Environmental and Climatological Evolution
of the early Paleogene Southern Ocean

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Milieu- en Klimaatsevolutie van de
Zuidelijke Oceaan tijdens het vroeg Paleogeen

(met een samenvatting in het Nederlands)

Proefschrift

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'It is only by combing the information furnished by all the Earth Sciences that we can hope to determine "truth" here'

- **Alfred Wegener (1880-1930)**

Contents

GENERAL INTRODUCTION AND SYNOPSIS

Dinopterygium cladoides Deflandre, 1935
Cyst is characterized by the possession of a narrow equatorial cingulum, with two very large membranes, and by the development of a system of fine wing-like, crested membranes, which delineate the tabulation. It has a goniodomacean tabulation and a disintegration archeopyle.

Since the beginning of the industrial revolution about 150 years ago, the burning of fossil hydrocarbons is increasing the partial pressure of atmospheric carbon dioxide (pCO$_2$) at an unprecedented rate. If CO$_2$ emissions continue unabated, pCO$_2$ levels will approach 1,000 parts per million by volume (ppmv) at the end of the 21st century, much higher compared to ~390 ppmv at present and ~280 ppmv prior to massive industrialization (IPCC 2007). CO$_2$, as a greenhouse gas, absorbs outgoing long-wave radiation (heat) and emits it to all directions, also to the Earth's surface. Higher CO$_2$ levels are therefore expected to result in a warming of the lower atmosphere. Moreover, the residence time of CO$_2$ is long enough for the Earth to experience the effects: the emitted CO$_2$ will reside in the atmosphere-ocean system for 100,000 years (Walker and Kasting 1992; Dickens 2003; Archer 2005). The climatic response to pCO$_2$ increase is already detectable in field observations (IPCC 2007). Computer model experiments are conducted to predict the climatic and environmental effects of the predicted greenhouse gas increase. A fundamental assumption in these computer models is the principle of actualism: the forces and feedbacks at hand in present-day climates also apply in the future. The question is whether these feedbacks are similar in different climatic conditions. One way to test the validity of computer model output is to look in the Earth's geologic past: investigating what the Earth looked like in the past with conditions analogous to our future, but also how sensitive it responded to perturbations of the climate system, may provide insights in future climate change.

To study a world with pCO$_2$ approaching that predicted for the future, paleoclimatologists are inspired to go far back in time. The available evidence suggests that pCO$_2$ did not exceed 1,000 ppmv during the last 35 million years of Earth history. Most of that time, pCO$_2$ was below 500 ppmv (Beerling and Royer 2011). The time period from beyond 35 million years ago (Ma) is characterized by pCO$_2$ values that may have been well above 1,000 ppmv (Pearson and Palmer 2000; Pagani et al., 2005; Zachos et al., 2008). Evidence for much warmer climates in that time period is particularly convincing in polar regions. Vegetation reconstructions from northern high latitudes indicate that during the Eocene Epoch, 56 to 34 Ma, the continents fringing the Arctic Ocean were forested with (amongst others) *Metasequoia*, now flourishing in temperate rainforests (Francis 1988). Crocodile skulls were found in Alaska (Markwick 1998), and footprints reveal that ancestors of modern-day hippos roamed marshes on what now is Svalbard. In fact, some mammals were year-round residents of the high Canadian Arctic (e.g., Eberle et al., 2009). During the zenith of Eocene warmth, at ~54 Ma, palms flourished on the landmasses fringing the Arctic Ocean (Sluijs et al., 2009a), and at ~50 Ma, the Arctic Ocean itself was blanketed by thick mats of the free-floating fresh water fern *Azolla* (Brinkhuis et al., 2006; Collinson et al., 2010). The latter phenomenon and the reconstructed vegetation on land evidences that, at least seasonally, profound wet precipitation regimes existed in the northern high latitudes (Pagani et al., 2006; Speelman et al., 2010; Barke et al., 2011).

At southern high latitudes, Paleogene macrofossil remains have been found on outcrops on the Antarctic continent, evidencing angiosperm forests (Francis and Poole 2002). New Zealand (50° South paleolatitude, Slat. (Crouch 2001)) biomes contained a diversity of palms in the late Eocene (Hartwich et al., 2010). Clay

13

mineralogical studies from Eocene Antarctic coastal sediments suggest wet weathering regimes in the Antarctic hinterland (Ehrmann 1998). The discovery of warm and wet conditions is enigmatic because polar environments in the Paleogene must still have been limited in daylight for half of the year. In order to explain the warmth on Eocene Antarctica, a hypothesis was proposed involving the effect of Southern Hemisphere continental configuration changes. Because Australia was still more or less attached to Antarctica during most of the Eocene (Lawver and Gahagan 2003) and also the conduit between South America and Antarctica was only shallowly open (Lagabrielle et al., 2009), the modern-day isolating Antarctic Circumpolar Current (ACC) could not circulate the Antarctic continent. Instead, low latitude, warm surface currents were thought to have reached and warmed Antarctica (Kennett 1977; Fig. 1). However, this hypothesis was later disputed by the results of general circulation model experiments (Sloan and Rea 1995). Applying these models to the Paleogene revealed that even in absence of an ACC, low latitude, warm currents could not reach the Antarctic coastline. Instead, large clockwise gyres existed in the Southern Ocean, which effectively prevented low latitude currents to reach Antarctica (Sloan and Rea 1995; Huber et al., 2004; Fig. 1). The currently leading hypothesis explaining the warmth of the early Paleogene now is related to the high concentrations of greenhouse gases, and associated feedbacks (Zachos et al., 2008), although sufficient direct evidence to support this statement is lacking.

In stark contrast to the overwhelming evidence for general warmth are the enigmatic findings of ice rafted debris, sea-ice diatoms and evidence for frost in the Arctic Ocean from 46 million years onwards (Eldrett et al., 2007; St. John 2008; Stickley et al., 2009). Although Paleogene climates seemed to have passed their warmest by then (Zachos et al., 2008), still both temperatures and pCO$_2$ concentrations are considered too high to allow for the presence of large ice caps on the poles (DeConto et al., 2008). On the Southern Hemisphere, remains of surface-dwelling dinoflagellates show a dramatic shift towards dominance of forms that are endemic to the Southern Ocean (e.g., Wrenn and Beckman, 1982; Wrenn and Hart, 1988; Brinkhuis et al., 2003a; Warnaar 2006). Intuitively, such an endemic dinocyst community suggests cooling of Southern Ocean surface waters. This is in contradiction with the reduced meridional temperature gradients as inferred from fossil leaf analyses (Greenwood and Wing 1995), which would promote poleward migration of biota rather than polar endemism. Moreover, the absence of geographic barriers separating the Southern Ocean from adjacent, warmer basins would further promote proliferation of low-latitude dinoflagellates into the Southern Ocean.

In general, while a wealth of evidence suggests that the early Eocene was warm, and the remainder of the Eocene had experienced cooling towards the icehouse world of the Oligocene and Neogene, still many questions remain on timing, nature and causes of the greenhouse-icehouse transition. Improved understanding of high-latitude climate evolution of the Southern Ocean starts with continuous, chronostratigraphically well-constrained time series of early Paleogene high-latitude temperature, climatic and biotic changes. The record from the Arctic Coring Expedition (Moran et al., 2006) provides a unique first insight in Northern high-latitude paleo-environments in the Paleogene greenhouse, but the record is patchy, of

low resolution and with large hiatuses (Sluijs et al., 2008a). The Southern Ocean, however, has much better coverage of sedimentary archives provided by the Deep Sea Drilling Project (DSDP), and subsequently the Ocean Drilling Program (ODP) and the on going Integrated Ocean Drilling Program (IODP). One issue that hampers investigations of these cores has been the lack of biostratigraphic tools to date these sediments: conventional calcite-based microfossils are usually absent in high-latitude sediments. In part because of this, still relatively little is known about Southern high-latitude climate evolution in the early Paleogene.
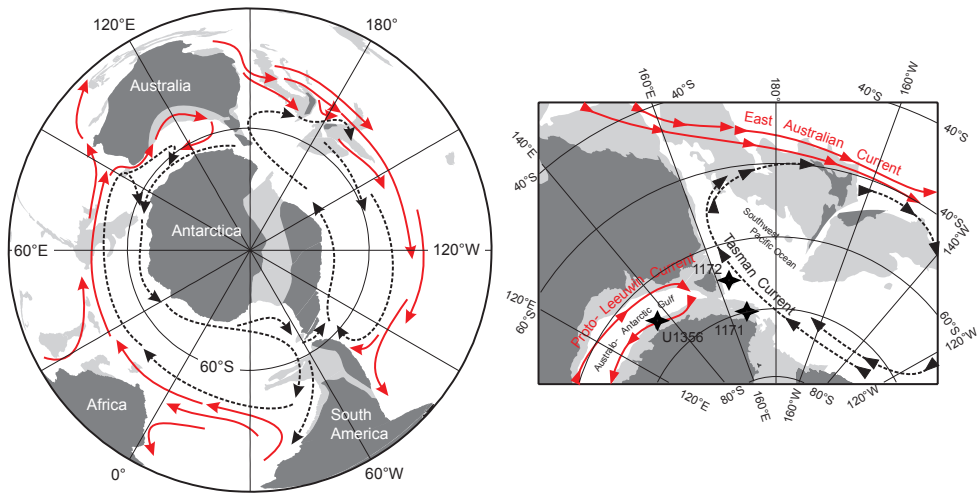
Figure 1. Left: Paleogeographic reconstruction of the Southern Ocean. Tectonic configurations were obtained from de odsn website: www.odsn.de. Note that Australia was much closer to Antarctica than it is today. The light grey areas indicate continental blocks that were submerged in the Eocene. Also plotted is the generalized Eocene surface current configuration as interpreted from computer model studies applying the Eocene continental configuration (Huber et al., 2004). Right: detailed paleogeographic reconstruction of the southwest Pacific Ocean. Black asterisks indicates the main drill sites from which sediments were used in this thesis: Ocean Drilling Program (ODP) Site 1172 on the East Tasman Plateau (ETP); ODP Site 1171 on the South Tasman Rise (STR) and Integrated Ocean Drilling Program (IODP) Site U1356 on the Wilkes Land Margin.

An indirect approach to evaluate Paleogene Southern Ocean climate evolution may be captured in the compilation of benthic foraminiferal oxygen isotopes ($\delta^{18}O$; Zachos et al., 2008). The $\delta^{18}O$ value of benthic foraminifera is mainly influenced by water temperature and changes in $\delta^{18}O$ of the seawater. The latter is, in turn, influenced by ice volume and salinity. In the absence of major continental ice, salinity changes mainly influence the seawater $\delta^{18}O$ in these times. The salinity and temperature of intermediate and deep water depend on the source and mechanisms of deep and intermediate water formation. The Southern Ocean is thought to have been the main source for deep-water formation in the Eocene (Thomas et al., 2003). Hence, the signal obtained from benthic foraminiferal calcite produced within deep and intermediate waters may in fact bear the signatures of climatic evolution of Southern Ocean surface waters.

15

The benthic $\delta^{18}O$ stack (Zachos et al., 2008) shows a general cooling from the Cretaceous-Paleogene boundary, towards a minimum in the mid-Paleocene (~58 Ma; Zachos et al., 2008; Fig. 2). Subsequently, temperatures rise gradually until ~52 Ma, where long-term Cenozoic peak temperatures are reached until ~50 Ma: the Early Eocene Climatic Optimum (EECO). Here, intermediate waters reached temperatures 13°C warmer than today, assuming ice-free conditions and normal salinity. The subsequent 15 million years, from 49 to 34 Ma, are characterized by gradual cooling of intermediate waters (Zachos et al., 2008; Fig. 2), eventually culminating in the onset of major glaciation of Antarctica (Zachos et al., 1992), at around 34 million years ago (Zachos et al., 2008). Because the presence of ice prior to the early Oligocene isotope step is still under debate (Eldrett et al., 2007; St. John 2008), it is as yet unknown how much of the middle and late Eocene $\delta^{18}O$ increase is related to cooling, how much to ice volume and how much to a change in sources of deep water formation.

Superimposed on these trends, several intervals of more rapid climate change in a transient fashion have been recognized. The Paleocene-Eocene Thermal Maximum (PETM) is a brief (~150 kyrs) interval of 5-8°C warming documented in the tropics and in mid- and high-latitude sites on both hemispheres (Kennett and Stott 1991; Sluijs 2006; see Sluijs et al., 2007a for a review). Transient warming at this event is accompanied by a prominent negative excursion in the stable carbon isotopic composition of sedimentary components (Kennett and Stott 1991; Sluijs et al., 2007a). This evidences an input of ^{13}C-depleted carbon into the global exogenic carbon pool. Because ample evidence shows that the amount of carbon being added to the exogenic carbon pool at the PETM is in the same order as is expected for the future after all fossil fuels have been burned, this climate event is seen as best analogue to future climate change. Younger, less extreme events with many similarities to the PETM are recognized in the interval from the PETM (~56 Ma) up to 51 Ma (e.g., Cramer et al., 2003; Nicolo et al., 2007; Stap et al., 2010; Zachos et al., 2010). These so-called 'hyperthermals' come with various acronyms, and are, like the PETM, characterized by a shift towards lower $\delta^{13}C$ values of the exogenic carbon pool.

A younger prominent feature of relatively short-term climate warming is the 'Middle Eocene Climate Optimum' (MECO; Bohaty and Zachos 2003), occurring at ~39 Ma (Bohaty et al., 2009). Similar to the PETM and other hyperthermal phases, the MECO stands out in deep-sea records as a phase of carbonate dissolution (Bohaty et al., 2009), suggestive of involvement of changes in the carbon cycle. In contrast to the early Eocene hyperthermal phases, the MECO is not accompanied by a shift towards lighter carbon isotopes (Bohaty et al., 2009). This suggests that, if CO_2 was involved in climate change, the source of carbon injected into the atmosphere is different compared to the early Eocene hyperthermal phases.

In this thesis, some of the outstanding questions will be addressed in order to better understand the climatic evolution of the Southern Ocean, such as: how did southern high-latitude surface waters evolve in the early Paleogene, particularly in comparison to that reconstructed for deeper waters? How sensitive were Southern Ocean surface waters to long- and short-term climatic oscillations in the early

Paleogene? What was the biotic response to the temperature and circulation changes? What was the role of atmospheric CO_2, and what of the opening and subsequent deepening of the Tasmanian Gateway? What was the climate sensitivity of the Southern Ocean in the Paleogene greenhouse?

