otto-Bliesner et al., 2002), Eocene (Huber and Sloan, 2001) and Pliocene (Haywood and Valdes, 2004).

3.1. Vegetation representation in the GCM

Vegetation is responsible for complex climatically important transfers of heat and moisture between the land surface and the atmosphere (Dickinson et al., 1986) and its representation in the GCM is a key aspect of this study. In the GCM configuration used here the Land-Surface-Transfer (LSX) component (Pollard and Thompson, 1995) uses three of 12 explicit vegetation types (Dorman and Sellers, 1989) and their associated parameters of canopy structure (heights), Leaf Area Index, fractional cover, seasonal phenology, leaf transmittance and reflectance (albedo), roughness length (turbulent transport) and biophysical control of evapotranspiration (stomatal resistance) to represent the vegetation cover in a given 2° × 2° land surface grid cell. LSX is a derivative of early land-surface/vegetation models e.g., the Biosphere–Atmosphere Transfer Scheme (BATS, Dickinson et al., 1986) and the Simple Biosphere Model (SiB, Sellers et al., 1986) and is similar in design to the Land Surface Model (LSM) in the NCAR CCSM (National Center for Atmospheric Research Community Climate System Model, Bonan, 1998). The model calculates radiative and turbulent fluxes through an upper and lower canopy to the soil or snow covered surface below while accounting for the interception of rain and snow, which subsequently drips or blows off. Vegetation temperatures, canopy air temperatures, and specific humidities are calculated from the atmospheric conditions above the upper canopy and the soil or snow conditions below. Prognostic variables are then passed back to the atmosphere allowing realistic interaction between the GCM's land surface and atmospheric components. In the model, partitioning of latent heat and sensible heat is dependent on canopy structure, soil moisture, holding capacity of the rooted soil layer and local rates of precipitation, evaporation and transpiration. Leaf and stem temperatures, air temperature, specific humidity, and intercepted precipitation are calculated for both the upper and lower canopies, as are soil and snow surface temperatures.

The GCM also includes a six-layer soil model extending to a depth of 4.25 m. The soil model calculates the downward diffusion of heat and moisture, desiccation of the rooted zone as a function of transpiration, surface runoff, subsurface drainage and ponding of liquid water on the surface. A three-layer snow model represents snow cover over soil, ice sheet and sea ice surfaces with total snow thickness and fractional coverage controlled by the melting and accumulation rates on the uppermost layer. In high latitudes, the explicit treatment of snow and vegetation including biome-specific canopy height and seasonal phenology allows for the seasonal masking of snow with an important climatic effect on albedo and net radiation as discussed below.

The capability to simulate the physical climatic effects of vegetation is of particular importance to the study of a partially glaciated Palaeogene Antarctica when vast areas of the continent were ice-free and susceptible to changing distributions of terrestrial ecosystems. Earlier versions of the GENESIS GCM have been used to test the sensitivity of global Cretaceous (Otto-Bliesner and Upchurch, 1997) and Miocene (Dutton and Barron, 1996) climates to different global distributions of vegetation assuming either an ice-free (Cretaceous) or mostly ice-covered (Neogene) Antarctica. These studies and others clearly point to the importance of vegetation type in palaeoclimate simulations. This is especially true over high latitudes where the replacement of tundra with evergreen forest has been shown to have a warming effect of up to ~14 °C during spring and summer (Bonan et al., 1992; DeConto et al., 1999a,b; Dutton and Barron, 1996;

Robinson and Kukla, 1985). Most of the warming stems from a reduction of albedo in areas with significant seasonal snowfall where evergreen trees with tall snow-free canopies hide the high albedo of snow-covered soil or low-canopy biomes such as tundra (Bonan et al., 1995).

During the development of the SiB biome categories used in this study Dorman and Sellers (1989) simulated the surface albedo of several vegetation cover categories. Needle leaf evergreen trees ranged from 0.10– 0.12 (boreal summer to winter) and 0.28 when snow-covered (although this latter figure they considered too low). This illustrates the propensity of tall trees *not* to retain snow cover, which is commonly dislodged in strong winds, resulting in a relatively low albedo year-round. In contrast, the low stature tundra biome had a relatively high albedo even when snow-free with a simulated summer albedo between 0.12 and 0.26 (depending on the soil background colour). Therefore, it is evident that tall green vegetation, if it remains free of snow cover, is a highly effective radiative trap for the visible wavelengths of light involved in photosynthesis and the issue of snow cover is important as land surface albedo can be altered significantly. Further, tree height in comparison to low stature tundra vegetation (roughness length) also influences the temperature of the near-surface through enhanced heat exchanges and turbulent mixing. The larger amount of photosynthetically active biomass in tree canopies in contrast to tundra absorbs more incoming radiation with the sun higher above the horizon during the summer months and also has higher latent heat flux due to increased evapotranspiration.

In a Late Cretaceous modelling study using the GENESIS GCM interactively coupled to a predictive vegetation model (EVE, Bergengren et al., 2001), DeConto et al. (1999a,b) recognized an atmospheric CO_2 threshold, below which forest could not be maintained on an ice-free Antarctic continent. As tundra spread at the expense of forest, cooling accelerated via snow-albedo feedbacks leading to further expansion of tundra. Similar tundra-cooling feedbacks have been suggested as a potential contributor to Quaternary glaciation (Gallimore and Kutzbach, 1996) and may have been an important contributor to episodes of Antarctic cooling prior to the full permanent glaciation of the continent sometime in the Neogene.

4. Experimental design and boundary conditions

To address the potential climatic effects of changing Antarctic vegetation around the time of the E/O transition we ran a suite of six equilibrium climate model simulations using the palaeogeographic boundary conditions from prior climate-ice sheet simulations of the Oi1 glaciation (DeConto and Pollard, 2003a) with varying CO_2 concentrations, ice sheet configurations, orbital parameters and Antarctic vegetation (Table 2). The 34 Ma global palaeogeography is based on a global tectonic reconstruction using the plate rotations in Hay et al. (1999) with superimposed early Oligocene shorelines and topography (Ronov et al., 1989). Antarctic topography is based on an initially isostatically rebounded ice-free topography with the bedrock elevations (Bamber and Bindschadler, 1997) modified to account for the load of the prescribed ice sheets (Brotchie and Sylvester, 1969). Two sets of simulations were run with 2× and 3× pre-industrial levels of atmospheric CO_2 representing the end-member greenhouse gas concentrations before and after the E/O 'Greenhouse–Icehouse' transition, as determined by a long time-continuous GCM-ice sheet simulation of the Oi1 glaciation (DeConto and Pollard, 2003b). Therefore, we assume that the 3× CO_2 simulations are representative of conditions just prior to glaciation (preglacial) and the 2× CO_2 simulations more representative of an early Oligocene Antarctica (both interglacial and glacial). Although prior coupled GCM-ice sheet simulations have shown that the continent would likely be fully glaciated at 2× CO_2 it is considered worth exploring this scenario as it is close to the CO_2 threshold for ice sheet initiation. Antarctic climate and ice volume has been shown to be highly sensitive to orbital forcing close to the glaciation threshold (Pollard

Table 2
Model simulations and relevant model inputs

Model run	CO_2 mixing ratio[a]	Orbital parameters (ecc., obl., pre.)[b]	Ice sheet geometry[c]	Antarctic vegetation[d]
1	$2\times CO_2$	0.04, 23.5, 90	interglacial	TUND
2	$2\times CO_2$	0.04, 23.5, 90	interglacial	EVER
3	$2\times CO_2$	0.01, 22.5, 270.0	glacial	TUND
4	$2\times CO_2$	0.01, 22.5, 270.0	glacial	EVER
5	$3\times CO_2$	0.04, 23.5, 90	preglacial	TUND
6	$3\times CO_2$	0.04, 23.5, 90	preglacial	EVER

Palaeogeographic boundary conditions were the same as used in prior climate-ice sheet simulations of the Oi1 transition (DeConto and Pollard, 2003a).

[a] CO_2 mixing ratios in multiples of the pre-industrial level of 280 ppm (Houghton et al., 1990).

[b] Idealised orbital parameters (eccentricity, obliquity and precession) producing relatively cold or warm austral summers. Values of precession are shown as the prograde angle between perihelion and the Northern Hemisphere vernal equinox.

[c] Ice sheet configurations representing a partly glaciated (preglacial, interglacial), and fully glaciated (glacial) East Antarctica.

[d] EVER, needle leaf evergreen forest vegetation; TUND, tundra vegetation.

and DeConto, 2005) and vegetation feedbacks may also be important.

Three different ice sheet geometries are provided by prior coupled GCM-ice sheet model simulations of the E/O boundary (DeConto and Pollard, 2003a) at 3× and 2× CO_2 and with orbital parameters representing 'preglacial', 'interglacial' and 'glacial' Antarctic climate conditions (Fig. 3). Small ice caps in the 3× CO_2 simulations are representative of a late Eocene preglacial scenario while two different ice sheet configurations, along with their associated orbital parameters, represent the range of expected climate (glacial vs. interglacial) variability in the earliest Oligocene after the Oi1 transition. Orbital parameters were chosen from those used in prior coupled GCM-ice sheet model simulations (DeConto and Pollard, 2003a) to represent conditions generally favourable or unfavourable for ice growth. While the prescribed orbits do not produce the extremes in Antarctic summer insolation the medium eccentricity and obliquity in the 'preglacial' and 'interglacial' cases produce relatively warm austral summers. Conversely, the low eccentricity and obliquity in the 'glacial case' is more favourable for ice growth. In the case of low eccentricity, the choice of precession becomes less important and low eccentricity, low obliquity orbits have been implicated in early Miocene glacial advances (Naish et al., 2001; Zachos et al., 1997).