To answer these questions, a combination of techniques is applied that have proven particularly useful to reconstruct high-latitude climate change. In contrast to calcite-based microfossil groups, the fossil remains of surface-dwelling dinoflagellates, i.e., their organic-walled cysts (dinocysts) have been found abundantly in high-latitude settings. In most high-latitude settings, particularly further back in the geologic record, dinocysts are the only biostratigraphic tool to tie high-latitude sections to international time scales (Brinkhuis et al., 2003a; Eldrett et al., 2004; Williams et al., 2004). On top of this, dinocyst assemblages provide valuable qualitative information on a suite of physical properties of the surface waters, such as nutrients, temperature, salinity and coastal proximity (Dale 1996; Pross and Brinkhuis 2005). Specifically for southern high latitudes, dinocyst assemblages consist of species that are endemic to the Southern Ocean (Wrenn and Beckman 1982; Wrenn and Hart 1988). These species were referred to as transantarctic flora, because it was suggested that they evidence the existence of a transantarctic seaway (Wrenn and Beckman 1982), and because at that time, dinoflagellates were apparently considered flora. Although dinoflagellates constitute of both heterotrophic and autotrophic forms, and it can be debated whether the data from Wrenn and Beckman (1982) truly supports the presence of a transantarctic seaway, the name of this dinocyst community is maintained in this thesis. It was shown that transantarctic floral species dominate in regions that are under influence of Antarctic-derived surface currents (Huber et al., 2004). This observation intuitively argues for temperature as the controlling environmental factor on the biogeographic distribution of endemic dinocysts, although this has not yet been tested adequately.

In addition to the dinocyst assemblages, organic geochemical paleo-environmental tools can often be applied on dinocyst-bearing sediments, owing to the similarity in preservation requirements for both the organic compounds and dinocysts. The organic geochemical toolbox to reconstruct climate change has been widened in the recent years, with the invention of new tools, such as the TEX_{86} proxy for sea surface temperature (SST; Schouten et al., 2002; Kim et al., 2010) and pCO$_2$ based on ^{13}C of long chain alkenones (Pagani 2002). Combining organic geochemical and palynological tools on the wealth of sedimentary records from the Southern Ocean may help to reconstruct the dynamics and high-resolution climatic evolution of the early Paleogene high southern latitudes.

This thesis provides new insights into the paleoenvironmental and paleoclimatological evolution of the early Paleogene Southern Ocean. In **Chapters 1, 2,** and **3**, a framework for portraying the early Paleogene paleoclimatological and paleoenvironmental evolution in the Southern Ocean is provided, in terms of dinocyst bio-magnetostratigraphy, organic biomarker paleothermometry and dinocyst biogeographic evolution of the Southern Ocean. **Chapters 4, 5,** and **6** zoom in on critical intervals of more rapid climate change, in chronological order (Fig. 2).
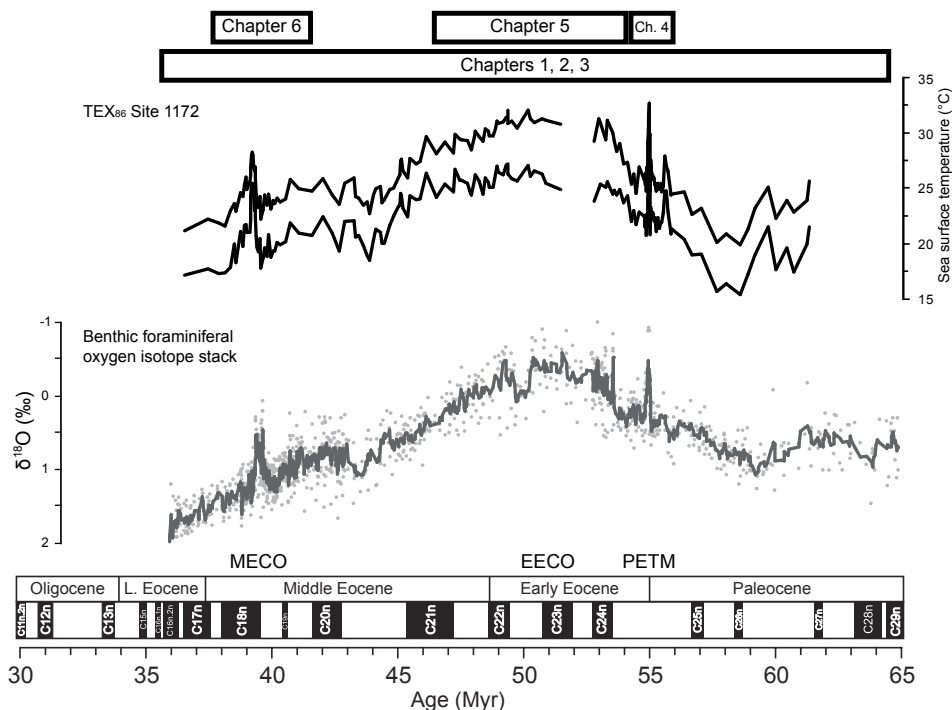
Figure 2. Early Paleogene Southwest Pacific Ocean TEX$_{86}$ sea surface temperature reconstructions, and the stack of benthic foraminiferal oxygen isotopes ($\delta^{18}O$; as modified from Zachos et al. (2008); see Chapter 2). Two calibrations for TEX$_{86}$ are used (Kim et al., 2010). Also indicated is the structure of the thesis per chapter. Chapter 1, 2 and 3 cover the long-term Paleocene and Eocene climate development, Chapter 4 focuses on the Paleocene-Eocene Thermal Maximum, Chapter 5 focuses on the early-middle Eocene boundary and Chapter 6 concerns the Middle Eocene Climatic Optimum.

In **Chapter 1** a calibrated organic-walled dinoflagellate cyst biostratigraphic zonation scheme is proposed for upper Paleocene - upper Eocene sediments from the South Pacific Ocean. We compare chronologies from Ocean Drilling Program (ODP) Holes 1171D and 1172A/D to that of Integrated Ocean Drilling Program (IODP) Hole U1356A, and subsequently to dinocyst zonations from New Zealand. Based on these correlations, we present a revised magnetostratigraphic age model for ODP Holes 1171D and 1172A/D and propose a South Pacific dinocyst zonation scheme consisting of thirteen dinocyst zones. The South Pacific Dinocyst Zonation is calibrated to the International Time Scale of Gradstein et al. (2004) with additional independent age control from stable carbon isotope profiles, i.e., those correlated to the Paleocene-Eocene Thermal Maximum and the Middle Eocene Climatic Optimum, and magnetostratigraphy.

In **Chapter 2** a uniquely continuous and chronostratigraphically well-calibrated TEX$_{86}$ record of SST is provided from Ocean Drilling Program Leg 189 Site 1172 at the East Tasman Plateau (~65° Slat.; Fig. 1). We show that southwest Pacific

SSTs rose above present-day tropical values (to 34°C) during the early Eocene (53 Ma) and had gradually decreased to about 21°C by the early late Eocene (36 Ma). Our results suggest that there was almost no latitudinal SST gradient between sub-equatorial and sub-polar regions during the early Eocene (~55–50 Ma). The latitudinal gradient markedly increased in the middle Eocene. In theory, if Eocene cooling was largely driven by a decrease in atmospheric greenhouse gas concentration (Zachos et al., 2008, and references cited therein), additional processes are required to explain the relative stability of tropical SSTs (Pearson et al., 2007) given that there was significant cooling at higher latitudes.

Despite warm polar climates and low meridional temperature gradients, a number of different high-latitude plankton assemblages were, to varying extents, dominated by endemic species during most of the Paleogene. To document the evolution of Paleogene plankton endemism in the high southern latitudes, in **Chapter 3** the spatio-temporal distribution of dinocysts is investigated. Moreover, we assess their response to changes in regional SST. Paleocene and early Eocene (65–50 Ma) Southern Ocean dinocyst assemblages were largely cosmopolitan in nature, but a distinct switch from cosmopolitan-dominated to endemic-dominated assemblages (the so-called "transantarctic flora") occurred around the early-middle Eocene boundary (50 Ma). The spatial distribution and relative abundance patterns of this transantarctic flora correspond well with surface water circulation patterns, as reconstructed through general circulation model experiments, throughout the Eocene. Dinocyst assemblages are quantitatively compared with the TEX_{86}– based SST reconstructions of Chapter 2 through the lower and middle Eocene from ODP Site 1172. One conclusion is that the middle Eocene onset of the proliferation of the transantarctic flora is not linearly correlated with regional SST records. Only after the transantarctic flora became fully established later in the middle Eocene, possibly triggered by large-scale changes in surface-ocean nutrient availability, abundances of endemic dinocysts show a high degree of correlation with regional SST variations.

In **Chapter 4** a record of the PETM is presented from ODP Site 1172. The organic paleothermometer TEX_{86} suggests that southwest Pacific sea surface temperatures increased from ~26°C to ~33°C during the PETM. Such temperatures before, during and after the PETM are >10°C warmer than predicted by paleoclimate model simulations for this latitude. Hence, not only Arctic, but also Antarctic temperatures may be underestimated in simulations of ancient greenhouse climates by the current generation of fully coupled climate models. An early influx of abundant *Apectodinium* is recognized at Site 1172, in accord with that in expanded PETM sections, such as those from New Jersey. This suggests once more that anomalous environmental change preceded the CIE on a global scale. Organic dinoflagellate cyst assemblages suggest a local decrease in the amount of river run-off reaching the core site during the PETM, possibly in concert with eustatic rise. Moreover, the assemblages suggest changes in seasonality of the regional hydrological system and perhaps storm activity. Finally, significant variation in dinoflagellate cyst assemblages during the PETM indicates that southwest Pacific climates varied significantly over time scales of $10^3 – 10^4$ years during this event, a finding comparable to similar studies of PETM successions from the New Jersey Shelf.

In **Chapter 5** the role of the opening of a critical Southern Ocean gateway, the Tasmanian Gateway in Antarctic climate change is investigated. Proxy-records are presented from recently recovered sediments from the Wilkes Land Margin, Antarctica (Escutia et al., 2011), deposited west of the Tasmanian Gateway in the Australo-Antarctic Gulf. Integrated marine and terrestrial microfossil and organic geochemical analyses show a strong cooling of east Antarctic surface waters (3-4°C) and adjacent landmasses (~6°C), at ~50 Ma. Critically, this cooling coincides with the first incursion of typical southwest Pacific phytoplankton that migrated across an apparently shallow southern opening of the Tasmanian Gateway at that time. Collectively, this implies that the onset of the relatively cool westward Antarctic Counter Current through the Tasmanian Gateway diminished the influence the warmer Australo-Antarctic waters. This suggests a significant role of the shallow opening of the Tasmanian Gateway in East Antarctic cooling, presetting this region for massive cryosphere development at the Eocene-Oligocene boundary at ~34 Ma.

The long-term warmth of the Eocene (~56-34 Ma) is commonly associated with elevated partial pressure of atmospheric carbon dioxide (pCO_2). However, a relationship between the two has not been established for short-term climate perturbations. In **Chapter 6** both pCO_2 and SST are presented over an episode of transient global warming called the Middle Eocene Climatic Optimum (MECO; ~39 Ma). Organic molecular paleothermometry indicates a 3-6°C warming of southwest Pacific SSTs whilst reconstructions of pCO_2 indicate a concomitant two- to three-fold increase. The marked consistency between SST and pCO_2 trends during the MECO suggests that elevated pCO_2 played a major role in global warming during the MECO.

In summary, this thesis provides a detailed image of the early Paleogene greenhouse-icehouse transition of the Southern Ocean. The dinocyst zonation for a large part of the early Paleogene may serve as guide for future biostratigraphy and age-assessment, which is crucial because few biostratigraphic tools are available in Eocene Southern high-latitude sediments. The SST evolution of the South Pacific Ocean mimics that of benthic foraminiferal oxygen isotopes, suggesting that the South Pacific Ocean was a region of deep-water formation. However, absolute temperatures seem much higher than those inferred from the benthic foraminifera. Part of this may be due to the effect of seasonal biases, i.e., that the deep water was formed mainly in winter and that the reconstructed Southern Ocean surface temperatures were skewed towards summer. The biotic response to the SST evolution in the Southern Ocean is controversial: dominance of endemic dinocyst assemblages seems unrelated to SST. The working hypothesis intuitively was that the proliferation of an endemic phytoplankton community at high latitudes was the result of outcompeting warm-water taxa when temperatures cooled. Because SST seems unrelated to the onset of proliferation of endemic dinocysts, an alternative suggestion is proposed relating the shift to an increase of surface-water nutrient levels in the middle Eocene. Indeed, several independent other proxies conspire to suggest a switch in nutrient input from the late early Eocene (~50 Ma) onwards, from basin-scale Pacific Ocean export productivity to inter-basin increase in benthic foraminiferal carbon isotopic gradients suggesting more effective aging of intermediate waters. The PETM interval in the South Pacific Ocean also stands out with a rapid sea surface temperature increase: 5-

8°C. Also an acme of the tropical dinocyst genus *Apectodinium* is observed. As for some other PETM sections, the PETM record from the South Pacific Ocean reveals major environmental change preceding the negative carbon isotope excursion (CIE). This is in concert with suggestions that initial climate changes preceded the release of carbon in the exogenic carbon pool that resulted in the CIE. In turn, this suggests that the carbon input in the exogenic carbon pool is a response rather than a forcing factor. Another ongoing debate that was touched upon in this thesis is the role of oceanic gateway openings and its role in climate change. Newly drilled (2010) sediment archives from the Wilkes Land Margin, Antarctica, showed that through flow of the Antarctic Counter Current was initiated around the early-middle Eocene boundary (~52-50 Ma), concomitant to amplified cooling of Antarctic continental and coastal temperatures. The opening of the Tasmanian Gateway occurred at the time when the Earth started cooling gradually towards the icehouse state of the past 35 million years. Hence, Tasmanian Gateway opening may have set the stage for climate deterioration, although atmospheric CO_2 was likely still the major forcing factor of early Paleogene climate evolution. Indeed, we still see that transient climate warming at the MECO coincides with increased concentrations of atmospheric CO_2, suggesting an important role for atmospheric CO_2 as driver of climate.

CHAPTER 1

A magnetostratigraphically calibrated dinocyst zonation for the early Paleogene South Pacific Ocean

Spinidinium macmurdoense (Wilson, 1967a) Lentin and Williams, 1976
Asymmetric pericyst with a well developed left antapical horn and a pentagonal endocyst. Pericyst tabulate with small penitabular spines (length 5-9 μm). Periarcheopyle is isodeltaform with operculum being posteriorly attached.

We propose an organic-walled dinoflagellate cyst (dinocyst) zonation for the late Paleocene to late Eocene (58 – 36 Million years ago; Ma) of the South Pacific Ocean, calibrated using magnetostratigraphy and stable isotope stratigraphy. Dinocyst chronologies were generated from Ocean Drilling Program (ODP) Holes 1171D on the South Tasman Rise, 1172A/D on the East Tasman Plateau and Integrated Ocean Drilling Program (IODP) Hole U1356A on the Wilkes Land margin. Correlation to dinocyst zonations from New Zealand lead to revisions of the magnetostratigraphic age model at Holes 1171D and 1172A/D. Stable carbon isotope profiles provide correlations to the Paleocene-Eocene Thermal Maximum (~56 Ma) and the Middle Eocene Climatic Optimum (~40 Ma). The resulting zonation consists of thirteen dinocyst zones and is calibrated to the International Time Scale of Gradstein et al. (2004).

1. Introduction

Throughout the past decades, interest has risen for southern high-latitude sediments from the early Paleogene Greenhouse period (~65 to 35 Ma), particularly for paleoclimate reconstructions (Wrenn and Beckman 1982; Wrenn and Hart 1988; Hannah 1997; O'Brien et al., 1998, 2004). Precise dating of the sediments has proven to comprise the major obstacle, because the well-calibrated calcite fossil groups, such as foraminifera and calcareous nannoplankton, are typically poorly or not preserved in high-latitude settings (Wilson et al., 1998; Florindo et al., 2003; Barker et al., 2007). In addition, siliceous microfossils such as diatoms and radiolaria dissolve and are typically not preserved in lower Paleogene sediments in the southern ocean. Organic-walled dinoflagellate cysts (dinocysts) are, however, often abundant and well-preserved in southern high-latitude marginal marine sediments (Wrenn and Hart 1988; Brinkhuis et al., 2003a, b), but this group still lacks a well-calibrated biostratigraphic framework in the Southern Ocean.
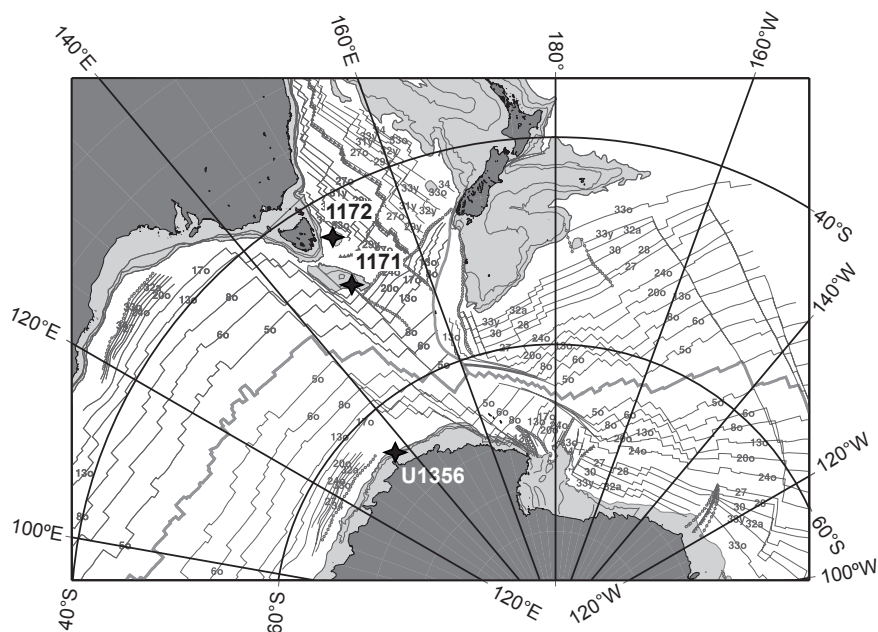
Figure 1. Present-day site locations of Ocean Drilling Program Sites 1172 and 1171 on the East Tasman Plateau (ETP) and South Tasman Rise (STR) in the Southwest Pacific, and Integrated Ocean Drilling Program Site U1356 on the Wilkes Land Margin, Antarctica (map modified from Cande and Stock (2004)). Dark grey areas indicate present-day shorelines, light-grey areas are submerged continental blocks above 3000 meters. White areas are below 3000 meters.

The Ocean Drilling Program (ODP) recovered early Paleogene sedimentary successions around Tasmania (Leg 189; at Sites 1171 and 1172; Fig. 1). An initial bio-magnetostratigraphic age model (including dinocysts) was published in the Scientific Results Volume (Stickley et al., 2004a). Extensive research at Site 1172 over the past

years has positively identified the Paleocene-Eocene Thermal Maximum (Chapter 4) and the Middle Eocene Climatic Optimum (Chapter 6), which represent chrono-stratigraphic tie points, and allows for a significant improvement of the age model. In addition, the recently recovered sedimentary succession at Site U1356 offshore the Wilkes Land Margin, Antarctica (Fig. 1) provides additional age constraints for early-middle Eocene dinocyst stratigraphy. We here evaluate the existing dinocyst chronology at Holes 1171D and 1172A/D and review the magnetostratigraphic calibrations made previously. Furthermore, we compare dinocyst chronologies at Holes 1171D and 1172A/D to that of Hole U1356A, and subsequently to dinocyst zonations from New Zealand. Based on these correlations, we present a revised bio-magnetostratigraphic age model for Holes 1171D and 1172A/D that allows for a South Pacific dinocyst zonation scheme.

2. Geological background and material

2.1 Tectonic evolution of the south Pacific and south Indian Ocean

The breakup of supercontinent Gondwana around Jurassic-Cretaceous boundary times (~145 Ma; Willcox and Stagg 1990) resulted in rifting between Australia and Antarctica. Continental crustal stretching and thinning occurred pulsewise, and formed the Australo-Antarctic Gulf. Continued rifting eventually resulted in the formation of oceanic crust in the Australo-Antarctic Gulf by the late Cretaceous (~83 Ma; Close et al., 2009). Sea floor spreading rates were slow initially (1.5-7.5 mm/yr; Close et al., 2009), and Antarctica and Australia remained attached at the Tasmanian side at least until the early Paleocene (~64 Ma; Woodburne and Case 1996). Before 64 Ma, Australian and South American terrestrial mammal assemblages exchanged, via a land connection over the Tasmanian land bridge (Woodburne and Case 1996). Diversifying mammal assemblages on Australia from 64 Ma onwards thus signifies the formation of a marine barrier, the Tasmanian Gateway. This Gateway must initially have been shallow, because, e.g., dinoflagellate cyst assemblages from either side of the Gateway are distinctly different until 50 Ma (Fig. 2a; Chapter 6). Crustal anomaly studies have shown a distinct increase in sea floor spreading rates in the Australo-Antarctic Gulf from 50 Ma onwards (Fig. 2b, c; Close et al., 2009). Continental blocks surrounding the Tasmanian Gateway slowly deepened, such as the East Tasman Plateau (Röhl et al., 2004a) and the South Tasman Rise (Cande and Stock 2004). Gradual drowning ultimately evolved in rapid deepening of continental blocks surrounding Tasmania starting at ~35.5 Ma (Stickley et al., 2004b), when the northward drift of the Australian continent commenced (Fig. 2d). This tectonic evolution dictated the surface oceanic circulation and thereby the distribution of dinocysts around the Tasmanian sector, with differentiation of dinocyst assemblages on either side of the gateway until 50 Ma (Chapter 3, 6), and major reorganizations of the dinocyst assemblages occurred when the Tasmanian Gateway deepened (Sluijs et al., 2003; Stickley et al., 2004b; Warnaar 2006). These dinocyst distributions are closely consistent with the output of ocean current configurations from computer model experiments (Sloan and Rea 1995; Huber et al., 2004).

2.2 Site 1172

2.2.1 Site and lithology description Holes 1172A/D

Site 1172 is located on the submerged flat western side of the East Tasman Plateau (ETP), at 45° Southern latitude (Slat.), at a water depth of 2620 meters (Fig. 1; Exon et al., 2004). Four holes were drilled in total, of which Hole 1172A extends to the mid-Lutetian (bottom of the hole at ~504 meters below sea floor (mbsf)). Hole 1172D contains sediments from mid-Lutetian to upper Maastrichtian, with good overlap between the two holes. Sediments used in this study are from ~358-630 mbsf. Paleocene - lower Eocene sediments are olive grey, with some variations to greyish brown and to greenish grey, with abundant bioturbation (Exon et al., 2001).
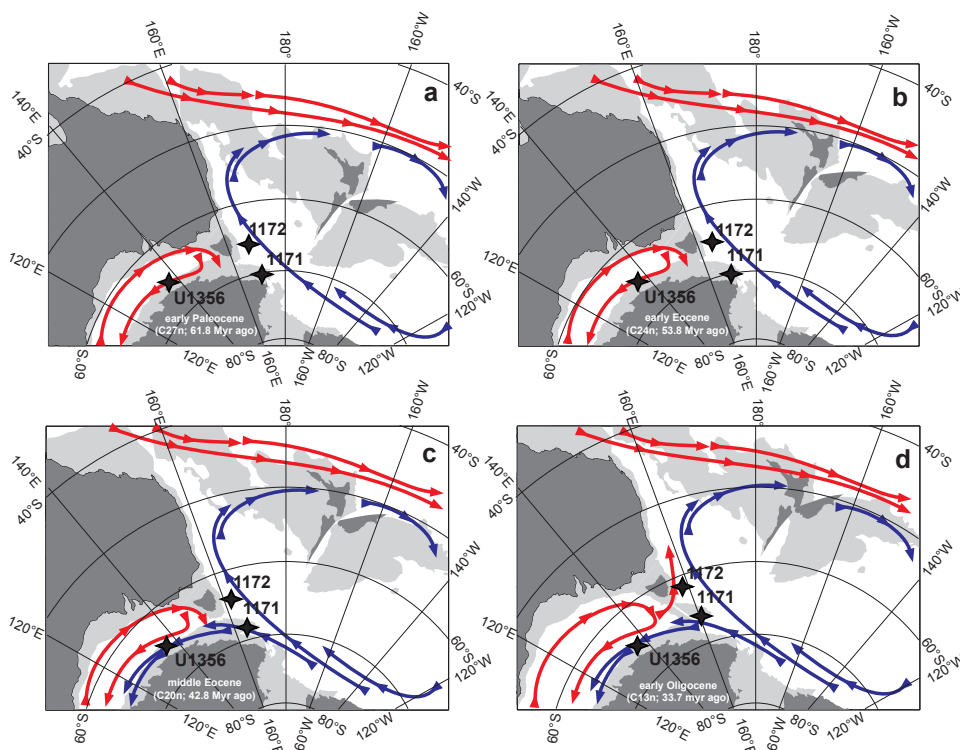
Figure 2. Paleogeographic reconstructions of the Australian Sector of the Southern Ocean, for (a) 61.1 Ma, (b) 53.3 Ma, (c) 43.8 Ma and (d) 33.5 Ma (maps modified from Cande and Stock (2004)). Also plotted are the reconstructed locations of the sites mentioned in this study, and the reconstructed surface current configurations. Shadings as in Fig. 1.

Sediments are virtually barren of any calcareous and siliceous microfossil groups, hampering conventional biostratigraphic dating (Stickley et al., 2004a). Middle - upper Eocene sediments contain increasing abundances of other microfossil groups, upsection, such as calcareous nannofossils, radiolarians and diatoms and eventually planktonic foraminifera (Stickley et al., 2004a). Prior to 35.5 Ma, lateral tectonic motion of the ETP was limited to slight eastward rift, away from Australia

(Fig. 2). Gradual subsidence of the East Tasman Plateau provided accommodation space for continuous shallow marine deposition. Middle Eocene strata are characterized by slightly deeper facies, increased carbonate content of the sediments and rapid shifts in dinocyst assemblages and elemental counts, all of which reflect third-order sea level cycles (Röhl et al., 2004a).

2.2.2 Age model Hole 1172A/D

Age control in the pre-Oligocene part of the succession relies on dinocysts, and occasionally on diatom and calcareous nannofossil stratigraphy, combined with magnetostratigraphy (Stickley et al., 2004a). Inclination data for Holes 1172A/D were published in the initial reports volume (Exon et al., 2001) and available on the ODP website (www-odp.tamu.edu), and magnetic reversals were identified from the inclination data as published by Fuller and Touchard (2004). Interpretation of the inclination data was complicated due to the strong normal overprint (Fuller and Touchard 2004). A first inventory of dinocyst species found during Leg 189 is published in the Scientific Results section of the Ocean Drilling Program (Brinkhuis et al., 2003a). The upper Paleocene and Eocene successions can be correlated to those in New Zealand, because many key dinocyst events were recognized at both locations (Brinkhuis et al., 2003a, b; Crouch and Brinkhuis 2005). Furthermore, high-resolution studies have added stratigraphic tie points and/or provided independent age constraints for Holes 1172A and D. For example, the identification of a ~3‰ negative excursion in the stable carbon isotopic composition ($\delta^{13}C$) of total organic carbon in Section 1172D-15R-4 reveals the position of the Paleocene-Eocene Thermal Maximum (PETM), along with dominance of dinocyst *Apectodinium* spp. and a transient 5-8°C sea surface temperature rise (Chapter 4). Also, bulk carbonate carbon isotope analyses revealed a profile within Core 1172A-46X to -42X, which allowed correlation to the Middle Eocene Climatic Optimum (MECO; Chapter 6). We incorporated these new insights into a revised magnetostratigraphic interpretation of ODP Site 1172 in the range chart of selected dinocyst species (Fig. 3). The magnetostratigraphic interpretation follows that in Chapter 6.

2.3 Site 1171
2.3.1 Site and lithology description Hole 1171D

Site 1171 is located in lower bathyal water depths of about 2150 meters on a gently south-westerly slope on the southern part of the South Tasman Rise (STR), at 48° Slat. (Fig. 1; Stickley et al., 2004a). The age model for this site indicates the presence of strata from early Eocene (mid-Subchron C24n; Röhl et al., 2004b) to early Oligocene times. The middle Eocene section does contain occasional calcareous nannofossils, but the lithology is predominantly carbonate-low, dark greenish grey to olive grey claystone, with increasing abundance of calcareous nannofossils going upsection, rich in dinocysts throughout the section. High sedimentation rates of >12 cm/kyr evidence steady subsidence of the sub-basins that were formed during faulting of the STR.