The two different floral community end members in transition across the E/O boundary in Antarctica are each assigned the closest comparable Dorman and Sellers (1989) biome categories as vegetation boundary conditions for the GCM. Category '4' (needle leaf evergreen trees, EVER) was chosen to reflect the forest community control case and category '10' (dwarf trees and shrubs with groundcover, TUND) to reflect the tundra. Despite the mainly broad-leaved nature of the fossil plant record across the E/O boundary, the pre-

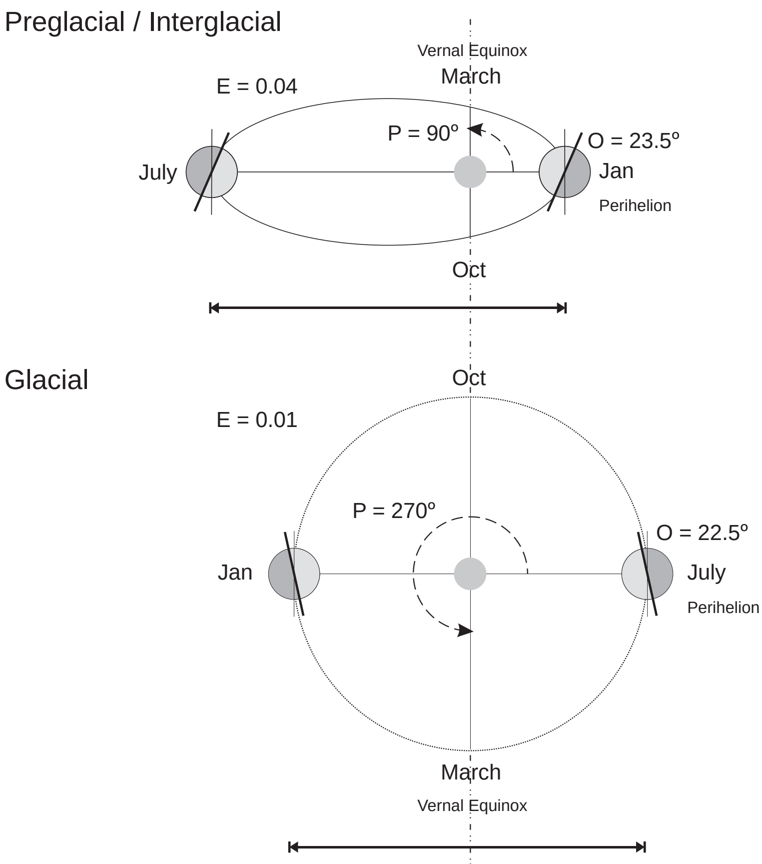

Fig. 3. Geometry of the synthetic orbital parameters used in the two sets of 'preglacial/interglacial' and 'glacial' climate model simulations based on prior climate-ice sheet simulations of the Eocene/Oligocene boundary (DeConto and Pollard, 2003a). The 'preglacial/interglacial' configuration uses medium obliquity (O) and eccentricity (E), with precession (P) placing perihelion in January to produce relatively warm austral summers. In contrast, the 'glacial' orbital configuration uses low obliquity and eccentricity, with perihelion in July to produce relatively cold austral summers and an overall reduction in seasonality.

dominantly evergreen nature of this palaeovegetation is considered to be of significant importance in relation to the effect on the modelled climate. Biome category '4' was chosen to reflect this seasonality in the forest foliage within the limitations of the scheme and to ensure a significant difference to the tundra designation. This choice of alternative model biomes allows an exploration of the widest possible envelope of vegetation feedback. For any given model experiment the appropriate biome category was supplied as a blanket land surface cover over Antarctica. Category '6' (broadleaf trees with groundcover-savanna) was used as a general mixed biome for the remainder of the terrestrial regions of the globe. The simplicity of this approach allows for a direct comparison between the modelled equilibrium climates that evolve using each of the different high latitude biome boundary conditions.

5. Results

Each of the experiments (Table 2) were run for 25–30 model years, providing more than ten years of equilibrium climate in each case. Two-metre screen temperature (degrees Celsius) and precipitation rate (mm per day) were averaged from the last ten years of each simulation over the terrestrial grid cells of the Antarctic continent (Table 3). Mean annual and mean seasonal data were analysed for the whole continent and each of the key regions separately (the Antarctic Peninsula and the Ross Sea regions). Mean austral summer (DJF) and winter (JJA) data were averaged from December/January/February (90 model days) and June/July/August (92 model days), respectively. Additional GCM fields analysed included surface-atmosphere sensible and latent heat fluxes, surface albedo, and sea ice and snow cover fractional extent and thickness. Areal averages were calculated for comparison between geographic areas (to account for the spatial shrinking of model grid cells near the South Pole); however, raw averages were used for significance testing. Student's t-test was applied to the model data to confirm significant differences between the EVER and TUND climatologies.

5.1. Antarctic climate sensitivity to vegetation boundary conditions

5.1.1. Preglacial ($3\times$ CO_2) and interglacial ($2\times$ CO_2) orbital experiments

Modelled mean annual temperature (MAT) data for the interglacial $2\times$ CO_2 scenarios indicate that a blanket TUND vegetation across Antarctica results in a significantly cooler climate over much of the landmass (on average by 1.2 °C) than if Antarctica was covered in EVER (Fig. 4, Table 3). In a warm preglacial climate ($3\times$ CO_2) with identical orbital configuration, mean annual temperatures are ~5 °C warmer than the cooler $2\times$ CO_2, interglacial scenario (Table 3). Both EVER and TUND interglacial climates (with relatively limited continental ice sheets and a warm summer orbital configuration) are above 0 °C on average during the summer (Fig. 5, Table 3). In the warm preglacial scenario, mean summer temperatures reach 7.4 °C in the EVER climate and 5.8 °C in the TUND climate. As expected, the influence of tundra versus evergreen forest vegetation evident in mean seasonal temperature comparisons is greatest during the spring and summer months. Significant summer continental-averaged cooling is 2.5 °C in the $2\times$ CO_2 interglacial scenario and 1.6 °C in the $3\times$ CO_2 scenario (Fig. 5). During austral summer,

Table 3
Model simulation results for near-surface temperature and precipitation rate averaged from 10 model years for the land area of Antarctica

			Near-surface temperature (°C)			Precipitation rate (mm/yr)		
			MAT	DJF	JJA	MAP	DJF	JJA
$2\times CO_2$:	Interglacial	TUND	−12.5	1.6	−20.9	423	86	109
		EVER	−11.3	4.1	−21.5	442	99	114
		Difference	*−1.2*	*−2.5*	*0.6*	*−19*	*−13*	*−5*
	Glacial	TUND	−26.8	−17.6	−33.4	219	58	54
		EVER	−25.1	−15.7	−31.9	318	60	80
		Difference	*−1.7*	*−1.9*	*−1.5*	*−99*	*−2*	*−26*
$3\times CO_2$:	Preglacial	TUND	−7.1	5.8	−15.0	475	97	127
		EVER	−6.1	7.4	−14.6	518	115	135
		Difference	*−1.0*	*−1.6*	*−0.4*	*−43*	*−18*	*−8*

TUND minus EVER results are quoted as 'Difference' and are all >99% significant based on Student's t-test analysis of the raw averages. EVER= needle leaf evergreen forest, TUND= tundra. MA*, mean annual; DJF, mean summer; JJA, mean winter; T, temperature; P, precipitation. All figures are areal averages taking into account the spatial shrinkage of model grid cell area towards the South Pole. All data is rounded to one decimal place.

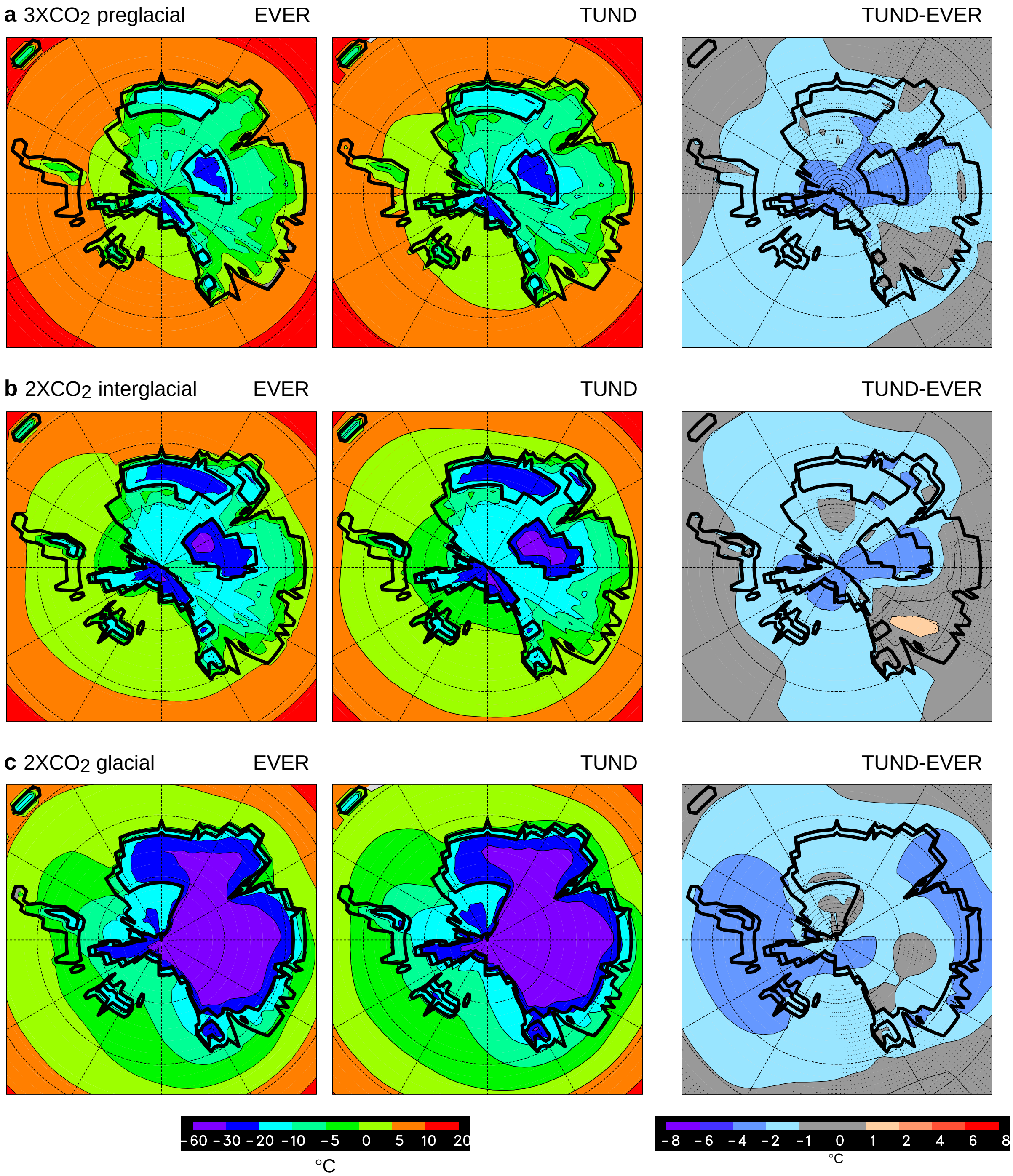

Fig. 4. Simulated mean annual two-metre surface temperature (°C) and differences for tundra (TUND) and evergreen (EVER) cases. (a) $3\times CO_2$, small ice sheets and a warm austral summer orbit, (b) $2\times CO_2$, medium ice sheets and a warm austral summer orbit, and (c) $2\times CO_2$, a large continental-scale ice sheet and a cold austral summer orbit are used to represent 'preglacial' late Eocene, 'interglacial' early Oligocene, and 'glacial' early Oligocene conditions, respectively. As in subsequent figures, thick black lines show the continental outline and interior ice sheet extent. Differences (right) are shown as the TUND case minus EVER to show the magnitude of tundra-induced cooling over high southern latitudes. Colour shading is limited to regions with temperature differences greater than 1 °C. Differences under non-stippled areas are statistically significant at the 95% confidence level as determined by Student's t-test.