Depth (mbsf)

ODP Site 1172

Thanetian	Ypresian		Lutetian	Stage
C26n	C24n C23n	C22n	C21n	Magnetostratigraphy
4	5 6 7 8			Europe[1]
P.gol	A. homomorphum	D.wai W.orn K.ed K.edw	M. perforatum W. echinosuturatum	New Zealand[2]
NZP5	NZE1 NZE2 NZE3 NZE4			New Zealand[3,4]
SPDZ1	SPDZ2 SPDZ3 SPDZ6	SPDZ7 SPDZ8	SPDZ9 SPDZ10 SPDZ11	This study

dinocyst zonations

Hystrichosphaeridium tubiferum
Palaeocystodinium golzowense
Eisenackia reticulata
Eisenackia circumtabulata
Elytrocysta brevis
Vozzhennikovia apertura
Glaphyrocysta ordinata
Glaphyrocysta pastielsii
Thalassiphora delicata
Hystrichosphaeridium truswelliae
Spiniferites sp. B Brinkhuis et al., 2003
Spinidinium schellenbergii
Areoligera senonensis
Lingulodinium machaerophorum
Corrudinium obscurum
Deflandrea phosphoritica
Polysphaeridium subtile
Cassidium filosum
Deflandrea convexa
Apectodinium homomorphum
Deflandrea antarctica
Glaphyrocysta delicata
Phthanoperidinium stockmansii
Thalassiphora pelagica
Florentinia reichartii
Spinidinium macmurdoense
Impagidinium crassimuratum
Samlandia delicata
Glaphyrocysta retiintexta
Impagidinium cassiculum
Vozzhennikovia stickleyae
Batiacasphaera cassicula
Ynezidinium waipawaense
Enneadocysta sp. B
Samlandia septata
Vozzhennikovia cf. *roehliae*
Schematophora spp.
Hystrichokolpoma rigaudiae
Charlesdowniea coleothrypta
Homotryblium tasmaniense
Membranophoridium perforatum
Kisselevia sp. A
Dracodinium waipawaense
Wetzeliella samlandica
Wetzeliella articulata
Octodinium askiniae
Hystrichokolpoma truncatum
Hystrichokolpoma bullatum
Hystrichokolpoma spinosum
Wilsonidium ornatum
Damassadinium crassimuratum
Hystrichostrogylon membraniphorum
Cleistosphaeridium diversispinosum
Charlesdowniea edwardsii
Selenopemphix spp.
Corrudinium regulare
Nematosphaeropsis labirinthus
Arachnodinium antarcticum
Hystrichokolpoma pusillum
Eocladopyxis peniculata
Enneadocysta multicornuta
Phthanoperidinium comatum
Enneadocysta dictyostila
Wilsonidium echinosuturatum

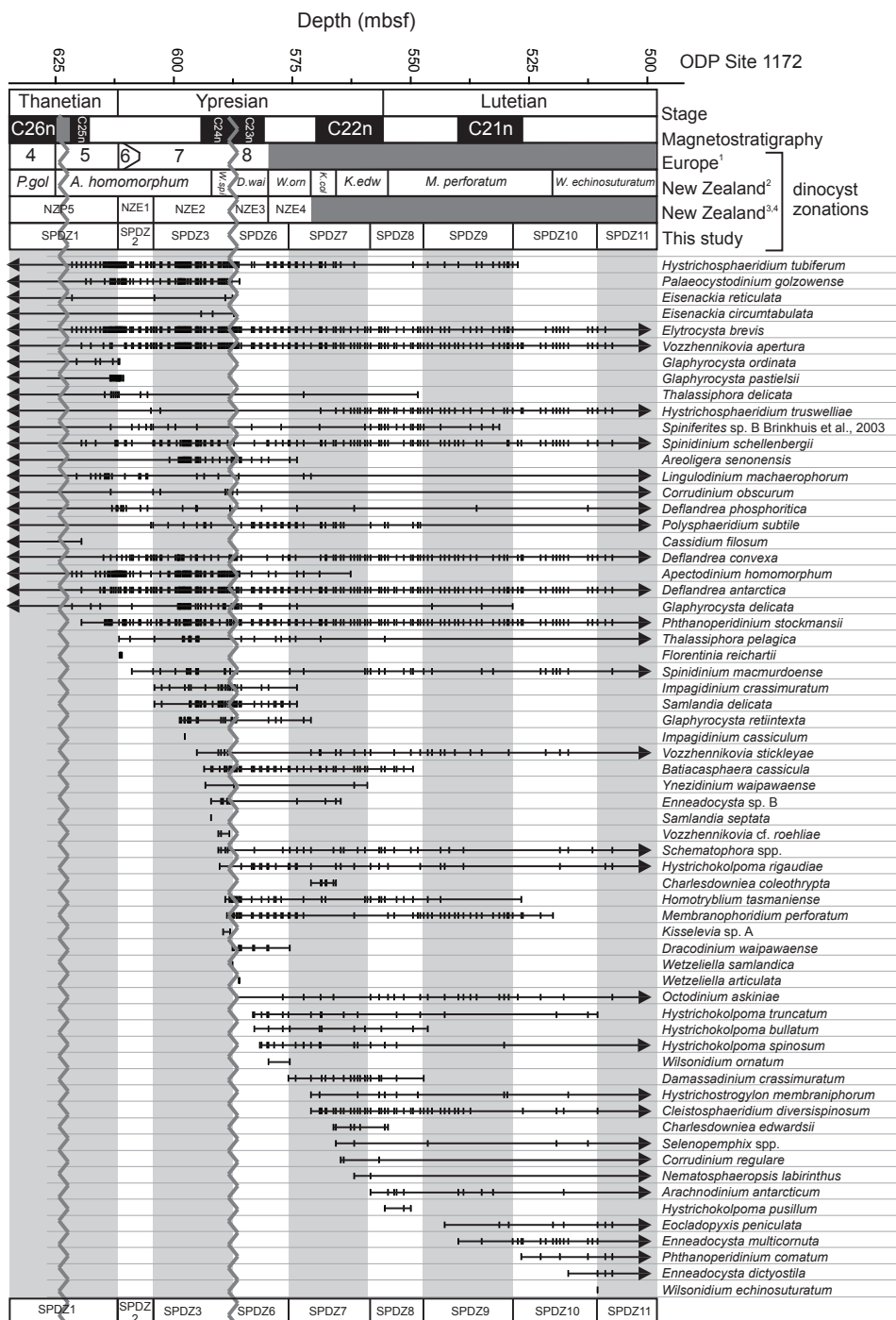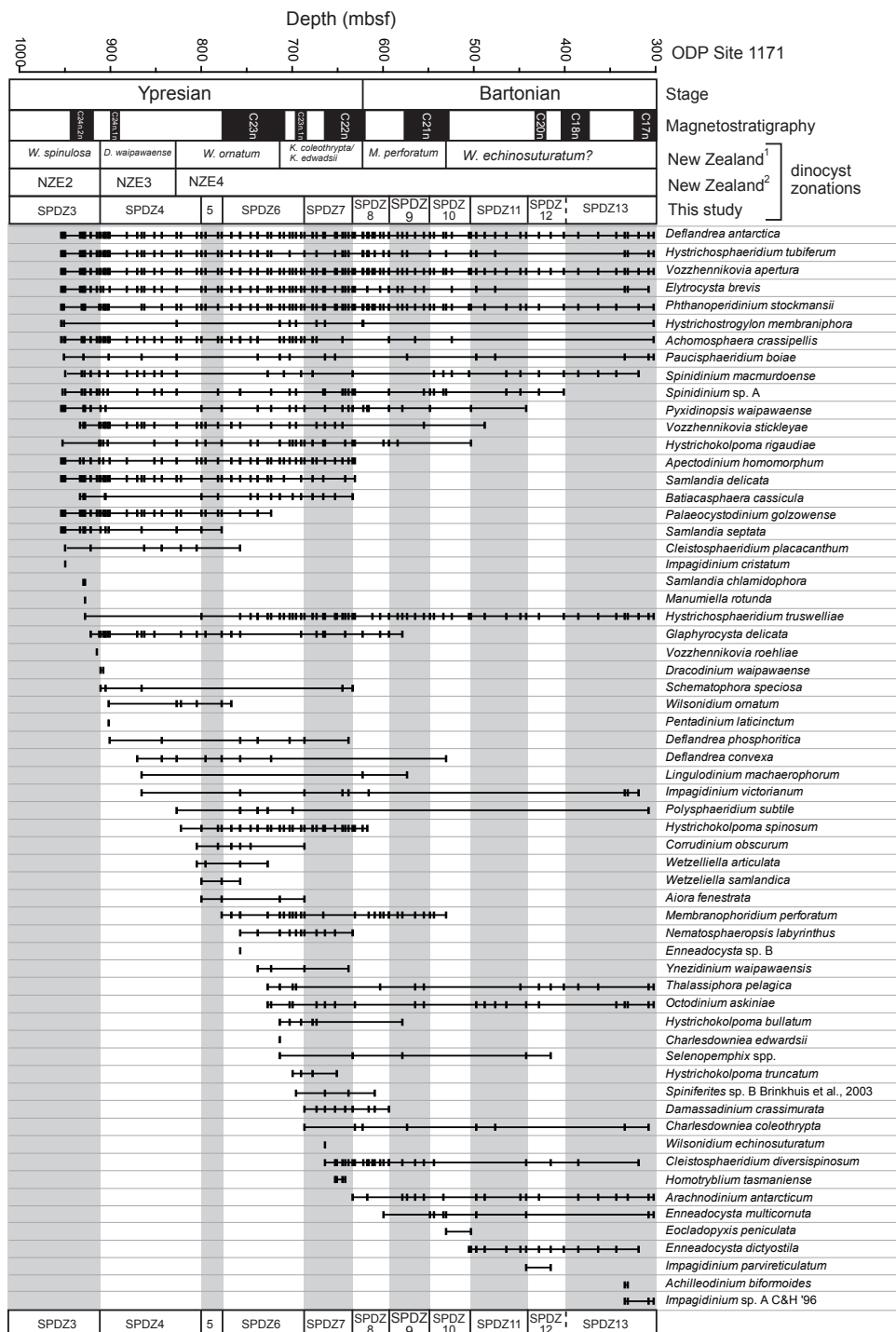| SPDZ1 | SPDZ2 | SPDZ3 | SPDZ6 | SPDZ7 | SPDZ8 | SPDZ9 | SPDZ10 | SPDZ11 |

Figure 3 (this page and previous page). Ranges of dinocyst taxa for ODP Site 1172. a (previous page): Thanetian-Lutetian strata. b (this page): Lutetian – Priabonian strata. Also plotted are the magnetostratigraphic interpretation, and other dinocyst zonations such as those from Heilmann-Clausen (1985)(1), Wilson (1988)(2), Crouch and Brinkhuis (2005)(3) and Crouch (2010)(4).

Figure 4 (next page). Ranges of dinocyst taxa for ODP Site 1171, with magnetostratigraphic interpretation and other dinocyst zonations such as those from Wilson (1988)(1) and Crouch and Brinkhuis (2005)(2).

Depth (mbsf)

ODP Site 1171

Ypresian						Bartonian								Stage

Magnetostratigraphy

W. spinulosa	D. waipawaense	W. ornatum		K. coleothrypta/ K. edwadsii		M. perforatum	W. echinosuturatum?				New Zealand[1]	dinocyst zonations
NZE2	NZE3	NZE4									New Zealand[2]	
SPDZ3	SPDZ4	5	SPDZ6	SPDZ7	SPDZ 8	SPDZ 9	SPDZ 10	SPDZ11	SPDZ 12	SPDZ13	This study	

Deflandrea antarctica
Hystrichosphaeridium tubiferum
Vozzhennikovia apertura
Elytrocysta brevis
Phthanoperidinium stockmansii
Hystrichostrogylon membraniphora
Achomosphaera crassipellis
Paucisphaeridium boiae
Spinidinium macmurdoense
Spinidinium sp. A
Pyxidinopsis waipawaense
Vozzhennikovia stickleyae
Hystrichokolpoma rigaudiae
Apectodinium homomorphum
Samlandia delicata
Batiacasphaera cassicula
Palaeocystodinium golzowense
Samlandia septata
Cleistosphaeridium placacanthum
Impagidinium cristatum
Samlandia chlamidophora
Manumiella rotunda
Hystrichosphaeridium truswelliae
Glaphyrocysta delicata
Vozzhennikovia roehliae
Dracodinium waipawaense
Schematophora speciosa
Wilsonidium ornatum
Pentadinium laticinctum
Deflandrea phosphoritica
Deflandrea convexa
Lingulodinium machaerophorum
Impagidinium victorianum
Polysphaeridium subtile
Hystrichokolpoma spinosum
Corrudinium obscurum
Wetzeliella articulata
Wetzeliella samlandica
Aiora fenestrata
Membranophoridium perforatum
Nematosphaeropsis labyrinthus
Enneadocysta sp. B
Ynezidinium waipawaensis
Thalassiphora pelagica
Octodinium askiniae
Hystrichokolpoma bullatum
Charlesdowniea edwardsii
Selenopemphix spp.
Hystrichokolpoma truncatum
Spiniferites sp. B Brinkhuis et al., 2003
Damassadinium crassimurata
Charlesdowniea coleothrypta
Wilsonidium echinosuturatum
Cleistosphaeridium diversispinosum
Homotryblium tasmaniense
Arachnodinium antarcticum
Enneadocysta multicornuta
Eocladopyxis peniculata
Enneadocysta dictyostila
Impagidinium parvireticulatum
Achilleodinium biformoides
Impagidinium sp. A C&H '96

| SPDZ3 | SPDZ4 | 5 | SPDZ6 | SPDZ7 | SPDZ 8 | SPDZ 9 | SPDZ 10 | SPDZ11 | SPDZ 12 | SPDZ13 |

Depth (mbsf)

IODP Site U1356

Ypresian / Lutetian / R.	Stage
C24n.3n / C24n.1n / C22n	Magnetostratigraphy
W.spin / Dracodinium waipawaense	New Zealand[1]
NZE2 / NZE3	New Zealand[2]
SPDZ3 / SPDZ4 / 5 / SPDZ8 / SPDZ9	This study

dinocyst zonations

Deflandrea antarctica
Phthanoperidium stockmansii
Vozzhennikovia apertura
Aiora fenestrata
Heteraulacacysta spp.
Thalassiphora pelagica
Apectodinium homomorphum
Impagidinium crassimuratum
Turbiosphaera filosa
Deflandrea convexa
Impagidinium cassiculum
Elytrocysta brevis
Hystrichosphaeridium tubiferum
Palaeocystodinium golzowense
Samlandia delicata
Spinidinium schellenbergii
Hystrichosphaeridium truswelliae
Rhombodinium subtile
Vozzhennikovia stickleyae
Octodinium askiniae
Samlandia septata
Polysphaeridium spp.
Schematophora obscura
Eocladopyxis sp. A
Eocladopyxis peniculata
Eisenackia circumtabulata
Batiacasphaera sp. A
Melitasphaeridium pseudorecurvatum
Enneadocysta sp. B
Batiacasphaera cassicula
Dracodinium varielongitudum
Spiniferites sp. B Brinkhuis et al., 2003
Hystrichokolpoma truncatum
Cordosphaeridium sp. A
Apectodinium hyperacanthum
Hystrichokolpoma bullatum
Corrudinium obscurum
Wetzelliella spp. (pars.)
Wetzeliella samlandica
Homotryblium tasmaniense
Kisselevia ornata
Kisselevia insolens
Vozzhennikovia roehliae
Damassadinium crassimuratum
Charlesdowniea cf. clathrata
Glaphyrocysta cf. pastielsii
Adnatosphaeridium multispinosum s.s.
Phthanoperidinium alectrolophum
Membranophoridium perforatum
Arachnodinium antarcticum
Areoligera cf. senonensis
Achomosphaera alcicornu
Spinidinium macmurdoense
Wilsonidium ornatum
Rhombodinium rhomboideum
Tectatodinium spp.
Pentadinium laticinctum s.s.
Charlesdowniea edwardsii
Enneadocysta multicornuta
Enneadocysta dictyostila
Malvinia escutiana
Reticulatosphaera acticoronata
Deflandrea sp. A

| SPDZ3 | SPDZ4 | 5 | SPDZ8 | SPDZ9 | |

Figure 5 (previous page). Ranges of dinocyst taxa for Site U1356. The stages are indicated, where R. stands for Rupelian. The shipboard magnetostratigraphy is plotted, with grey areas indicating intervals with no magnetostratigraphic information due to disturbed core or core gaps. Dinocyst zonations are shown from Wilson (1988)(1) and Crouch and Brinkhuis (2005)(2).

2.3.2 Age model Hole 1171D

Site 1171 has been less intensively studied than Site 1172. Despite earlier claims (Exon et al., 2001; Exon et al., 2003) the PETM was not recovered at this site (Röhl et al., 2004b): oldest recovered sediments are correlated to Subchron C24n. Furthermore, many and large core gaps in the mid-middle Eocene interval compromise proper investigations of the expression of the MECO at this site. An integrated magneto- biostratigraphic age model was published in the Scientific Results Volume (Stickley et al., 2004a), combining dinocyst and few calcareous nannoplankton events with magnetostratigraphy. Higher resolution dinocyst analyses since then has increased confidence in the chronology at this site (Fig. 4). Sedimentation rates in the lower Eocene are much higher (12 cm/kyr) compared to Site 1172. Importantly, the record at Site 1171 does cover the interval from mid-Subchron C24n to mid Subchron C23n, an interval truncated in a hiatus at Site 1172 (Chapter 2, 6).