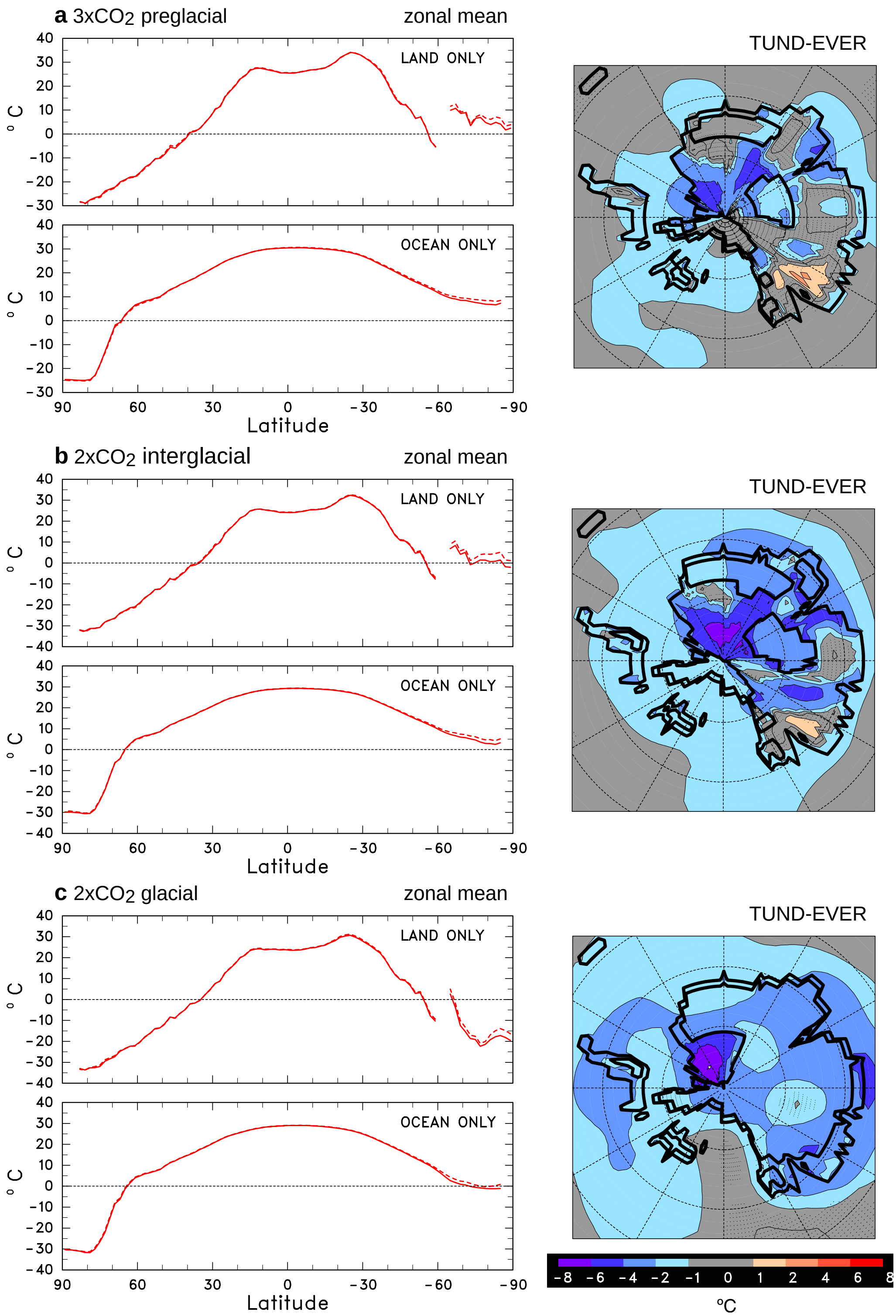

Fig. 5. Simulated austral summer (DJF) two-metre surface temperature (°C) shown as zonal means over land and ocean (left) and differences (right) for the EVER and TUND cases as in Fig. 4. Zonal mean temperatures show the direct influence of evergreen (dashed red line) versus tundra (solid red line) vegetation over ice-free continental regions, and indirect effects on high-latitude sea surface conditions.

cooling is >4 °C across large regions of the continental interior. This tundra-induced cooling effect is reflected in the much higher albedo of the Antarctic landmass in the TUND climate of both CO_2 scenarios (Fig. 6). During the Spring, particularly prior to snow melt, most of the ice-free areas have albedo values up to 0.8, whereas this decreases to 0.2 in the EVER climate (Fig. 6). This is consistent with prior palaeoclimate-

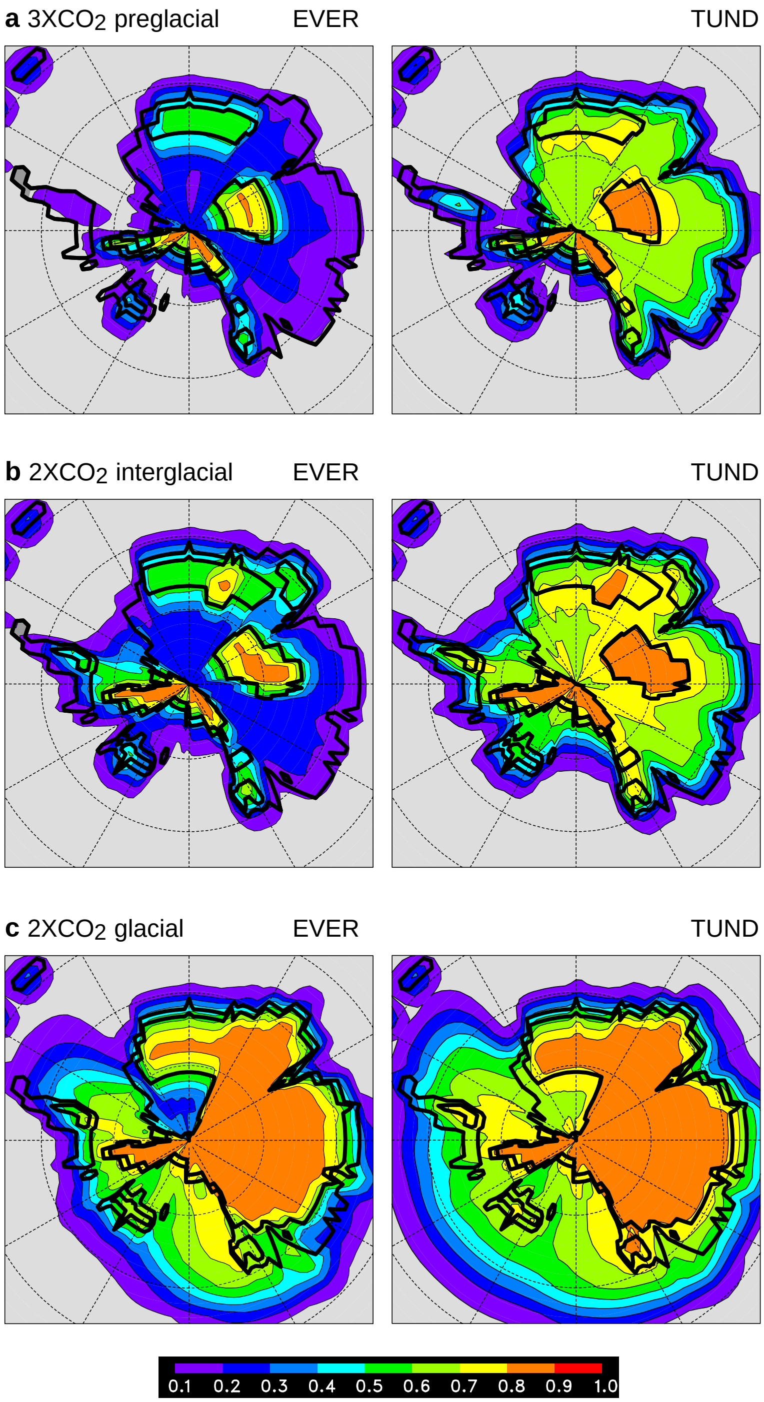

Fig. 6. Austral spring (SON) albedo calculated by the model in response to the distribution of spring snow cover, sea ice, and land surface conditions for the EVER (left) and TUND (right) cases.

vegetation studies (DeConto et al., 1999a,b; Dutton and Barron, 1996; Otto-Bliesner and Upchurch, 1997) where the effect of vegetation type on vegetation-atmosphere heat exchanges and land surface albedo is shown to be most prominent during the spring and summer months while the effect of roughness length on turbulent mixing in the boundary layer occurs all year. The albedo effects clearly dominate however, as reflected in the model results where the TUND cooling effect of up to 2.5 °C during the interglacial summer could have a potentially significant impact on the survivability of snow patches throughout the year and consequent ice sheet development. This is especially true in the 2× CO_2 interglacial case (Fig. 5b) where tundra-induced summer cooling maintains zonal mean Antarctic temperatures close to or just below zero. In contrast, summer temperatures in the EVER case are generally well above zero and therefore unfavourable for perennial snow cover and glaciation. In contrast to the summer cooling effect of tundra, during the winter months the TUND climate exhibits smaller, but statistically significant warming in the 2× CO_2 interglacial scenario (Table 3). In some areas the warming is between 2 and 4 °C. This is likely in response to the longer roughness length of the higher stature evergreen forest vegetation and its influence on turbulent mixing and sensible heat flux. This seasonal warming decreases the mean annual cooling effect of tundra vegetation across the Antarctic landmass (Fig. 4). Nonetheless, tundra-induced cooling clearly dominates and extends well offshore; decreased sea surface temperatures, especially during the summer and in the 2× CO_2 scenarios (Figs. 4 and 5), increasing seasonal sea ice extent and thickness in the seaway between East and West Antarctica (e.g., Fig. 7). The effect of tundra vegetation on Southern Ocean sea surface temperatures and sea ice combined with the seasonal thermal inertia of the ocean extends the zonal mean cooling effect into the winter

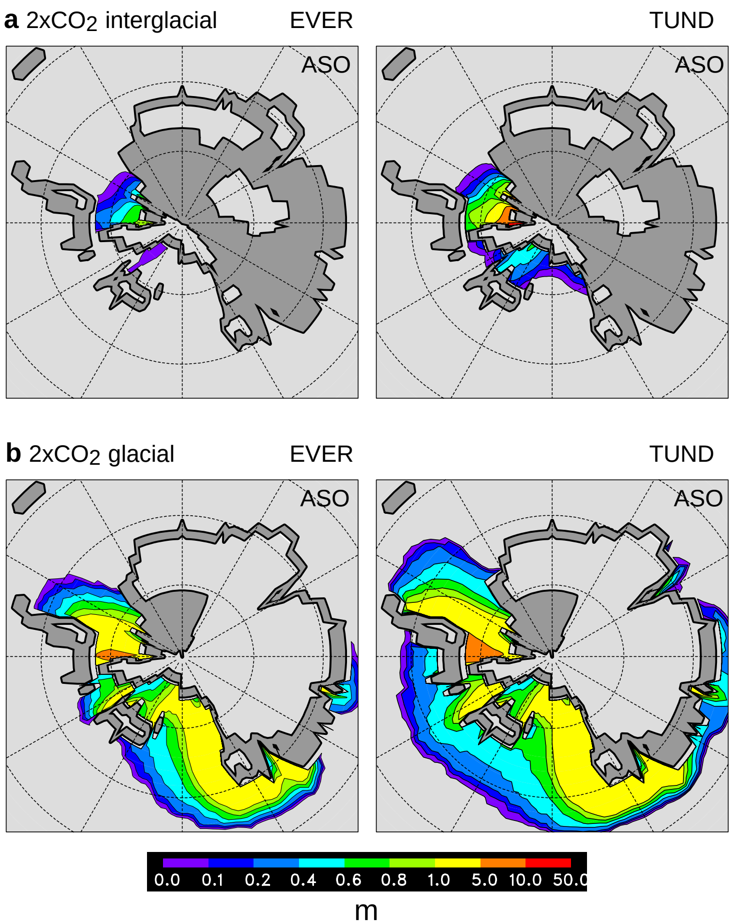

Fig. 7. Simulated, maximum austral spring (ASO) sea ice thickness (metres) for the interglacial (a) and glacial (b) 2× CO_2 scenarios, with EVER and TUND continental vegetation. No seasonal sea ice forms in the 3× CO_2 cases (not shown).

months (JJA) despite the lack of winter insolation and vegetation-albedo feedbacks and despite the relative winter-warming effects of tundra vegetation.