2.4 Site U1356

2.4.1 Site and lithology description Hole U1356A

Site U1356 is located on the Wilkes Land Margin at the transition between the continental rise and the abyssal plain at 4003 meters below sea level. Eocene sediments were recovered between 890 and 1000 mbsf, and consist of of dark green bioturbated claystone that is faintly stratified (millimeter to meter scale) as defined by color variations (light to dark bands), followed by interbedded stratified and massive sandstones, diamictites, silty claystones, and siltstones with graded bedding and parallel lamination (Escutia et al., 2010). While the drill site is now at abyssal depths, it was located on the Antarctic continental shelf in the Eocene, when the Wilkes Land shelf was much broader than at present (Close et al., 2009). The outer shelf submerged during the late Eocene-early Oligocene when rapid spreading widened the Australo-Antarctic Gulf.

2.4.2 Age model Hole U1356A

Shore-based high-resolution dinocyst analyses (Fig. 5; Chapter 5) have increased confidence in the shipboard integrated dinocyst- and magneto-stratigraphic age model for the Eocene section at Site U1356 (Escutia et al., 2011). The succession covers Subchrons C24n to C23n, followed by strata of middle Eocene age (Magnetochron C22/C21).
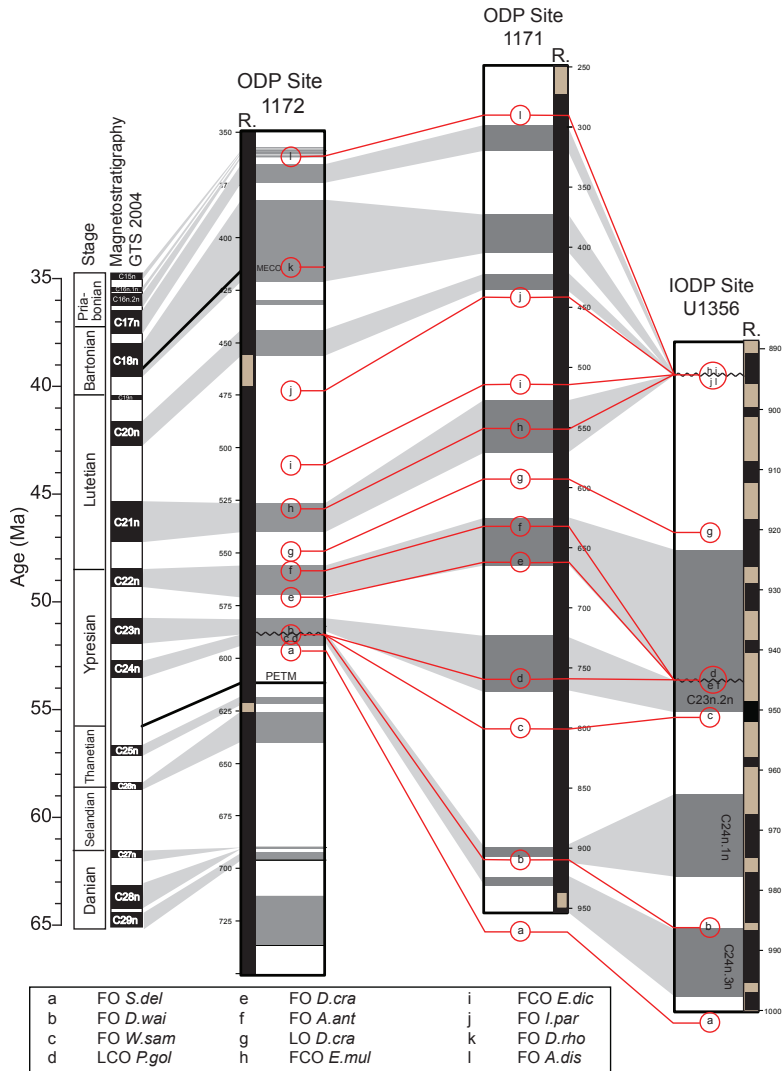
Figure 6. Correlation of Eocene dinocyst events between ODP Sites 1171 and 1172, and IODP Site U1356, and magnetostratigraphic calibration. In red the dinocyst bioevents are indicated: First Occurence (FO) of *Samlandia delicata* (*S.del*), *Dracodinium waipawaense/varielongitudum* (*D.wai*), *Wetzeliella samlandica* (*W.sam*), *Damassadinium crassimuratum* (*D.cra*), *Arachnodinium antarcticum* (*A.ant*), *Impagidinium parvireticulatum* (*I.par*), *Dracodinium rhomboideum* (*D.rho*) and *Alterbidinium distinctum* (*A.dis*); Last Common Occurrence (LCO) of *Paleocystodinium golzowense* (*P.gol*); Last Occurence (LO) of *Damassadinium crassimuratum* (*D.cra*); First Common Occurrence (FCO) of *Enneadocysta multicornuta* (*E.mul*) and *Enneadocysta dictyostila* (*E.dic*). Grey bars indicate intervals with normal magnetic polarity; white indicates reversed polarity. Magnetochrons are correlated between the sites based on the dinocyst events in red. Identification of the K-Pg boundary (Schellenberg et al., 2004) and chemostratigraphic correlations to the Paleocene-Eocene Thermal Maximum (PETM; Chapter 4), and Middle Eocene Climatic Optimum (MECO; Chapter 6) further ties ODP Site 1172 to the Geomagnetic Polarity Time Scale of Gradstein et al., (2004).

3. Methods

3.1 Palynology

Sediments were processed for palynology using standard palynological processing techniques at the Laboratory of Palaeobotany and Palynology (LPP) of Utrecht University. About 10 grams of sediment was collected, freeze dried, and weighed. Processing involved an initial watering with Agepon detergent, followed by addition of hydrochloric acid (10%) to dissolve carbonates and a treatment of hydrofluoric acid (38%) to dissolve silicates. After each acid step samples were washed by decanting after 24 hours (hrs) of settling, followed by filling up with water, centrifuging and decanting again. The hydrofluoric step included 2 hrs shaking at ~250 rotations per minute (rpm) and adding 30% hydrochloric acid to remove fluoride gels. Subsequently, samples were repeatedly washed in water and finally sieved through a 250-micrometer (μm) nylon mesh sieve to remove large palynodebris and pyrite particles. The <250-μm fraction was then sieved using a 15-μm nylon mesh sieve. To break up clumps of residue, samples were placed in an ultrasonic bath for 5 minutes. The residue was transferred to a glass tube and centrifuged for 5 minutes at 2000 rpm. For slide preparation, residues were transferred to vials, homogenized and mounted on a slide with glycerine jelly and sealed. All material is stored in the collection of the LPP. Dinocyst taxonomy follows that cited in Fensome and Williams (2004). Recent additions to the latter include dinocyst species that were described or emended by (Levy and Harwood 2000; Clowes and Wilson 2006; Fensome et al., 2009; Sluijs and Brinkhuis 2009; Sluijs et al., 2009b; Houben et al., 2011).

4. South Pacific Dinocyst zonation

We recognized many key dinocyst events that allow for detailed correlation between Holes 1172A/D and Hole 1171D (Fig. 6; Table 1). These dinocyst events correlate well to the lower Eocene succession of the Wilkes Land Margin (Fig. 6). For the upper Paleocene - lower Eocene succession in Holes 1172A/D and 1171D, key dinocyst events were recognized that allow to correlate to the dinocyst stratigraphy of New Zealand (Crouch and Brinkhuis 2005). Younger sediments at the Tasmanian sector do not have a calibrated reference frame in New Zealand, but can be correlated to an uncalibrated dinocyst zonal scheme (Wilson 1988). The combined information allows for the recognition of inter-regional dinocyst index-events, which define thirteen formal dinocyst zones for the late Paleocene - late Eocene South Pacific Ocean (Fig. 1; Table 1). Many of our dinocyst index events are recognized in New Zealand dinocyst chronologies (Wilson 1988) and stratigraphies (Crouch and Brinkhuis 2005) as well. This means that those dinocyst events have supra-regional significance in the South Pacific Ocean. Where possible, we maintain the events that were used to define zones in New Zealand for our zonation. The top of each zone is defined by a dinocyst event, which also defines the base of the overlying zone. First occurrences (FO), last occurrences (LO), first common (~25%) occurrences (FCO) and last common (~25%) occurrences (LCO) of dinocyst taxa are used as zonal boundaries. We assign a type locality for each dinocyst zone and a base- and top sample for that zone is assigned at the type locality. In addition, the base and top samples in the paratype locality are

36

assigned where applicable. The age calibration of the top of each zone is given to the Geomagnetic Polarity Time Scale of Gradstein et al. (2004), including the associated magneto (sub-) chron. The dinocyst zonation is correlated to stages, with ages according to Gradstein et al. (2004). A comparison of our dinocyst zonation to those from previous studies is presented in Fig. 7.

Dinocyst zone	Type section	bottom sample	top sample	bottom depth	top depth	bottom uncertainty	top uncertainty
SPDZ1	1172D	17R-6W, 70-72 cm	15R-5W, 120-122 cm	633.17	612.7	0.895	0.1
SPDZ2	1172D	15R-5W, 100-102 cm	14R-6W, 40-42 cm	612.5	604.44	0.1	0.15
SPDZ3	1172D	14R-6W, 10-12 cm	13R-2W, 60-62 cm	604.14	587.85	0.15	0.1
SPDZ4	1171D	71R-2W,630-62 cm	59R-CCW, 10-12 cm	912.6	805.32	1.35	2.61
SPDZ5	1171D	59R-4W, 10-12 cm	55R-1W, 10-12 cm	800.1	757.3	2.61	0.025
SPDZ6	1171D	54R-CCW, 9-11 cm	47R-CCW, 17-19 cm	757.25	690.24	0.025	1.865
SPDZ7	1172D	12R-2W, 40-42 cm	10R-3W, 110-112 cm	575.84	559.2	0.61	0.35
SPDZ8	1172D	10R-3W, 40-42 cm	9R-2W, 40-42 cm	558.5	547.3	0.35	0.4
SPDZ9	1172D	9R-1W, 105-107 cm	6R-4W, 135-137 cm	546.5	522.55	0.4	0.535
SPDZ10	1172D	6R-4W, 38-40 cm	5R-3W, 54-56 cm	521.48	510.5	0.535	1.62
SPDZ11	1172A/D	5R-2W, 38-40 cm	51X-4W, 85-87 cm	508.88	475.15	1.62	0.75
SPDZ12	1172A	51X-3W, 85-87 cm	45X-2W, 32-34 cm	473.65	414.02	0.75	0.4
SPDZ13	1172A	45X-1W, 102-104 cm	39R-7W, 45-47 cm	413.22	364.05	0.4	0.07

Table 1. Samples and depths for the zone boundaries in the type locality of the proposed dinocyst zone. SPDZ = South Pacific Dinocyst Zone. The top and bottom depths are in meters below sea floor (mbsf); uncertainties indicate the depth uncertainty (±) in meters.

1. South Pacific Dinocyst Zone (SPDZ) 1
Base definition: LCO *Palaeoperidinium pyrophorum*
Top definition: FCO *Apectodinium homomorphum*
Type locality: ODP Site 1172
Base sample: 1172D-18R-1W, 40-42 cm (634.98 mbsf)
Top sample: 1172D-15R-5W, 120-122 cm (612.7 mbsf)
Calibration: the base is correlated to the base of Subchron C26n. The top of this zone is correlated to just prior to the Paleocene- Eocene boundary, Subchron C24r.
Age: late Thanetian (58.7 Ma to 55.8 Ma)
Characteristic species: *Elytrocysta brevis, Senegalinium* cpx., *Cerebrocysta bartonensis*
Significant bioevents: LO *Eisenackia reticulata*, FO *Cleistosphaeridium placacanthum,* FO *Deflandrea antarctica,* FO *Glapyrocysta delicata*, FO *Phthanoperidinium stockmansii*
Remarks: this zone equates to New Zealand Paleocene Zone 5 of Crouch and Brinkhuis (2005).

2. Zone SPDZ2
Base definition: FCO *Apectodinium homomorphum*
Top definition: FO *Samlandia delicata*
Type locality: ODP Site 1172
Base sample: 1172D-15R-5W, 100-102 cm (612.5 mbsf)

Top sample: 1172D-14R-6W, 40-42 cm (604.44 mbsf)

Calibration: at the type locality (ODP Site 1172), bulk carbon isotope analyses have revealed the stratigraphic position of the Paleocene-Eocene boundary within this dinocyst zone (at 611.8 mbsf). This calibrates the oldest strata within this dinocyst zone to the latest Thanetian. The top of this zone is correlated to the top of Subchron C24r

Age: latest Thanetian- early Ypresian (55.8 to 54.0 Ma)

Characteristic species: *Apectodinium homomorphum, Senegalinium* cpx., *Cordosphaeridium fibrospinosum, Elytrocysta brevis, Deflandrea antarctica, Glaphyrocysta retiintexta, Deflandrea convexa, Hystrichosphaeridium tubiferum*

Significant bioevents: ACME *Apectodinium homomorphum,* FO *Spinidinium macmurdoense*

Remarks: This dinocyst zone includes the quasi-globally recorded dominance of the genus *Apectodinium* associated with the PETM, calibrated with high accuracy using the negative excursion in stable carbon isotopes ($\delta^{13}C$) of the exogenic carbon pool (Crouch et al., 2001; Sluijs et al., 2007a). Regionally, this zone is equivalent to Zone New Zealand Eocene zone 1 (NZE1) of (Crouch and Brinkhuis 2005). Centimeter-scale dinocyst and stable carbon isotope studies of expanded PETM sections have revealed that the *Apectodinium* acme precedes the carbon isotope excursion by several thousands of years in New Jersey, in the North Sea and perhaps at Tawanui, New Zealand (Sluijs et al., 2007b), and in the Arctic Ocean (Harding et al., 2011). At Site 1172, the type section of this dinocyst zone, the acme of *Apectodinium* at the CIE is preceded by a separate latest Paleocene acme, prior to the carbon isotope excursion of the PETM (Chapter 4). Because of the limited constraints on sedimentation rates, it remains uncertain if that early acme correlates to the pre-CIE acmes at other sites or if regional environmental forcing caused some degree of diachronism (Chapter 4). Either way, the resulting maximum uncertainty is in the order of 10^3 or 10^4 years, much smaller than the uncertainty related to the other zones described here. An acme of *Deflandrea* spp. in sample 1172D-15R-1, 10-12 cm (605.6 mbsf) likely correlates to a subzone proposed for New Zealand.