Regionally, the influence of vegetation type on modelled climate in the more maritime Antarctic Peninsula appears less prominent than in the Ross Sea region where continental influences dominate. The Ross Sea region, in both EVER and TUND 2× CO_2 interglacial climates, is significantly cooler than the Antarctic Peninsula with an approximate 8.5 °C difference in MAT. Mean annual temperatures in the tundra climate are significantly cooler than the evergreen forest climate throughout the year in both regions. In the Antarctic Peninsula both EVER and TUND climates (−2.0 and −3.0 °C MAT, respectively) are substantially warmer than the mean annual temperatures for the entire continent (−12.5 and −11.3 °C) with both summer and winter temperatures within 6.5 °C of zero. In the 3× CO_2 preglacial scenarios, mean annual and seasonal temperatures at the Antarctic Peninsula are again significantly warmer than both the Ross Sea region and Antarctica as a whole. Mean annual temperature even in the TUND climate is above zero (1.7 °C) with mean summer temperatures reaching 8.8 °C in the EVER climate. The Ross Sea region mean annual temperatures for both the EVER and TUND climates mirror that of the whole continent (−6.1 and −7.1 °C, respectively), although the mean summer temperatures are significantly colder than the continental average.

Over the Antarctic landmass in all interglacial and preglacial scenarios the annual precipitation rate is moderate, between 1.16 and 1.42 mm/day (~400–500 mm p.a.) (Table 3), and is within the range of a modern boreal forest (taiga) biome with cold winters (when precipitation falls as snow) and warm, humid, but short summers. Overall, mean annual precipitation is higher in the preglacial 3× CO_2 climates than the interglacial 2× CO_2 climates and lower in the TUND compared with the EVER scenarios, although in general there is little significant difference (Fig. 8). During the 2× CO_2 interglacial scenarios the winter season is slightly wetter with most of this precipitation falling as snow. The influence of vegetation type on Antarctic mean annual precipitation rate is greatest in the 3× CO_2 preglacial climate with the TUND climate having 0.12 mm/day (43 mm p.a.) less on average than the EVER climate. In contrast, only 0.05 mm/day (19 mm p.a.) less on average falls with tundra vegetation in the 2× CO_2 interglacial climate. Higher average regional precipitation in the Antarctic Peninsula and Ross Sea region than the continent-wide average is consistent with coastal regions being wetter

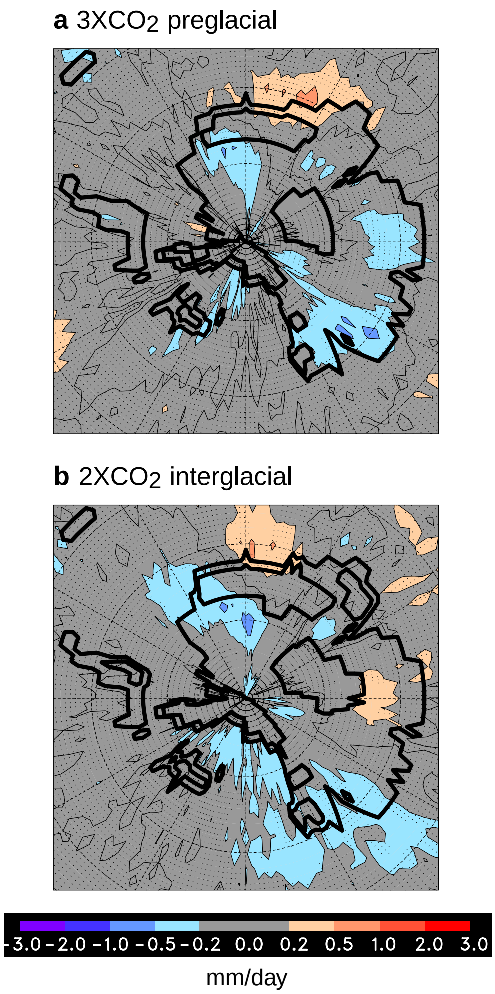

Fig. 8. Differences (TUND minus EVER) in simulated mean annual precipitation rate (mm/day). Precipitation differences are shown for (a) the 3× CO_2 and (b) 2× CO_2 interglacial cases described in the text. Colour shading is limited to regions with precipitation differences greater than 0.2 mm/day. Differences under non-stippled areas are statistically significant at the 95% confidence level as determined by Student's t-test.

than the interior, a situation relevant to 34 Ma as well as modern geography.

Mean annual sensible and latent heat fluxes over the whole of Antarctica during both the 2× CO_2 interglacial and 3× CO_2 scenarios are generally upward (positive) and significantly lower in the TUND than the EVER climate (e.g., Fig. 9). As expected, significant seasonal variability in sensible heat flux occurs in both the TUND and EVER climates with higher positive flux during the summer. During the summer months, tundra vegetation particularly in the interior of continental East Antarctica, causes a significantly lower upward sensible heat flux than evergreen forest due in part to the lower temperatures of the single lower

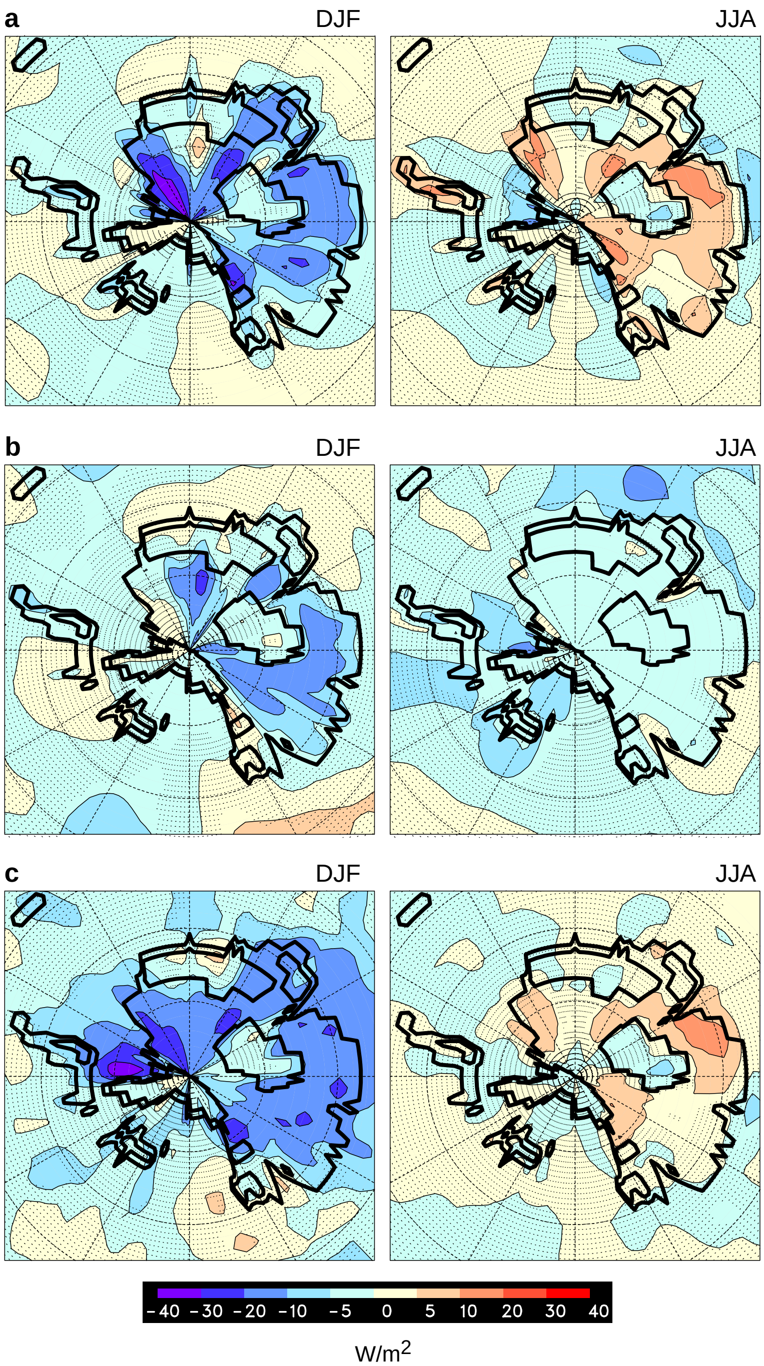

Fig. 9. Differences in simulated seasonal heat and radiative fluxes (W/m^2) for the 2× CO_2 interglacial case with EVER versus TUND vegetation. Differences, shown as TUND minus EVER, are calculated for (a) upward sensible heat flux, (b) upward latent heat flux, and (c) downward net radiation at the surface. Differences under non-stippled areas are statistically significant at the 95% confidence level as determined by Student's *t*-test.

canopy. Maximum summer values of over 50 W/m^2 occur in the ice-free region east of the Pensacola Mountains in the EVER climate, however, as expected the offshore influence of vegetation type on heat flux is insignificant except in places where it influences the presence or absence of sea ice. Comparing sensible heat flux between the EVER and TUND climates indicates a much greater effect in the summer than the winter months in both the 2× CO_2 interglacial and 3× CO_2 cases, with reductions in sensible heat flux >30 W/m^2

in some regions. The model results for both CO_2 cases and vegetation types yield a lower latent heat flux during these winter months consistent with a lack of evapotranspiration during the Polar Night and a lower specific humidity gradient. In all of the TUND cases latent heat flux is significantly reduced compared with the EVER case in every season particularly during summer over ice-free regions (e.g., Fig. 9). Vegetation type has the largest influence on latent heat flux in the 3× CO_2 scenario due to the higher rates of evapotranspiration associated with a warmer environment.