3. Zone SPDZ3

Base definition: FO *Samlandia delicata*

Top definition: FO *Dracodinium waipawaense*

Type locality: ODP Site 1172

Base sample: 1172D-14R-6W, 10-12 cm (604.14 mbsf)

Top sample: 1172D-13R-2W, 60-62 cm (587.85 mbsf); 1171D-71R-4W, 30-32 cm (915.3 mbsf)

Calibration: the top of this zone is truncated by a hiatus at ODP Site 1172; ODP Site 1171 indicates a tentative correlation to the top of Subchron C24n for this zone. At the Wilkes Land Margin, the top of this dinocyst zone is correlated to the top of Subchron C24n.3n

Age: mid-Ypresian (54.0 Ma to 53.3 Ma)

Characteristic species: *Apectodinium homomorphum*, *Areoligera* spp., *Deflandrea antarctica*, *Membranosphaera* spp., *Thalassiphora pelagica, Cordosphaeridium fibrospinosum*, *Palaeocystodinium golzowense*

Significant bioevents: FO *Enneadocysta* sp. B, LO *Eisenackia circumtabulata*

Remarks: This zone equates to dinocyst zone NZE2 of Crouch and Brinkhuis (2005), indicating the supra-regional stratigraphic significance of the dinocyst species that define the zone. These species have also been recognized in the Australo-Antarctic Gulf. However, on that side of the Tasmanian Gateway, *Dracodinium varielongitudum*, a taxon characteristic for the Northern Hemisphere, has been identified rather than *D. waipawaense*. The differences between the two species are subtle at most, and the stratigraphic ranges are identical. Hence it can be debated whether the subtle differences are eco-phenotypical in nature or have evolutionary significance. We consider these species as interchangeable with respect to the zonation.

4. Zone SPDZ4

Base definition: FO *Dracodinium waipawaense*
Top definition: FO *Wetzeliella samlandica*
Type locality: ODP Site 1171
Base sample: 1171D-71R-2W, 60-62 cm (912.6 mbsf); 1172D-13R-2W, 60-62 cm (587.85 mbsf)
Top sample: 1171D-59R-CCW, 10-12 cm (805.32 mbsf); 1172D-13R-2W, 40-42 cm (587.65 mbsf)
Calibration: At ODP Site 1171 and IODP Site U1356 the top of this zone correlates to the onset of Subchron C23n. This zone is within a core gap at ODP Site 1172
Age: mid-Ypresian (53.3 Ma to 51.9 Ma)
Characteristic species: *Senegalinium* cpx., *Phthanoperidinium stockmansii*, *Deflandrea antarctica*, *Palaeocystodinium golzowense*, *Spinidinium/ Vozzhennikovia* cpx.
Significant bioevents: FO *Polysphaeridium subtile,* FO *Corrudinium obscurum*
Remarks: The base of this dinocyst zone equates to the base of NZE3 of Crouch and Brinkhuis (2005), of which the top is defined by the FO of *Wilsonidium ornatum*. While this species is recognized at both Site 1171 and 1172, our calibration against magnetostratigraphy shows that this species apparently has a diachronous first occurrence in the region. We prefer *Wetzeliella samlandica* as a zonal boundary definition because this species apparently has a synchronous first occurrence, also in Wilkes Land. Unfortunately, *Wetzeliella samlandica* has not been reported from New Zealand thus far.

5. Zone SPDZ5

Base definition: FO *Wetzeliella samlandica*
Top definition: LCO *Palaeocystodinium golzowense*
Type locality: ODP Site 1171
Base sample: 1171D-59R-4W, 10-12 cm (800.1 mbsf); 1172D-13R-2W, 60-62 cm (587.85 mbsf)

Top sample: 1171D-55R-1W, 10-12 cm (757.3 mbsf); 1172D-13R-2W, 40-42 cm (587.65 mbsf)

Calibration: at ODP Site 1171 the top of this zone correlates to the top of Subchron C23n. This zone is within a hiatus at ODP Site 1172

Age: mid-Ypresian (51.9 Ma to 50.7 Ma)

Characteristic species: *Senegalinium* cpx., *Phthanoperidinium stockmansii*, *Deflandrea antarctica*, *Palaeocystodinium golzowense*, *Wetzeliella samlandica*, *W. articulata*

Significant bioevents: FO *Polysphaeridium subtile,* FO *Membranophoridium perforatum,*

LO *Samlandia septata*, FO *Ynezidinium waipawaense*

Remarks: The last common occurrence of *Palaeocystodinium golzowense* correlates well between the Tasmanian sector, uncalibrated records from New Zealand sections and tentatively to Wilkes Land.

6. Zone SPDZ6

Base definition: LCO *Palaeocystodinium golzowense*

Top definition: FO *Damassadinium crassimuratum*

Type locality: ODP Site 1171

Base sample: 1171D-54R-CCW, 9-11 cm (757.25 mbsf); 1172D-13R-2W, 40-42 cm (587.65 mbsf)

Top sample: 1171D-47R-CCW, 17-19 cm (690.24 mbsf); 1172D-12R-3W, 10-12 cm (577.06 mbsf)

Calibration: the base of this zone is truncated by a hiatus at ODP Site 1172. At ODP Site 1171 and 1172 the top of this zone is correlated to the onset of Subchron C22n

Characteristic species: *Polysphaeridium subtile, Wetzeliella samlandica, Wilsonidium ornatum, Membranophoridium perforatum, Hystrichokolpoma* spp.

Age: late Ypresian (50.7 Ma to 49.4 Ma)

Significant bioevents: FO *Selenopemphix* spp., FO *Hystrichokolpoma truncatum*

Remarks: The base of the youngest dinocyst zone of Crouch and Brinkhuis (2005) is defined by the FO of *Wilsonidium ornatum*. The FO of this species is diachronous between ODP Sites 1171 and 1172. The FO of other key species for the late Ypresian of New Zealand, such as *Charlesdowniea coleothrypta* (as *Kisselovia coleothrypta*) and *Charlesdowniea edwardsii* (as *Kisselovia edwardsii*) (Wilson 1988) were recognized at ODP Sites 1171 and 1172, but are not synchronous. The FO of *Damassadinium crassimuratum*, however, correlates well between 1171 and 1172 is also reported from the uncalibrated sections in New Zealand (as *Danea crassimurata* (Wilson 1988)). We hence consider this the best-suited key dinocyst event to employ for our zonation.

7. Zone SPDZ7

Base definition: FO *Damassadinium crassimuratum*

Top definition: FO *Arachnodinium antarcticum*

Type locality: ODP Site 1172

Base sample: 1172D-12R-2W, 40-42 cm (575.84 mbsf); 1171D-47R-6W, 10-12 cm (686.51 mbsf)

Top sample: 1172D-10R-3W, 40-42 cm (558.5 mbsf); 1171D-42R-5W, 10-12 cm (638.1 mbsf)

Calibration: the top of this zone is correlated to the top of Subchron C22n

Age: late Ypresian (49.7 Ma to 48.6 Ma)

Characteristic species: *Cleistosphaeridium diversispinosum, Deflandrea antarctica, D. convexa*, *Damassadinium crassimuratum*

Significant bioevents: LO *Apectodinium homomorphum*, FO *Charlesdowniea coleothrypta,* FO *C. edwardsii*, LO *Samlandia delicata*

Remarks: *Arachnodinium antarcticum* is a distinct dinocyst species that is found as a characteristic element of the transantarctic flora (Wrenn and Beckman 1982) from basal middle Eocene sediments upwards on the Antarctic shelf (Wrenn and Hart 1988). However, this species has not been found in New Zealand. One dinocyst event that roughly coincides with the FO of *Arachnodinium antarcticum* and that does allow for correlation to New Zealand is the LO of *Apectodinium homomorphum* (Hollis et al., 2009). We however do not consider this dinocyst event to be a useful zone definition, because this species reaches high abundance in the lower Eocene and hence is very likely to be found reworked in younger sediments.

8. Zone SPDZ8

Base definition: FO *Arachnodinium antarcticum*

Top definition: LO *Damassadinium crassimuratum*

Type locality: ODP Site 1172

Base sample: 1172D-10R-3W, 40-42 cm (558.5 mbsf); 1171D-42R-2W, 10-12 cm (633.6 mbsf)

Top sample: 1172D-9R-2W, 40-42 cm (547.3 mbsf); 1171D-38R-1W, 10-12 cm (593.7 mbsf)

Calibration: the top of this zone is correlated to mid-Subchron C21r

Age: early Lutetian (48.6 Ma to 47.9 Ma)

Characteristic species: *Arachnodinium antarcticum, Damassadinium crassimuratum Deflandrea antarctica, Vozzhennikovia/Spinidinium* cpx*., Phthanoperidinium* spp.

Significant bioevents: LO *Batiacasphaera cassicula,* LO *Polysphaeridium subtile,* FCO *Octodinium askiniae,* LO *Charlesdowniea edwardsii*

Remarks: Also the LO of *Damassadinium crassimuratum* correlates well between ODP Sites 1171 and 1172. Because the range of this species is relatively short, the likelihood of reworking is small. Therefore, this LO is a robust zonal boundary criterion.

9. Zone SPDZ9

Base definition: LO *Damassadinium crassimuratum*

Top definition: FCO *Enneadocysta multicornuta*

Type locality: ODP Site 1172

Base sample: 1172D-9R-1W, 105-107 cm (547.3 mbsf); 1171D-37R-CCW, 17-19 cm (593.56 mbsf)

Top sample: 1172D-6R-4W, 135-137 cm (522.55 mbsf); 1171D-33R-CCW, 0-2 cm (555.02 mbsf)

Calibration: the top of this zone is correlated to mid-to top Subchron C21n

Age: early Lutetian (47.9 Ma to 46.8 Ma)

Characteristic species: *Arachnodinium antarcticum*, *Deflandrea antarctica*, *Membranophoridium perforatum*, *Vozzhennikovia apertura*

Significant bioevents: FO *Enneadocysta multicurnuta*

Remarks: The FO of *Enneadocysta multicornuta* is near the top of the previous zone, and subsequently increases in abundance. We prefer to use the FCO of this species as a zone definition rather than its first occurrence because that has a more robust FO compared to the FO.

10. Zone SPDZ10

Base definition: FCO *Enneadocysta multicornuta*

Top definition: FCO *Enneadocysta dictyostila*

Type locality: ODP Site 1172

Base sample: 1172D-6R-4W, 38-40 cm (521.48 mbsf); 1171D-33R-3W, 10-12 cm (548.6 mbsf)

Top sample: 1172D-5R-3W, 54-56 cm (510.5 mbsf); 1171D-30R-CCW, 12-14 cm (524.58 mbsf)

Calibration: the top of this zone is correlated to early Subchron C20r

Age: mid- Lutetian (46.8 Ma to 45.8 Ma)

Characteristic species: *Enneadocysta multicornuta, Deflandrea convexa*

Significant bioevents: FO *Enneadocysta dictyostila,* LO *Membranophoridium perforatum*

Remarks: This zone and all overlying zones are characterized by monotypic assemblages of dinocysts, alternating by *Deflandrea antarctica*, *Enneadocysta multicornuta* and *E. dictyostila*. Because of the low diversity of the samples, few dinocyst events are recognized.

11. Zone SPDZ11

Base definition: FCO *Enneadocysta dictyostila*

Top definition: FO *Impagidinium parvireticulatum*

Type locality: ODP Site 1172

Base sample: 1172D-5R-2W, 38-40 cm (508.88 mbsf); 1171D-28R-CCW, 5-7 cm (505.73 mbsf)

Top sample: 1172A-51X-4W, 85-87 cm (475.15 mbsf); 1171D-22R-CCW, 12-14 cm (448.84 mbsf)

Calibration: the top of this zone is correlated to the top of Subchron C20r

Age: mid-Lutetian (45.8 Ma to 42.8 Ma)

Characteristic species: *Enneadocysta dictyostila, Deflandrea antarctica, Eocladopyxis peniculata, Thalassiphora pelagica*

Significant bioevents: LCO *Deflandrea convexa*

Remarks: Similar to *Enneadocysta multicornuta, E. dictyostila* has a first occurrence in the underlying zone, followed by rapid increase in abundance. We hence prefer to use the FCO of *E. dictyostila* rather than its FO.

12. Zone SPDZ12

  Base definition: FO *Impagidinium parvireticulatum*
  Top definition: FO *Dracodinium rhomboideum*
  Type locality: ODP Site 1172
  Base sample: 1172A-51X-3W, 85-87 cm (473.65 mbsf); 1171D-22R-3W, 10-12 cm (442.7 mbsf)
  Top sample: 1172A-45X-2W, 32-34 cm cm (414.02 mbsf); 1171D: not found
  Calibration: the top of this zone is calibrated to the onset of Subchron C18n, also with use of isotope stratigraphy around the Middle Eocene Climatic Optimum
  Age: mid-Lutetian to early Bartonian (42.8 Ma to 39.4 Ma)
  Characteristic species: *Thalassiphora pelagica, Arachnodinium antarcticum, Deflandrea antarctica, Enneadocysta dictyostila, E. multicornuta*
  Significant bioevents: LO *Cleistosphaeridium diversispinosum*
  Remarks: The LO of *Dracodinium rhomboideum* is also calibrated to the onset of Magnetochron C18n in Northern Hemisphere sites (Eldrett and Harding 2009). That allows for correlation of the top of this zone to global dinocyst stratigraphies.

13. Zone SPDZ13

  Base definition: FO *Dracodinium rhomboideum*
  Top definition: FO *Alterbidinium distinctum*
  Type locality: ODP Site 1172
  Base sample: 1172A-45X-1W, 102-104 cm (413.22 mbsf); 1171D: not found
  Top sample: 1172A-39X-7W, 45-47 cm (364.05 mbsf); 1171D-4R-2W, 55-57 cm (278.26 mbsf)
  Calibration: the top of this zone is correlated to the onset of C16n.2n
  Age: Bartonian-Priabonian (39.4 Ma to 36.3 Ma)
  Characteristic species: *Spinidinium macmurdoense, Phthanoperidinium stockmansii, Vozzhennikovia apertura*, *Enneadocysta* spp., *Glaphyrocysta pastielsii*
  Significant bioevents: FO *Deflandrea cygniformis,* FCO *Spinidinium macmurdoense*
  Remarks: This zone is characterized by the first occurrence of several transantarctic floral elements with short ranges such as *Deflandrea cygniformis.*

5. Concluding remarks

The combined magneto-, isotope- and biostratigraphically calibrated sediment records from ODP Sites 1172 and 1171 in the Tasmanian sector and IODP Site U1356 from the Wilkes Land Margin, Antarctica now provide a solid framework for dinocyst stratigraphy in the Southern Ocean. It also portrays the evolution of Southern Ocean shelf settings from greenhouse to icehouse conditions. We here provide the calibrated ages for key dinocyst events based on the Geomagnetic Polarity Time Scale of Gradstein et al., (2004) in Table 2, as an update compared to those published earlier (Williams et al., 2004; Escutia et al., 2011). Antarctic-endemic as

well as cosmopolitan species are included, and an error is given for each datum. These errors include those caused by calibration uncertainties, sample spacing and core recovery. Ongoing work on the late Eocene - early Oligocene should allow continuation of the dinocyst stratigraphy upwards. The result will be a proper tool for dating Paleogene sediments from marginal marine settings throughout the Southern Ocean.