Vegetation-albedo/cooling effects on downward net surface radiation are clearly seen in the 2× CO_2 interglacial simulations (Fig. 9c) where net radiation is reduced by 10–40 W/m^2 over most of ice-free continental interior and over ocean regions with increased sea ice fractional cover and thickness.

5.1.2. Glacial (2× CO_2) orbit experiments

As expected the simulated glacial climate conditions in the south polar region for both EVER and TUND scenarios are much colder and drier than the preglacial and interglacial experiments (Table 3, Figs. 4 and 5). The glacial climate mean annual temperatures are ~14 °C cooler than the interglacial climate for both vegetation types with a much higher overall albedo due primarily to the direct snow-masking effect of a more extensive ice sheet (Fig. 6). The magnitude of the difference between the glacial and interglacial EVER and TUND temperatures is similar, but with a slightly different response to the change in Antarctic vegetation. Overall, the influence of tundra vegetation in a glacial scenario cools the mean annual temperature of the Antarctic landmass by 1.7 °C. Surprisingly this is slightly more than in the interglacial climate (1.2 °C) due to a continued tundra cooling effect during the winter. During the summer months when temperatures are important with respect to the survivability of snow and ice, tundra cools the land area significantly, by 1.9 °C on average, in comparison to evergreen forest although this effect is less compared to the interglacial climate (2.5 °C cooling) due to a smaller vegetated ice-free area (Fig. 5). The maximum effect of tundra vegetation on the glacial climate occurs during the spring with 2.2 °C of cooling compared to the evergreen forest scenario across Antarctica. This slight reduction in the influence of vegetation type on the glacial model climate compared to the interglacial climate is expected due to a more extensive ice sheet, generally colder temperatures and less land area available for vegetation influences. In the natural world, colder glacial climate would reduce the geographic range of vegetation cover and therefore evapotranspiration rates and latent heat flux between the land surface and the atmosphere. As in the interglacial climate the tundra cooling effect in the glacial scenario decreases Southern Ocean sea surface temperatures significantly and increases the fractional cover of sea ice, which forms in both EVER and TUND climates mostly in the seaway between East and West Antarctica and around the Victoria Land coast (Figs. 4 and 7).

Regionally, mean annual temperatures on the Antarctic Peninsula are ~18 °C warmer in both EVER and TUND scenarios (−6.2 and −8.8 °C, respectively) than the mean annual temperature for the entire Antarctic continent (−25.1 and −26.8 °C). In the Ross Sea region, mean annual temperatures are again more influenced by continental effects and are much closer to the continental average at −21.8 °C (EVER) and −23.2 °C (TUND). Mean summer temperatures in the Ross Sea for both vegetation types are −12 °C (EVER) and −13 °C (TUND) compared to the near zero mean summer temperatures of the Antarctic Peninsula region at 0.1 and −1.6 °C. Mean winter temperatures plummet to −30 °C for both EVER and TUND climates in the Ross Sea region and to −11.3 °C (EVER) and −14.7 °C (TUND) in the Antarctic Peninsula.

Precipitation rates are significantly lower in the glacial climates compared with the interglacial and preglacial climates (Table 3). In addition, the glacial EVER climate is wetter than the glacial TUND climate by 0.27 mm/day (99 mm p.a.) mean precipitation, however, summer precipitation remains similar in each of the different vegetation cases. The highest precipitation rates occur during the austral autumn. Model mean annual precipitation data suggests a wetter glacial climate on the maritime Antarctic Peninsula (400 mm p.a., EVER; 274 mm p.a., TUND) compared to the mean annual precipitation rate for the whole of the Antarctic continent (318 mm p.a.; 219 mm p.a.).

5.2. Model versus proxy palaeoclimate

This study focuses on the model data for the terrestrial areas south of 60° Latitude. Additional information from model-data comparisons could be derived in the future from expanding the analysis north with additional proxy data without the need to run additional simulations.

To compare the quantitative palaeoclimate interpretations from the Antarctic fossil plant record spanning the E/O boundary (Table 1, Fig. 2) with the modelled 34 Ma climates for the terrestrial regions of Antarctica, the model data need to be simplified. The six model experiments represent contrasting preglacial (3× CO_2) and glacial/interglacial (2× CO_2) scenarios with distinct ice sheet and orbital configuration boundary conditions based on prior coupled GCM-ice sheet simulations across the Oi1 transition (DeConto and Pollard, 2003a). While we use idealised orbital values within the assumed range of Cenozoic eccentricity and obliquity (Berger and Loutre, 1991; Laskar et al., 2004) it is supposed that average Antarctic conditions during the earliest Oligocene consisted of some combination of the 2× CO_2 model climatologies. Further, the two different simple vegetation boundary conditions (EVER and TUND) used to test the model sensitivity to Antarctic vegetation type are necessarily over-simplistic compared with a more realistic distribution of earliest Oligocene Antarctic biomes. Differences in environmental conditions across the landscape, e.g., in altitude, aspect, soil type and drainage, would have occurred as they do in modern vegetated environments. This would have resulted in a mosaic of vegetation communities across the ice-free areas of the continent (if the climate was warm enough for vegetation growth) each with their own level of temperature and moisture feedback to the climate. Therefore, the EVER and TUND model biomes for all paired scenarios are combined for proxy comparative purposes. The model data are therefore simplified by averaging model temperature and precipitation data for both the 2× CO_2 and 3× CO_2 scenarios (to incorporate mixed vegetation, ice sheet and orbital configurations) to produce two sets of summary figures for comparison with the available proxy data (Table 4).

The range of quantitative palaeoclimate temperature and precipitation figures available from studies on the Antarctic fossil plant record from the latest Eocene through the Oligocene floral 'transition zone' is summarised in Table 4 (refer also to Table 1 and Fig. 2). The results indicate that mean annual temperature in the combined 2× CO_2 scenario models is ~7 °C colder than the minimum proxy temperature and −20 °C below the mid-point. Mean summer and especially winter temperatures reach 24 °C colder than the proxy minimum. A similar comparison occurs within the key regional areas with simulated mean annual temperatures closest to the proxy range in the Ross Sea region, which are ~5 °C colder than the proxy minimum. In the model, mean annual temperatures on the Antarctic Peninsula are 10 °C colder than the proxy data. However, the averaged 3× CO_2 model temperatures are in good agreement with the proxy record and are *within* the proxy temperature range for mean annual temperature (at 5 °C warmer than the minimum) and mean summer temperature (3 °C warmer than the minimum) continent-wide. Simulated mean winter temperatures are again significantly colder than the proxy range, however, the minimum mean winter proxy temperatures may represent even lower temperatures due to uncertainties in the modern analogue interpretation. Regionally, the 3× CO_2 mean annual temperatures also agree well with the proxy record. In the Ross Sea region the model mean annual temperature is within the proxy range and it is only 3 °C colder than the minimum proxy temperature in the Antarctic Peninsula region. Simulated mean summer temperatures are only 3 °C colder in the Ross Sea region.

Both the 2× CO_2 and 3× CO_2 simulations appear to significantly underestimate the precipitation rate in the Antarctic Peninsula region (Table 4) suggesting that the modelled climate is too dry in the coastal regions of Antarctica. However, the range of modern analogue data for this climatic parameter is broad and we only

Table 4
Comparison of combined results from the model simulations with quantitative palaeoclimate data summarised from the Antarctic latest Eocene to Oligocene plant fossil record

		Whole of Antarctica			Ross Sea region			Antarctic Peninsula region			
		Near-surface temperature (°C)									Precipitation rate (mm)
		MAT	DJF	JJA	MAT	DJF	JJA	MAT	DJF	JJA	MAP
Model summary[1]	2 × CO_2	−19	−7	−27	−17	−8	−23	−5	1	−10	350
	3 × CO_2	−7	7	−15	−7	1	−11	2	8	−1	496
Proxy range[2]		−12 to 15	4 to 12	< −3 to 7	−12 to 5	4 to 12	< −3 to 7	5 to 15	nd	nd	600–4300
Combined model data ([1]) minus proxy data minimum ([2]):	2 × CO_2	−7	−11	< −24	−5	−12	< −20	−10	nd	nd	−250
	3 × CO_2	5	3	< −12	5	−3	< −8	−3	nd	nd	−104

Model figures are derived from areally averaged data across the Antarctic continent from each of the six experiments as described in the text; combining the effects of the different prescribed boundary conditions. For sources of proxy data refer to Table 1. Model figures are rounded to the nearest integer for ease of comparison with the proxy data. nd= no data.

have comparable proxy data from the Antarctic Peninsula, which the modelling suggests has a significantly different climate to the Ross Sea region and the continental averages. Furthermore, medium-resolution spectral GCMs do a poor job representing the steep orographic slopes of coastal margins especially with regards to precipitation. Therefore, in terms of precipitation an estimate of the model's success over the broader Antarctic remains equivocal.

Throughout these model-proxy data comparisons it must be emphasised that the comparison of model data with the proxy record relies heavily on the accuracy of palaeoclimate interpretations from fossil plant assemblages. In the Antarctic these are sparse and have relatively low diversity in the floral 'transition zone' across the E/O boundary. So, a larger proxy dataset would significantly improve model calibration and boundary conditions. Palaeobotanical techniques are becoming increasingly sophisticated for extracting quantitative information from fossil plants, which can only serve to improve the empirical testing of GCMs in the future.

6. Discussion

Altering vegetation type across Antarctica between a needle leaf evergreen forest and low stature tundra vegetation has important effects on all the simulated climates. The significant cooling effect of tundra versus evergreen forest vegetation of up to 2.5 °C during the interglacial summer months supports the importance of vegetation feedbacks during glaciation transition seen in previous studies (e.g., Bonan et al., 1992; DeConto and Pollard, 2003a; Dutton and Barron, 1996; Robinson and Kukla, 1985). With a predominantly evergreen-forested Antarctica a significant warming effect could potentially delay the initiation of a continental-scale ice sheet due to enhanced summer ablation. Quantification of this potential delay on ice sheet initiation due to vegetation type is currently being analysed in coupled GCM-ice sheet-vegetation simulations for the same period. Vegetation-climate feedbacks would presumably become even more significant for Antarctic palaeoclimate in model scenarios closer to the glaciation threshold level of greenhouse gases, suggested by previous studies to be ~2.4– 2.8 × CO_2 (DeConto and Pollard, 2003b). Thus, the implications of significant vegetation-climate feedbacks are potentially crucial for coupled GCM-ice sheet modelling. For example, the cooling effects of tundra recognised in this study suggest that a prescribed 100% tundra land surface could allow an ice sheet to form at relatively high levels of atmospheric greenhouse gas concentrations. It is also noteworthy that the seasonal cooling in the continental interior caused by a replacement of evergreen vegetation with tundra is greater than that caused by the opening of the Tasmanian Gateway in a coupled AOGCM simulation of the Eocene (Huber et al., 2004) and a 20% reduction in poleward ocean heat transport in a slab-ocean GCM simulation of the Eocene–Oligocene boundary (DeConto and Pollard, 2003b). This suggests that climate feedbacks related to shifting terrestrial ecosystems may have been as important during the Oi1 transition as the effects of opening Southern Ocean gateways, long assumed to be the most important contributor to Palaeogene Antarctic cooling (Kennett, 1977).

Prior coupled GCM-ice sheet model simulations suggesting the CO_2 threshold for widespread Antarctic glaciation was ~2.4– 2.8 × CO_2 used the same atmospheric model component as this study (DeConto and Pollard, 2003b). While climate sensitivity to CO_2 is model-dependent, comparison of these results using 2× and 3× CO_2 with our proxy estimates from the vegetation record gives us some idea as to the reasonableness of simulated Antarctic climate around the time of the Greenhouse–Icehouse transition.

Comparing the average temperatures for the whole of Antarctica from each of the model experiments to the proxy fossil plant record and then combining the effects of vegetation type and orbital configuration, the 2× CO_2 model climates are significantly cooler, while the 3× CO_2 climate agrees well with the proxy temperature ranges. The growing conditions suggested for the survival of *Nothofagus beardmorensis*, a prostrate woody beech species described from the Ross Sea region Sirius Group deposits (not included in the proxy record for this study due to disputed age control), include a minimum winter temperature of −22 °C and a mean annual temperature of −12 °C (Cantrill, 2001). The GCM 2× CO_2 interglacial temperatures are at the minimum required for survival of this extinct species with a mean annual temperature (continent-wide) of −11.3 °C (EVER) and −12.5 °C (TUND) and mean winter temperatures of −21.5 and −20.9 °C, respectively. Regionally, these criteria for growth are only comfortably met in the Antarctic Peninsula during these interglacial times suggesting that distribution of these low-growing (in comparison to tall stature trees) woody plants would be severely restricted over much of the continent with only the relatively far north and maritime Antarctic Peninsula providing a suitable climate for growth. It is interpreted, however, from the fossil record of the Ross Sea region that *Nothofagus* and other woody vegetation such as podocarps were significant components of the regional vegetation at the time of ice sheet initiation and may have formed a low scrub or closed forest of intermediate stature between that of the Eocene and the limited late Oligocene to early Miocene vegetation (Raine and Askin, 2001). The preglacial 3× CO_2 model data, however, even in the TUND simulations provide more equable temperatures continent-wide and are well within the suggested growing conditions for *Nothofagus beardmorensis*. Mean annual temperatures of −6.1 °C (EVER) and −7.1 °C (TUND) are ~5 °C above the suggested mean annual temperature criteria and mean winter temperatures are ~7 °C above the minimum suggested. The Antarctic Peninsula region may well have even been too warm for this prostrate species with mean annual temperatures above zero and mean summer temperatures reaching ~9 °C in the EVER climate. The fossil record suggests the presence of a *Nothofagus*-dominated forest with conifers, ferns, Proteaceae and other angiosperms growing during the Eocene to earliest Oligocene in the Antarctic Peninsula region (Askin, 1997).

The GENESIS GCM therefore appears to be simulating too cold a climate at 34 Ma with 2× CO_2 to support significant woody vegetation across Antarctica as seen in the empirical record. However, temperatures more comparable with the fossil record are simulated in the 3× CO_2 scenario suggesting that the GCM is producing realistic climates for the period prior to the Oi1 transition.

There may be several reasons for cooler than expected climates simulated for the earliest Oligocene interglacial scenario, which might also affect the accuracy of the 3× CO_2 simulations. These include our choice in prescribed ice sheet configurations, orbital parameters, or issues with the comparative proxy information. For example, only two orbital configurations out of a possible five from DeConto and Pollard's (2003b) synthetic 40 kyr orbital cycle were chosen for this study — a 'cool, interglacial' and a 'cold, glacial' scenario. While producing warmer austral summer temperatures than our glacial climate scenarios we note that the orbital parameters used to represent 'interglacial' conditions are on the cool side of the full range of possible Cenozoic orbital configurations and likely contribute a cold bias to our model results. Considering the sensitivity of Antarctic climate to the prescribed boundary conditions used in these experiments, additional simulations should be run using a broader representation of orbital parameters to see if any of the resulting model climates were closer again to the proxy palaeoclimate data from fossil plants.

In addition, the static vegetation boundary conditions forming the mainstay of the experimental design relied on the choice of appropriate biome categories. Within the limitations of the scheme, however, it was not possible to recognize the broad-leaved or deciduous components of the 34 Ma vegetation recognized in the Antarctic fossil plant record. This may have a significant effect on the modelled climate due to seasonal changes in albedo and surface to atmosphere heat exchanges. Future refinements to these experiments should therefore include additional vegetation biomes reflecting more accurately the communities in the fossil record. In addition, enhanced flexibility over the resolution of community distribution, not only across the south polar landmass, but the remaining terrestrial regions of the globe would improve vegetation boundary conditions. Dynamic coupling of a complex vegetation model with the GCM would let the modelled climate and vegetation communities evolve together allowing a more complete analysis of vegetation-climate feedbacks and richer model-data comparisons.

Further, inherent assumptions in model boundary conditions for which no fossil record exists have to be based by necessity on modern or poorly constrained values. An obvious example is our reconstruction of global palaeogeography and more specifically, ancient Antarctic palaeotopography. Similarly, the prescription of ice shelves are not included in the model simulations, which could result in even colder modelled climates over the regions where the model and data are compared. However, the existence and extent of early Oligocene ice shelves is unknown.

Despite the uncertainties discussed above the results from this study suggest that the GENESIS GCM successfully simulated Antarctic climate for conditions representative of a time just prior to the development of a permanent East Antarctic ice sheet ~34 Ma. Furthermore, the choice of 2× and 3× CO_2 experiments does appear to span the presumed range of CO_2 values around the time of this Oi1 transition lending some confidence in the assumptions made in prior coupled GCM-ice sheet simulations of the event (e.g., DeConto and Pollard, 2003b).

7. Summary

This study contributes to the investigation of palaeoclimate during the initiation of the East Antarctic Ice Sheet ~34 Ma by comparing GCM simulations with contemporaneous information derived from the proxy fossil plant record. Covering the polar landmass of Antarctica with first needle leaf evergreen trees then tundra tested the sensitivity of Antarctic climate to different simple static vegetation boundary conditions. The polar vegetation categories were chosen to reflect the change in composition of the fossil plant assemblages throughout a 'transition zone' described across the E/O boundary from an evergreen mixed broadleaf and needle leaf temperate forest to a tundra/low shrub community. Preglacial, interglacial and glacial model scenarios were modelled using idealized orbital parameters and related ice sheet configurations, 2× and 3× pre-industrial atmospheric levels of CO_2 and 34 Ma palaeogeography.

Simulated preglacial and interglacial climates produced mean summer temperatures above zero degrees Celsius continent-wide with the preglacial 3× CO_2 climate ~5 °C warmer on average. Precipitation is moderate across the continent at 400–500 mm p.a. in these scenarios and comparable to the modern day taiga biome. Regionally, the Ross Sea region mean temperatures compare closely to continent-wide means, however, the Antarctic Peninsula models significantly warmer than continental averages as expected due to its more northerly position and maritime influences. The glacial climates (2× CO_2) are significantly colder and drier than the preglacial and interglacial scenarios.

The importance of high latitude vegetation-atmosphere feedbacks to modelled palaeoclimate in the south polar region has been demonstrated by the observation of statistically significant differences in near-surface temperature, precipitation rate and surface-atmosphere heat exchanges over the Antarctic landmass. Preglacial and interglacial mean annual temperatures across Antarctica are on average ~1 °C cooler, ~4 °C cooler in ice-free areas, with a tundra rather than a needle leaf evergreen forest vegetation covering the continent. The influence of vegetation type is greatest during the spring and summer with continental cooling up to 2.5 °C with tundra compared with evergreen forest vegetation in the interglacial scenario. The large effect on summer temperature would likely have significant implications for the development of a continental-scale ice sheet in a scenario of global cooling as presumed to have occurred around the Eocene–Oligocene boundary. The tundra cooling effect is mostly due to the enhanced albedo of tundra cover compared to forest, particularly in snow-covered regions. Sensible and latent heat fluxes are generally lower with tundra vegetation across the continent in all simulations, as is net downward radiation. In austral winter in the interglacial scenario, the tundra case is slightly warmer than the evergreen forest case, however, the winter warming is more than offset by the summer cooling effect. Southern hemispheric winter temperatures remain cooler overall in the tundra case despite the regional winter warming. This appears to be due to the seasonal lag of high latitude sea surface temperatures, which remain cooler in the tundra than in the evergreen forest case throughout the year. The tundra vegetation simulations also exhibit significant expansion in coastal sea ice, further enhancing and spatially propagating the albedo-related cooling feedbacks associated with the spread of tundra.

The glacial (cool austral summer orbits and bigger continental ice sheets) climates are colder and drier than the preglacial and interglacial climates with significantly different model sensitivity to the vegetation boundary conditions. At 2× CO_2, the temperatures suggest that woody vegetation may not survive across most of the continent, perhaps being restricted to the relatively mild Antarctic Peninsula region. However, prior coupled GCM-ice sheet simulations have shown that the continent would be fully glaciated at 2× CO_2 anyway. Sensible heat fluxes are similarly lower with south polar tundra in the glacial model as in the interglacial, but latent heat fluxes with both vegetation types are significantly lower in the glacial scenario. Modelled mean annual precipitation for interglacial and glacial experiments is comparable to the modern taiga biome.

Quantitative palaeoclimate information is summarised from the literature for fossil plant assemblages described from the Antarctic Peninsula and Ross Sea regions and divided into three zones: Late Palaeocene to the end of the Middle Eocene– Evergreen, mixed broadleaf and needle leaf temperate forests; Late Eocene to earliest Miocene-transition: Low-diversity forest to shrubland to tundra/low shrub; and Early Miocene–Tundra/low shrub. These vegetation zones reflect an overall climatic cooling throughout the late Palaeocene to earliest Miocene. Climate model temperature and precipitation data from the four experiments are summarised and combined to reflect more realistic mixed vegetation and orbital boundary conditions and compared with the proxy ranges derived from modern analogue analysis. The 2× CO_2 model results are shown to be much colder than the minimum proxy temperatures (on average by 7 °C) continent-wide, but particularly during the winter (by 24 °C). Further, the Antarctic Peninsula precipitation rate is 250 mm p.a. less than the minimum proxy data suggests. In contrast, the 3× CO_2 model results are in good agreement with the proxy data, so providing a realistic Antarctic climate for conditions just prior to the Oi1 transition. Mean annual (−7 °C) and mean summer (7 °C) temperatures continent-wide are within the suggested empirical ranges (−12 to 15 °C and 4 to 12 °C, respectively). Regionally, modelled mean annual temperature for the Ross Sea region (−7 °C) is also within the proxy data range (−12 to 5 °C) and with comparable figures modelling only slightly colder (3 °C) than the minimum proxy temperatures in the Antarctic Peninsula. The closest model scenario to the proxy record of palaeoclimate is the preglacial, evergreen forest simulation, although modelled winter temperatures are still significantly colder. This 3× CO_2 case provides comfortable growing conditions suggested for woody vegetation across the continent, even in the tundra climate, based on suitable criteria interpreted for the survival of *Nothofagus beardmorensis* from the Sirius Group. These results remind us that model data and proxy data comparisons rely on modern analogue interpretations from the sparse Antarctic fossil record that are difficult to constrain due to the inherent differences, despite apparent similarities in taxonomy and diversity, between vegetation growing ~34 Ma and in the present day.