Dinocyst species	South Pacific Ocean				Other Oceans				
	FO	error	LO	error	FO	error	LO	error	references
Alterbidinium distinctum	36.40	0.20			transantarctic flora				
Apectodinium homomorphum	58.50	0.20	48.60	0.20					
Arachnodinium antarcticum	48.60	0.20	36.00	0.30	transantarctic flora				
Charlesdowniea coleothrypta	49.50	0.10							
Charlesdowniea columna	51.90	0.10			51.90	0.20	49.00	0.40	(Bujak and Mudge 1994)
Charlesdowniea edwardsii	49.00	0.10	48.40	0.10	51.60	0.20	47.20	0.20	(Bujak and Mudge 1994)
Cleistosphaeridium diversispinosum	49.50	0.20	38.50	0.30					
Corrudinium regulare	49.00	0.20			transantarctic flora				
Damassadinium crassimuratum	49.50	0.10	48.40	0.10					
Deflandrea antarctica	57.60	0.30			transantarctic flora				
Deflandrea convexa	57.60	0.30			transantarctic flora				
Dracodinium rhomboideum	39.30	0.10			39.30	0.10			(Eldrett and Harding 2009)
Dracodinium waipawaense/ D. varielongitudum	53.30	0.10	51.50	0.40	53.30	0.10	51.40	0.10	(Bujak and Mudge 1994)
Elytrocysta brevis	65.50	0.30	33..6	0.30					
Enneadocysta dictyostila	44.60	0.30	33.30	0.30	transantarctic flora				
Enneadocysta multicornuta	47.80	0.20	35.20	0.20					
Enneadocysta sp. B	53.10	0.10	51.90	0.10					
Eocladopyxis peniculata	47.20	0.10							
Glaphyrocysta intricata	39.20	0.20	38.50	0.20					
Glaphyrocysta ordinata	64.40	0.20	55.80	0.20					
Histiocysta palla	44.70	0.20							
Homotryblium tasmaniense	51.50	0.40	45.30	0.10					
Hystrichokolpoma bullatum	51.00	0.20	48.30	0.20					
Hystrichokolpoma rigaudiae	53.80	0.20							
Hystrichokolpoma spinosum	51.00	0.20							
Hystrichokolpoma truncatum	51.00	0.20	45.30	0.20					
Hystrichostrogylon membraniphorum	50.00	0.20							
Impagidinium cassiculum	53.90	0.10	53.40	0.10					
Impagidinium parvireticulatum	43.70	0.20	35.50	0.10					
Kisselevia insolens	51.90	0.20							
Lingulodinium machaerophorum	57.20	0.20							
Lophocysta spp.	40.40	0.10	38.50	0.20					
Membranophoridium perforatum	53.50	0.20	45.10	0.20					
Nematosphaeropsis labirinthus	49.40	0.20							
Octodinium askiniae	53.80	0.10			transantarctic flora				
Palaeocystodinium golzowense			51.10	0.20					
Palaeoperidinium pyrophorum	68.20	0.30	58.50	0.20	73.60	0.10	58.80	0.10	(Bujak and Mudge 1994)
Phthanoperidinium comatum	45.30	0.20							
Phthanoperidinium stockmansii	56.70	0.20							
Polysphaeridium subtile	59.00	0.30	38.50	0.30					
Samlandia delicata	54.20	0.20	48.40	0.30					
Schematophora obscura	53.60	0.10	51.90	0.10					
Selenopemphix spp.	43.50	3.00							
Spinidinium macmurdoense	55.50	0.20			transantarctic flora				
Spinidinium schellenbergii	62.10	0.20			transantarctic flora				
Spiniferites sp.B Brinkhuis et al., 2003	62.50	0.20							
Stoveracysta ornata	35.20	0.20							
Thalassiphora delicata	62.50	0.20			58.90	0.10	38.40	0.10	(Bujak et al., 1980; Köthe 1990)
Thalassiphora pelagica	55.80	0.20							
Vozzhennikovia apertura	65.10	0.40			transantarctic flora				
Vozzhennikovia cf. roehliae	53.50	0.20			transantarctic flora				
Vozzhennikovia sticklyeae	53.80	0.10			transantarctic flora				
Wetzeliella articulata	51.50	0.50	51.40	0.50					
Wetzeliella samlandica	52.30	0.10	51.80	0.10					
Wilsonidium echinosuturatum	44.90	0.20	39.90	0.20					
Wilsonidium ornatum	50.40	0.20	49.70	0.20					

Table 2. Magnetostratigraphically calibrated ages of key dinocyst species in the Southern Ocean, compared to calibrations from other sections in the world. Ages are calibrated to the geomagnetic polarity time scale of Gradstein et al. (2004). The errors refer to age uncertainty caused by correlation error, core recovery issues and/or sample spacing.
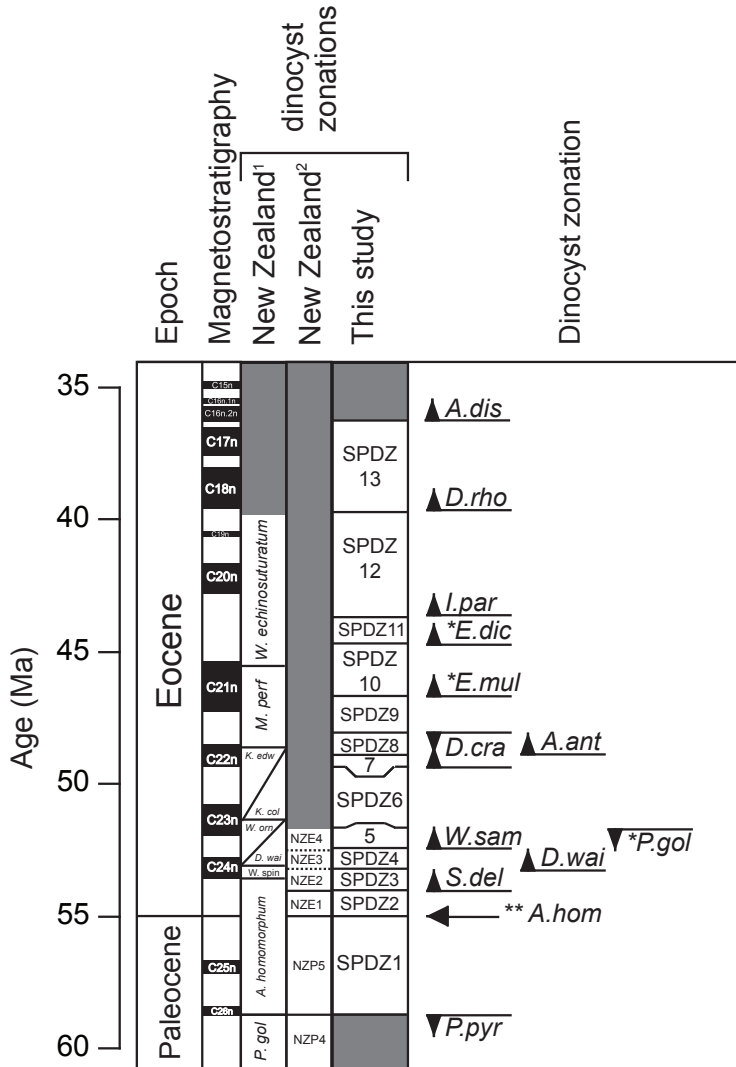
Figure 7. South Pacific dinocyst zonation scheme, with zonal definitions and Southern Ocean dinocyst events within the transantarctic flora. Also indicated are dinocyst zonations from Wilson (1988) (1) and Crouch and Brinkhuis (2005) (2). Abbreviations: *Palaeoperidinium pyrophorum* (*P.pyr*), *Apectodinium homomorphum* (*A.hom*; ** indicates acme), *Samlandia delicata* (*S.del*), *Dracodinium waipawaense/varielongitudum* (*D.wai*), *Wetzeliella samlandica* (*W.sam*), *Palaeocystodinium golzowense* (*P.gol*), *Damassadinium crassimuratum* (*D.cra*), *Arachnodinium antarcticum* (*A.ant*), *Enneadocysta multicornuta* (*E.mul*), *Enneadocysta dictyostila* (*E.dic*), *Impagidinium parvireticulatum* (*I.par*), *Dracodinium rhomboideum* (*D.rho*) and *Alterbidinium distinctum* (*A.dis*).

Dinoflagellate cyst species list

Achilleodinium biformoides

Achomosphaera crassipellis
Achomosphaera alcicornu
Achomosphaera ramulifera [Plate I; a]

Adnatosphaeridium multispinosum
Adnatosphaeridium sp. A [Plate I; b, c]
> Remarks: this species of *Adnatosphaeridium* is characterized by having processes that are united by traberculae connecting perforate distal platforms.

Aiora fenestrata

Aireiana spp.

Alterbidinium acutulum [Plate I; d]
Alterbidinium distinctum

Apectodinium homomorphum [Plate I; e, f, Plate VII; j]
Apectodinium hyperacanthum [Plate I; g, h, Plate VII; k]
Apectodinium parvum
Apectodinium quinquelatum

Apteodinium australiense

Arachnodinium antarcticum [Plate I; i, j]

Areoligera senonensis
Areoligera sentosa
Areoligera spp. [Plate I; k]

Areosphaeridium arcuatum

Batiacasphaera cassicula [Plate I; l]

Brigantedinium spp.

Cannosphaeropsis spp.

Cassidium filosum
Cassidium fragile

Cerebrocysta bartonensis
Cerebrocysta delicata comb. nov. [Plate I; m]

46

Remarks: Fensome et al. (2009) stated that the genus *Cerebrocysta* is "...very difficult to differentiate from *Pyxidinopsis*, originally described by Habib (1976)". Since the species assigned to *Pyxidinopsis* can be divided into a Jurassic and Neocomian group and a Cenozoic group, and *Cerebrocysta* only occurs in the Cenozoic, they preferred to assign the Cenozoic species of *Pyxidinopsis* to *Cerebrocysta*. In their paper, Fensome et al. (2009) only relocated *Pyxidinopsis waipawaensis* to *Cerebrocysta*. In agreement with their emendation of *P. waipawaensis*, we hereby transfer the mid-Late Eocene *Pyxidinopsis delicata* to *Cerebrocysta*.

Cerebrocysta waipawaensis comb. nov. [Plate I; n]

Remarks: this species has been assigned to Cerebrocysta following (Fensome et al., 2009). For explanation see remarks under *Cerebrocysta delicata.*

Charlesdowniea clathrata
Charlesdowniea clathrata [Plate I; o]
Charlesdowniea coleothrypta [Plate II; a, b]
Charlesdowniea columna
Charlesdowniea edwardsii [Plate II; c]

Cleistosphaeridium diversispinosum [Plate II; d]
Cleistosphaeridium placacanthum

Cordosphaeridium fibrospinosum [Plate II; e, Plate VII; l]
Cordosphaeridium gracile
Cordosphaeridium minimum
Cordosphaeridium? sp. A [Plate VII; m]

Remarks: this species is characterized by having, fibrous, sponge-like antapical processes, of which the antapical is elongate and slim. This species is tentatively assigned to *Cordosphaeridium*, and will be formally described elsewhere.

Corrudinium incompositum
Corrudinium obscurum
Corrudinium regulare (Clowes and Wilson 2006) [Plate II; f]

Cribroperidinium sp. A [Plate II; g]

Remarks: this species of *Cribroperidinium* is characterized by having a dark, thick shagreenate to scabrate wall.

Cribroperidinium spp.

Damassadinium crassimuratum [Plate II; h, i]

Dapsilidinium pastielsii

Deflandrea antarctica [Plate II; j]
Deflandrea convexa [Plate III; k]
Deflandrea cygniformis

Deflandrea phosphoritica
Deflandrea sp. A (Brinkhuis et al., 2003a)

Diphyes colligerum [Plate II; l]

Distatodinium paradoxum

Dracodinium rhomboideum
Dracodinium varielongitudum [Plate II; m]
Dracodinium waipawaense [Plate II; n]

Eisenackia circumtabulata [Plate II; o]
Eisenackia reticulata

Elytrocysta brevis [Plate III; a, Plate VII; n]

Enneadocysta brevistila
Enneadocysta dictyostila [Plate III; b, c]
Enneadocysta multicornuta [Plate III; d]
Enneadocysta pectiniformis
Enneadocysta sp. A (Brinkhuis et al., 2003a)
> Remarks: this species resembles *Enneadocysta dictyostila*, with the exception of the two, rather than one antapical processes that are distally connected into a perforate platform (see Brinkhuis et al., 2003a).
Enneadocysta sp. B [Plate III; e, f]
> Remarks: This species resembles *E. multicornuta*, but differs in having bifurcations on many larger processes mid-way the process. The stratigraphic occurrence of this species (Magnetochron C24) is also much older than that reported for *E. multicornuta* (Magnetochron C21).

Eocladopyxis peniculata [Plate III; g]
Eocladopyxis? sp. A [Plate III; h, i]
> Remarks: This dinocyst is questionably assigned to *Eocladopyxis*. The archeopyle seems to involve the apical plates and occasionally the precingular plates as well. It has a coarsely perforate surface, but lacks the processes typical for *E. peniculata*.

Fibrocysta bipolaris

Florentinia reichartii (Sluijs and Brinkhuis 2009)

Gelatia inflata

Glaphyrocysta delicata [Plate III; j]
Glaphyrocysta divaricata
Glaphyrocysta exuberans
Glaphyrocysta ordinata

Glaphyrocysta pastielsii [Plate III; k]
Glaphyrocysta retiintexta

Hafniasphaera septata

Hemiplaciphora semilunifera

Heteraulacacysta porosa
Histiocysta palla [Plate III; l]

Homotryblium tasmaniense [Plate III; m, n]

Hystrichokolpoma bullatum [Plate III; o, Plate IV; a]
Hystrichokolpoma pusillum
Hystrichokolpoma rigaudiae [Plate IV; b]
Hystrichokolpoma salacia
Hystrichokolpoma spinosum
Hystrichokolpoma truncatum [Plate IV; c]

Hystrichosphaeridium truswelliae [Plate IV; d, Plate VII; o]
Hystrichosphaeridium tubiferum [Plate IV; e]
> Remarks: *Hystrichosphaeridium truswelliae* is distinguished *H. tubiferum* in having hollow open infundibular processes that forms distally fenestrate, edged platforms, whereas the processes of *H. tubiferum* are tubiform. Occasionally, poor preservation of the distal ends of the processes compromises proper speciation.

Hystrichosphaeropsis spp.

Hystrichostrogylon membraniphorum [Plate IV; f]

Impagidinium aculeatum
Impagidinium cassiculum [Plate IV; g]
Impagidinium crassimuratum
Impagidinium dispertitum
Impagidinium maculatum [Plate IV; h, Plate VIII; a]
Impagidinium paradoxum
Impagidinium parvireticulatum [Plate IV; i]
Impagidinium victorianum

Kisselevia insolens [Plate IV; j; Plate VIII; b]
Kisselevia ornata [Plate IV; k]
Kisselevia sp. A [Plate IV; l; Plate VIII; c]
> Remarks: This species consists of 2 layers, which assigns it to *Kisselevia* rather than *Charlesdowniea* (following concepts of Lentin and Vozzhennikova (1989)). However, some features, such as the perforate platforms supported

by short processes, are characteristic of *Charlesdowniea*. The absence of a spherical endocyst assigns this species to *Kisselevia*.

Lejeunecysta hyalina

Lingulodinium machaerophorum

Lophocysta spp.

Mavinia esutiana (Houben et al., 2011)

Melitasphaeridium pseudorecurvatum

Membranophoridium perforatum [Plate IV; m, n]

Nematosphaeropsis labyrinthus

Ochetodinium spp.

Octodinium askiniae [Plate V; a]

Oligosphaeridium spp.