The two different CO_2 levels applied to the 34 Ma simulations in this study were chosen to represent assumed greenhouse gas levels in the atmosphere prior to (3× CO_2, 'Greenhouse') and following (2× CO_2, 'Icehouse') the Oi1 transition. These chosen levels were based on prior work, which identified a critical glaciation threshold for this period at atmospheric levels of CO_2 between 2.5 and 3 times preindustrial levels (280 ppm). Certainly, within this range of CO_2 concentrations, vegetation type appears to have a significant effect on Antarctic continental climate even if CO_2 is the dominant contributing factor to climate change at the E/O boundary as indicated by prior modelling studies (e.g., DeConto and Pollard, 2003b; Huber et al., 2004). Interestingly, vegetation type appears to have a greater effect, on temperature in particular, in the glacial 2× CO_2 case compared with the preglacial and interglacial orbital forcing scenarios. Vegetation type may have an even more pronounced effect on climate, particularly on snow accumulation rates, closer to the critical CO_2 level for ice sheet initiation at the E/O transition, which in this GCM may be nearer 2.8×CO_2. The climatic influence of vegetation cover on Antarctic interior climate is greater than that shown in model simulations of open versus closed Southern Ocean gateways, suggesting climate-vegetation feedbacks may have played an important role in the Oi1 transition and subsequent Cenozoic climatic events on Antarctica.

Future work will involve the interactive coupling of dynamic vegetation models with higher resolution coverage of diverse biomes and additional investigation of the tundra cooling effect. Quantification of the potentially important vegetation-climate feedbacks on Palaeogene ice sheet development described in this study awaits the results of additional modelling using coupled climate-ice sheet-vegetation models. This will test the cooling feedback effects of a needle leaf evergreen forest to tundra vegetation transition across Antarctica on the timing of the initiation of an extensive East Antarctic Ice Sheet. The results from this future work will allow a better assessment of the relative importance of vegetation-climate feedbacks compared with previously identified critical forcing factors such as greenhouse gas concentrations, orbital forcing, ocean gateways (e.g., the Drake and Tasmanian Passages) and mountain uplift.

Acknowledgements

We would like to acknowledge funding and support for this project from the New Zealand Foundation for Research, Science and Technology as part of Thorn's Postdoctoral Fellowship, the Royal Society of New Zealand Marsden Fund for contributions from the project 'The dynamic behaviour of Antarctic ice sheets on a warmer planet' and the Antarctic Research Centre, Victoria University of Wellington, New Zealand. This material is also based upon work supported by collaborative awards from the National Science Foundation (ATM-9905890/9906663 and ANT-0342484). We also thank David Pollard for his ongoing support of GENESIS GCM modelling activities.

References

Ashworth, A.C., Cantrill, D.J., 2004. Neogene vegetation of the Meyer Desert Formation (Sirius Group) Transantarctic Mountains, Antarctica. Palaeogeogr. Palaeoclimatol. Palaeoecol. 213, 65–82.

Askin, R.A., 1992. Late Cretaceous–early Tertiary Antarctic outcrop evidence for past vegetation and climates. Antarct. Res. Ser., AGU 56, 61–73.

Askin, R.A., 1997. Eocene–?earliest oligocene terrestrial palynology of Seymour Island, Antarctica. In: Ricci, C.A. (Ed.), The Antarctic Region: Geological Evolution and Processes. Terra Antart. Publication, Siena, pp. 993–996.

Askin, R.A., 2000. Spores and pollen from the McMurdo Sound Erratics, East Antarctica. In: Stilwell, J.D., Feldmann, R.M. (Eds.), Paleobiology and Paleoenvironments of Eocene rocks, McMurdo Sound, East Antarctica, Ant. Res. Ser., vol. 76. AGU, Washington, D.C., pp. 161–181.

Askin, R.A., Raine, J.I., 2000. Oligocene and early Miocene terrestrial palynology of the Cape Roberts drillhole CRP-2/2A, Victoria Land Basin, Antarctica. Terra Antartica 7 (4), 493–501.

Axelrod, D.I., 1984. An interpretation of cretaceous and tertiary biota in the polar regions. Palaeogeogr. Palaeoclimatol. Palaeoecol. 45, 105–147.

Bamber, J.A., Bindschadler, R.A., 1997. An improved elevation dataset for climate and ice-sheet modelling: validation with satellite imagery. Ann. Glaciol. 25, 439–444.

Bergengren, J.C., Thompson, S.L., Pollard, D., DeConto, R.M., 2001. Modeling global climate-vegetation interactions in a doubled CO_2 world. Clim. Change 50, 31–75.

Berger, A., Loutre, M.F., 1991. Insolation values for the climate of the last 10,000,000 years. Quat. Sci. Rev. 10, 297–317.

Birkenmajer, K., 1997. Tertiary glacial/interglacial palaeoenvironments and sea-level changes, King George Island, West Antarctica. An overview. Bull. Pol. Acad. Sci., Earth Sci. 44, 157–181.

Birkenmajer, K., Zastawniak, E., 1986. Plant remains of the Dufayel Island group (early Tertiary?), King George Island, South Shetland Islands (West Antarctica). Acta Palaeobot. 26, 33–54.

Birkenmajer, K., Zastawniak, E., 1989. Late Cretaceous–early Tertiary floras of King George Island, west Antarctica: their stratigraphic distribution and palaeoclimatic significance, Origins and Evolution of the Antarctic Biota. Spec. Publ.- Geol. Soc. Lond.- Geol. Soc. Lond, London, pp. 227–240.

Bonan, G.B., 1998. The land surface climatology of the NCAR land surface model coupled to the NCAR community climate model. J. Climate 11 (6), 1307–1326.

Bonan, G.B., Pollard, D., Thompson, S.L., 1992. Effects of boreal forest vegetation on global climate. Nature 359, 716–718.

Bonan, G.B., Chapin, F.S., Thompson, S.L., 1995. Boreal forest and tundra ecosystems as components of the climate system. Clim. Change 29 (2), 145–167.

Brotchie, J.F., Sylvester, R., 1969. On crustal flexure. J. Geophys. Res. 74 (22), 5240–5252.

Cantrill, D., 2001. Early oligocene *Nothofagus* from CRP-3, Antarctica: implications for the vegetation history. Terra Antartica 8 (4), 401–406.

Case, J.A., 1988. Paleogene floras from Seymour Island, Antarctic Peninsula, geology and paleontology of Seymour Island, Antarctic Peninsula. Mem. Geol. Soc. Am.Geol. Soc. Am, pp. 523–530.

Cox, P.M., 2001. Description of the "TRIFFID" Dynamic Global Vegetation Model. Hadley Centre Technical Note, vol. 24. Hadley Centre, Meteorological Office, Bracknell, Berkshire.

DeConto, R.M., Pollard, D., 2003a. A coupled climate–ice sheet modelling approach to the early Cenozoic history of the Antarctic ice sheet. Palaeogeogr. Palaeoclimatol. Palaeoecol. 198 (1), 39–52.

DeConto, R.M., Pollard, D., 2003b. Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO2. Nature 421, 245–249.

DeConto, R.M., Hay, W.W., Thompson, S.L., Bergengren, J.C., 1999a. Late Cretaceous climate and vegetation interactions: the cold continental interior paradox. In: Barrera, E., Johnson, C. (Eds.), The Evolution of Cretaceous Ocean/Climate Systems. Geol. Soc. Am., Boulder, CO, pp. 391–406.

DeConto, R.M., Thompson, S.L., Pollard, D., 1999b. Recent advances in paleoclimate modeling: toward better simulations of warm paleoclimates. In: Huber, B.T., Wing, S.L., MacLeod, K. (Eds.), Warm Climates in Earth History. Cambridge University Press, Cambridge, pp. 21–49.

Dickinson, R.E., Henderson-Sellers, A., Kennedy, P.J., Wilson, M.F., 1986. Biosphere–Atmosphere Transfer Scheme (BATS) for the NCAR community climate model. NCAR, Boulder, Colorado. 275+STR.

Dorman, J.L., Sellers, P.J., 1989. A global climatology of albedo, roughness length and stomatal resistance for atmospheric general circulation models as represented by the Simple Biosphere Model (SiB). J. Appl. Meteorol. 28, 833–855.

Dutton, J.F., Barron, E.J., 1996. GENESIS sensitivity to changes in past vegetation. Paleoclimates 1, 325–354.

Falcon-Lang, H.J., Cantrill, D.J., 2001. Gymnosperm woods from the Cretaceous (mid-Aptian) Cerro Negro Formation, Byers Peninsula, Livingston Island, Antarctica: the arborescent vegetation of a volcanic arc. Cretac. Res. 22 (3), 277–293.

Falcon-Lang, H.J., Cantrill, D.J., Nichols, G.J., 2001. Biodiversity and terrestrial ecology of a mid-Cretaceous, high-latitude floodplain, Alexander Island, Antarctica. J. Geol. Soc. Lond. 158, 709–724.

Flower, B.P., 1999. Cenozoic deep-sea temperatures and polar glaciation: the oxygen isotope record. In: Barrett, P., Orombelli, G. (Eds.), Geological Records of Global and Planetary Changes, Terra Antart. Rep., vol. 3. Terra Antart. Publication, Siena, pp. 27–42.

Foley, J.A., Prentice, I.C., Ramankutty, N., Levis, S., Pollard, D., Sitch, S., Haxeltine, A., 1996. An integrated biosphere model of land surface processes, terrestrial carbon balance, and vegetation dynamics. Glob. Biogeochem. 10, 603–628.

Foley, J.A., Levis, S., Costa, M.H., Cramer, W., Pollard, D., 2000. Incorporating dynamic vegetation cover within global climate models. Ecol. Appl. 10, 1620–1632.

Frakes, L.A., Francis, J.E., Syktus, J.I., 1992. Climate Modes of the Phanerozoic: The History of the Earth's Climate Over the Last 600 Million Years. Cambridge University Press, Cambridge.

Francis, J.E., 1991. Palaeoclimatic significance of Cretaceous–early Tertiary fossil forests of the Antarctic Peninsula. In: Thomson, M.R.A., Crame, A., Thomson, J.W. (Eds.), Geological Evolution of Antarctica. Cambridge University Press, New York, pp. 623–627.

Francis, J.E., 1999. Evidence from fossil plants for Antarctic Palaeoclimates over the past 100 million years. In: Barrett, P.J., Orombelli, G. (Eds.), Geological Records of Global and Planetary Changes, Terra Antart. Rep., vol. 3. Terra Antart. Publication, Siena, pp. 43–52.

Francis, J.E., 2000. Fossil wood from Eocene high latitude forests, McMurdo Sound, Antarctica. In: Stilwell, J.D., Feldmann, R.M. (Eds.), Paleobiology and Paleoenvironments of Eocene Rocks, McMurdo Sound, East Antarctica, Ant. Res. Ser., vol. 76. AGU, Washington, D.C., pp. 253–260.

Francis, J.E., Hill, R.S., 1996. Fossil plants from the Pliocene Sirius group, Transantarctic Mountains: evidence for climate from growth rings and fossil leaves. Palaios 11 (4), 389–396.

Francis, J.E., Poole, I., 2002. Cretaceous and early Tertiary climates of Antarctica: evidence from fossil wood. Palaeogeogr. Palaeoclimatol. Palaeoecol. 182 (1–2), 47–64.

Gallimore, R.G., Kutzbach, J.E., 1996. Role of orbitally induced changes in tundra area in the onset of glaciation. Nature 381, 503–505.

Hay, W.W., DeConto, R.M., Wold, C.N., Wilson, K.M., Voigt, S., Schulz, M., Wold-Rossby, A., Dullo, W.-C., Ronov, A.B., Balukhovsky, A.N., Söding, E., 1999. An alternative global Cretaceous paleogeography. In: Barrera, E., Johnson, C. (Eds.), The Evolution of the Cretaceous-Ocean Climate System. Geol. Soc. Am., Boulder, pp. 1–48.

Haywood, A.M., Valdes, P.J., 2004. Modelling Pliocene warmth: contribution of atmosphere, ocean and cryosphere. Earth Planet. Sci. Lett. 218, 363–377.

Henderson-Sellers, A., 1993. Continental vegetation as a dynamic component of a global climate model: a preliminary assessment. Clim. Change 23, 337–378.

Hicock, S.R., Barrett, P.J., Holme, P., 2003. Fragment of an Antarctic outlet glacier system near the top of the Transantarctic Mountains. Science 31 (9), 821–824.

Hill, R.S., 1989. Palaeontology fossil leaf. In: Barrett, P.J. (Ed.), Antarctic Cenozoic history from the CIROS-1 drillhole, McMurdo Sound, DSIR Bulletin, vol. 245. Department of Scientific and Industrial Research Publishing, Wellington, pp. 143–144.

Hill, R.S., Scriven, L.J., 1995. The angiosperm-dominated woody vegetation of Antarctica: a review. Rev. Palaeobot. Palynol. 86 (3–4), 175–199.

Houghton, J.T., Jenkins, G.J., Ephraums, J.J. (Eds.), 1990. Climate Change: The IPCC Scientific Assessment. Cambridge University Press, Cambridge.

Huber, M., Sloan, L.C., 2001. Heat transport, deep waters, and thermal gradients: coupled simulation of an Eocene greenhouse climate. Geophys. Res. Lett. 28 (18), 3481–3484.

Huber, M., Brinkhuis, H., Stickley, C.E., Doos, K., Sluijs, A., Warnaar, J., Schellenberg, S.A., Williams, G.L., 2004. Eocene circulation of the southern Ocean: was Antarctica kept warm by subtropical waters? Paleoceanography 19, PA4026.

Jones, T.P., Rowe, N.P., 1999. Fossil Plants and Spores: Modern Techniques. Geol. Soc. Lond., London.

Kennett, J.P., 1977. Cenozoic evolution of Antarctic glaciation, the circum-Antarctic oceans and their impact on global paleoceanography. J. Geophys. Res. 82, 3843–3859.

Laskar, J., Robutel, P., Joutel, F., Gastineau, M., Correia, A.C.M., Levrard, B., 2004. A long term numerical solution for the insolation quantities of the Earth. Astron. Astrophys. 428 (1), 261–285.

McGuffie, K., Henderson-Sellers, A., 1997. A Climate Modelling Primer. Research and Developments in Climate and Climatology. John Wiley & Sons, Chichester, England.

Mildenhall, D.C., 1989. Palaeontology — terrestrial palynology. In: Barrett, P.J. (Ed.), Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, DSIR Bulletin, vol. 245. Department of Scientific and Industrial Research Publishing, Wellington, pp. 119–127.

Miller, M.F., Mabin, M.C.G., 1998. Antarctic neogene landscapes — in the refrigerator or in the deep freeze? GSA Today 8 (4), 1–3.

Miller, K.G., Wright, J.D., Fairbanks, R.G., 1991. Unlocking the ice house: oligocene to miocene oxygen isotopes, eustasy, and margin erosion. J. Geophys. Res. 96, 6829–6848.

Naish, T.R., Woolfe, K.J., Barrett, P.J., Wilson, G.S., Atkins, C., Bohaty, S.M., Bucker, C.J., Claps, M., Davey, F.J., Dunbar, G.B., Dunn, A.G., Fielding, C.R., Florindo, F., Hannah, M.J., Harwood, D.M., Henrys, S.A., Krissek, L.A., Lavelle, M., van der Meer, J., McIntosh, W.C., Niessen, F., Passchier, S., Powell, R.D., Roberts, A.P., Sagnotti, L., Scherer, R.P., Strong, C.P., Talarico, F., Verosub, K.L., Villa, G., Watkins, D.K., Webb, P.N., Wonik, T., 2001. Orbitally induced oscillations in the east Antarctic ice sheet at the Oligocene/Miocene boundary. Nature 413 (6857), 719–723.

Otto-Bliesner, B.L., Upchurch, G.R.J., 1997. Vegetation-induced warming of high-latitude regions during the Late Cretaceous period. Nature 385, 804–807.

Otto-Bliesner, B.L., Brady, E.C., Shields, C., 2002. Late Cretaceous ocean: coupled simulations with the National Center for Atmospheric Research climate system model. J. Geophys. Res.-Atmos. 107 (D1–D2) (14 pp.)

Philip, G. (Ed.), 1998. Philip's Atlas of the World. George Philip Limited, London.

Pole, M., Hill, R.S., Harwood, D.M., 2000. Eocene plant macrofossils from erratics, McMurdo Sound, Antarctica. In: Stilwell, J.D., Feldmann, R.M. (Eds.), Paleobiology and Paleoenvironments of Eocene rocks, McMurdo Sound, East Antarctica, Ant. Res. Ser., vol. 76. AGU, Washington, D.C., pp. 243–251.

Pollard, D., DeConto, R.M., 2003. Antarctic ice and sediment flux in the Oligocene simulated by a climate-ice sheet-sediment model. Palaeogeogr. Palaeoclimatol. Palaeoecol. 198, 53–67.

Pollard, D., DeConto, R.M., 2005. Hysteresis in Cenozoic Antarctic ice-sheet variations. Glob. Planet. Change 45, 9–21.

Pollard, D., Thompson, S.L., 1995. Use of a Land-Surface-Transfer scheme (LSX) in a global climate model: the response to doubled stomatal resistance. Glob. Planet. Change 10, 129–161.

Pollard, D., Thompson, S.L., 1997. Driving a high-resolution dynamic ice-sheet model with GCM climate: ice sheet initiation at 116,000 BP. Ann. Glaciol. 25, 296–304.

Poole, I., Cantrill, D.J., 2001. Fossil woods from Williams points beds, Livingston island, Antarctica: a late Cretaceous southern high latitude flora. Palaeontology 44 (6), 1081–1112.

Prebble, J.G., Raine, J.I., Barrett, P., Hannah, M.J., this volume. Vegetation and climate from two Oligocene glacioeustatic sedimentary cycles (31 and 24 Ma) cored by the Cape Roberts Drilling Project, Victoria Land Basin, Antarctica. Palaeogeogr. Palaeoclimatol. Palaeoecol.

Prentice, I.C., Cramer, W., Harrison, S.P., Leemans, R., Monserud, R.A., Solomon, A.M., 1992. A global biome model based on plant physiology and dominance, soil properties and climate. J. Biogeogr. 19, 117–134.

Raine, J.I., 1998. Terrestrial palynomorphs from Cape Roberts Project drillhole CRP-1, Ross Sea, Antarctica. Terra Antartica 5 (3), 539–548.

Raine, J.I., Askin, R.A., 2001. Terrestrial palynology of Cape Roberts Project drillhole CRP-3, Victoria Land Basin, Antarctica. Terra Antartica 8 (4), 389–400.

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

Robinson, D.A., Kukla, G., 1985. Maximum surface albedo of seasonally snow-covered lands in the Northern Hemisphere. J. Clim. Appl. Meteorol. 24, 402–411.

Ronov, A., Khain, V., Balukhovsky, A., 1989. Atlas of Lithological–Paleogeographical Maps of the World: Mesozoic and Cenozoic of Continents and Oceans. Ministry of Geology, USSR, Leningrad.

Sellers, P.J., Mintz, Y., Sud, Y.C., Dalcher, A., 1986. A simple biosphere model (SiB) for use within general circulation models. J. Atmos. Sci. 43, 505–531.

Siegert, M.J., Hindmarsh, R.C.A., Hamilton, G.S., 2003. Evidence for a large surface ablation zone in central east Antarctica during the last ice age. Quat. Res. 59 (1), 114–121.

Thompson, S.L., Pollard, D., 1997. Greenland and Antarctic mass balances for present and doubled atmospheric CO_2 from the GENESIS version-2 global climate model. J. Climate 10, 871–900.

Trenberth, K.E., 1992. Climate System Modelling. Cambridge University Press, Cambridge.

Truswell, E.M., 1990. Antarctica: a history of terrestrial vegetation. In: Tingey, R. (Ed.), The Geology of Antarctica, Oxford Monographs on Geology and Geophysics. Clarendon Press, Oxford, pp. 499–537.

Veblen, T.T., Hill, R.S., Read, J., 1996. The Ecology and Biogeography of *Nothofagus* Forests. Yale University Press, New Haven.

Zachos, J.C., Quinn, T.M., Salamy, K.A., 1996. High-resolution (10^4 years) deep-sea foraminiferal stable isotope records of the Eocene–Oligocene climate transition. Paleoceanography 11, 251–266.

Zachos, J.C., Flower, B.P., Paul, H., 1997. Orbitally paced climate oscillations across the Oligocene/Miocene boundary. Nature 388 (6642), 567–570.

Zachos, J., Billups, K., Pagani, H., Sloan, L., Thomas, E., 2001. Trends, rhythms, and aberrations in global climate 65 Ma to present. Science 292 (5517), 686–693.