Operculodinium centrocarpum

Palaeocystodinium golzowense [Plate V; b]

Palaeoperidinium pyrophorum [Plate V; c, d]

Paucisphaeridium inversibuccinum
Paucisphaeridium spp. (pars.) [Plate IV; o, Plate V; e, f, late VIII; d, e]
  Remarks: several new species of *Paucisphaeridium* are identified at this site, but these will be formally described elsewhere.

Pentadinium laticinctum [Plate V; g]

Phthanoperidinium alectrolophum [Plate V; h]
Phthanoperidinium comatum
Phthanoperidinium crenulatum
Phthanoperidinium stockmansii [Plate V; i, j]
Phthanoperidinium sp. A (Goodman and Ford 1983)
Phthanoperidinium sp. B (Goodman and Ford 1983)

Polysphaeridium subtile [Plate V; k]

Reticulatosphaera acticoronata

Rhombodinium subtile [Plate V; l]
Rhombodinium draco

Samlandia delicata [Plate V; m]
Samlandia septata [Plate VI; a]
Samlandia chlamidophora

Schematophora obscura [Plate VI; b, c]
Schematophora speciosa

Selenopemphix armata
Selenopemphix nephroides

Senegalinium spp. (pars.) [Plate VI; d, e, f, Plate VIII, f, g, h]

Spinidinium luciae
Spinidinium macmurdoense [Plate VI; g, Plate VIII; i]
Spinidinium schellenbergii (Sluijs et al., 2009b) [Plate VI; h, i, Plate VIII; j, k]

Spiniferites pseudofurcatus
Spiniferites ramosus
Spiniferites sp. B (Brinkhuis et al., 2003a) [Plate VI; j]

Stoveracysta kakanuiensis
Stoveracysta ornata

Tectatodinium pellitum [Plate VI; k]

Thalassiphora delicata
Thalassiphora pelagica [Plate VI; l]

Turbiosphaera filosa
Turbiosphaera galeata
Turbiosphaera sagena (Levy and Harwood 2000)

Vozzhennikovia apertura [Plate VI; m, n]
Vozzhennikovia netrona (Levy and Harwood 2000)
Vozzhennikovia cf. *roehliae* (Sluijs et al., 2009b) [Plate VI; o, Plate VII; a]
       Remarks: this species of *Vozzhennikovia* occurs in Subchron C24n, and shows morphological resemblances to *V. roehliae*. Ornamentation (in the form hollow verrucae) is faintly visible, and the cingulum is slightly indent. However, the outline of the cyst is more rounded compared to *V. roehliae* with no apical or antapical horns. While *V. roehliae* has a consistent 3I archeopyle, the archeopyle of this species is consistent 2a, and the operculum is consistently adnate to 4''.
Vozzhennikovia stickleyae (Sluijs et al., 2009b) [Plate VII; b]

Wetzeliella articulata [Plate VII; c]
Wetzeliella astra
Wetzeliella samlandica [Plate VII; d, e, Plate VIII; l, m]

Wilsonidium echinosuturatum [Plate VII; f]
Wilsonidium ornatum [Plate VII; g]

Ynezidinium waipawaense

Dinocyst gen. et sp. indet. [Plate VII; h, I; Plate VIII; n, o]

Remarks: this dinocyst species is spherical in polar view, with a clear edged apical archaeopyle. The offset sulcal nodge and occasionally large antapical bulge clearly indicates dorso-ventrally compression. Based on these observations, this species belongs to the *Areoligera* genus. Species within *Areoligera* spp. are characterized by processes or process complexes indicating para-tabulation. The outer wall of this species however is finely rugulate with no expression of para-tabulation.

Dinocyst plates

Plate I. a: *Achomosphaera ramulifera* 1172D-11R-1W, 113-115 cm b: *Adnatosphaeridium* sp. U1356A-101R-1W, 10-12 cm c: *Adnatosphaeridium* sp. U1356A-101R-1W, 10-12 cm d: *Alterbidinium acutulum* 1172D-10R-2W, 10-12 cm   e: *Apectodinium homomorphum* 1172D-13R-3W, 135-137 cm   f: *Apectodinium homomorphum* U1356A-103R-4W, 20-22 cm g: *Apectodinium hyperacanthum* 1172D-15R-5W, 100-102 cm h: *Apectodinium hyperacanthum* 1172D-15R-5W, 100-102 cm i: *Arachnodinium antarcticum* 1172D-7R-6W, 135-137 cm j: *Arachnodinium antarcticum* U1356A-97R-1W, 60-63 cm k: *Areoligera* sp. 1172D-14R-1W, 55-57 cm l: *Batiacasphaera* sp. 1172D-12R-4W, 10-12 cm m: *Cerebrocysta delicata* 1172D-5R-3W, 38-40 cm n: *Cerebrocysta waipawaensis* 1172D-11R-3W, 135-137 cm o: *Charlesdowniea clathrata* U1356A- 101R-1W, 10-12 cm.

Scale bar = 25 micrometer (μm)

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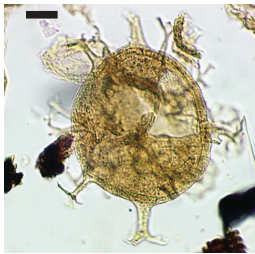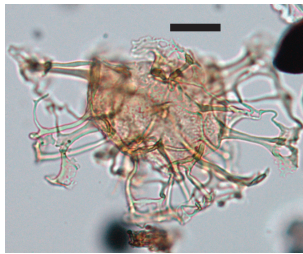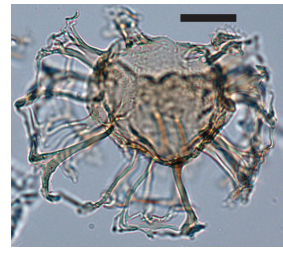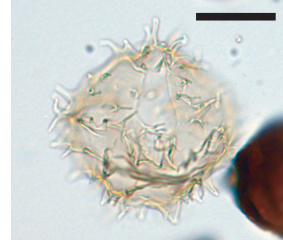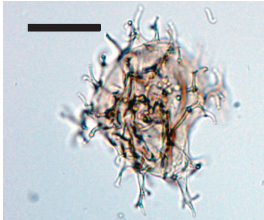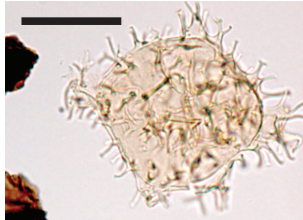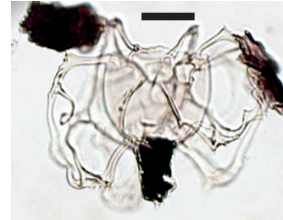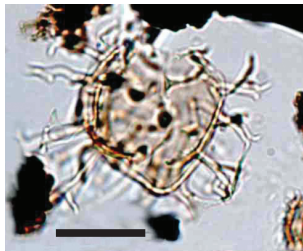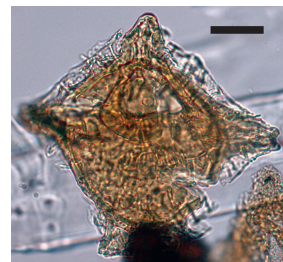
Plate II. a: *Charlesdowniea coleothrypta* 1172D-11R-5W, 40-42 cm b: *Charlesdowniea coleothrypta* U1356A-98R-5W, 0-3 cm   c: *Charlesdowniea edwardsii* 1172D-11R-1W, 111-113 cm   d: *Cleistosphaeridium diversispinosum* 1172D-11R-2W, 10-12 cm e: *Cordosphaeridium fibrospinosum* 1172D-15R-5W, 100-102 cm f: *Corrudinium regulare* 1172A-42X-1W, 85-87 cm g: *Cribroperidinium* sp. A U1356A-105R-6W, 95-98 cm h: *Damassadinium crassimuratum* 1172D-11R-6W, 40-42 cm i: *Damassadinium crassimuratum* 1172D-10R-2W, 10-12 cm j: *Deflandrea antarctica* 1172D-12R-1W, 10-12 cm k: *Deflandrea convexa* 1172D-5R-1W, 40-42 cm l: *Diphyes colligerum* 1172D-13R-2W, 90-92 cm m: *Dracodinium varielongitudum* U1356A-104R-2W, 120-122 cm n: *Dracodinium waipawaense* 1172D-12R-7W, 10-12 cm   o: *Eisenackia circumtabulata* U1356A- 105R-4W, 0-2 cm.

Scale bar = 25 μm
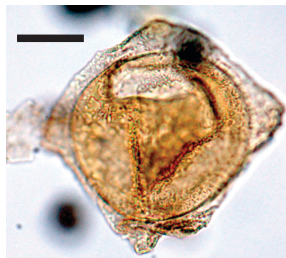
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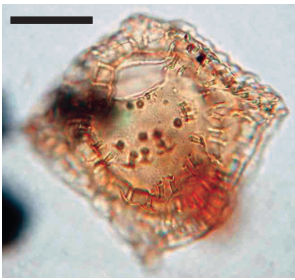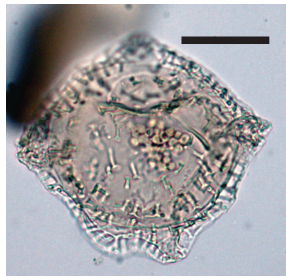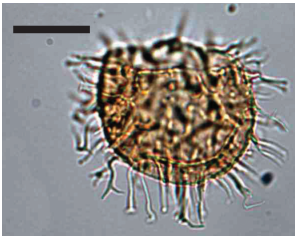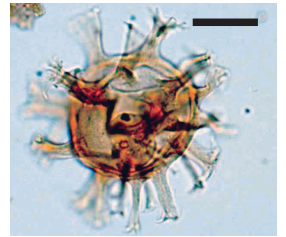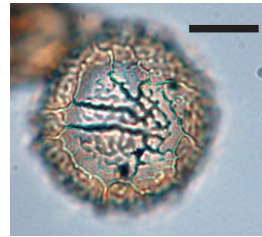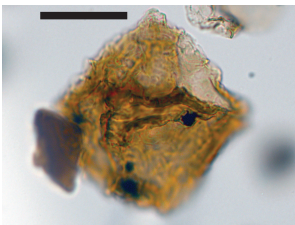
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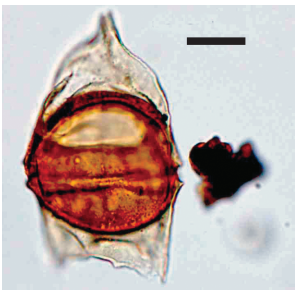
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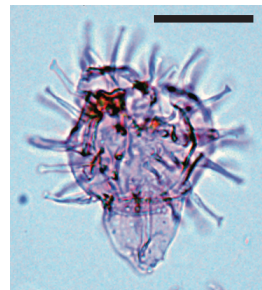
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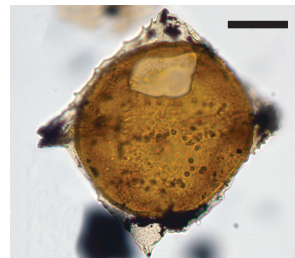
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Plate III. a: *Elytrocysta brevis* 1172D-14R-5W, 10-12 cm b: *Enneadocysta dictyostila* 1172D-5R-3W, 38-40 cm c: *Enneadocysta dictyostila* 1172D-5R-1W, 40-42 cm d: *Enneadocysta multicornuta* 1172D-13R-3W, 105-107 cm e: *Enneadocysta* sp. B U1356A-102R-1W, 10-12 cm f: *Enneadocysta* sp. B U1356A-103R-2W, 40-42 cm g: *Eocladopyxis peniculata* 1172D-5R-4W, 38-40 cm h: *Eocladopyxis* sp. A U1356A- 101R-1W, 60-62 cm i: *Eocladopyxis* sp. A U1356A-101R-1W, 60-62 cm j: *Glaphyrocysta delicata* 1172D-12R-7W, 10-12 cm k: *Glaphyrocysta pastielsii* U1356A-99R-2W, 40-43 cm l: *Histiocysta palla* U1356A-100R-1W, 10-13 cm m: *Homotryblium tasmaniense* 1172D-10R-2W, 10-12 cm n: *Homotryblium tasmaniense* U1356A-101R-1W, 60-63 cm o: *Hystrichokolpoma bullatum* 1172D-12R-2W, 10-12 cm.
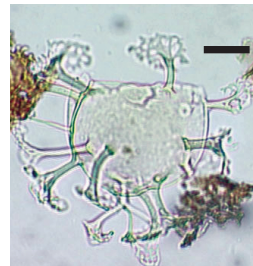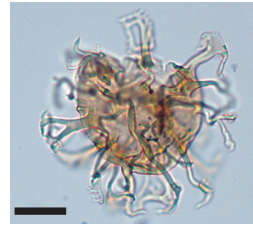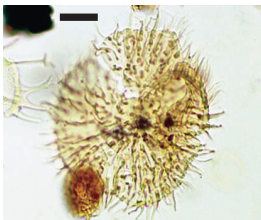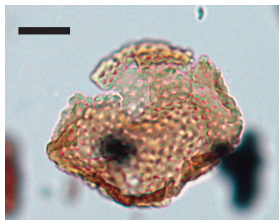
Scale bar = 25 μm

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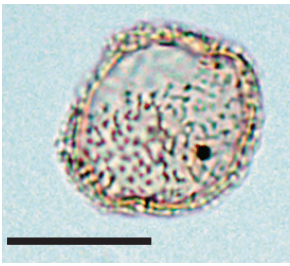
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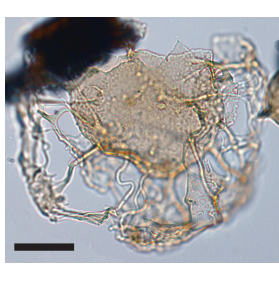
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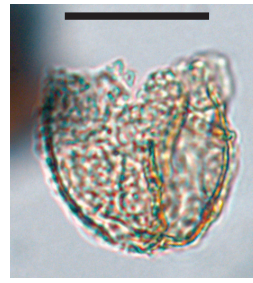
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Plate IV. a: *Hystrichokolpoma bullatum* 1172D-12R-7W, 10-12 cm b: *Hystrichokolpoma rigaudiae* U1356A-103R-1W, 10-12 cm   c: *Hystrichokolpoma truncatum* U1356A-98R-1W, 100-103 cm   d: *Hystrichosphaeridium truswelliae* U1356A-101R-1W, 80-82 cm e: *Hystrichosphaeridium tubiferum* U1356A-98R-1W, 100-103 cm f: *Hystrichos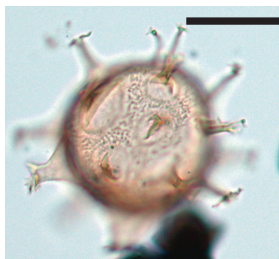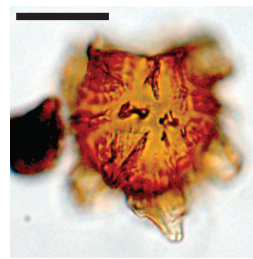togylon membraniphora* 1172D-11R-5W, 40-42 cm g: *Impagidinium cassiculum* U1356A-104R-2W, 40-42 cm   h: *Impagidinium maculatum* 1172D-12R-2W, 10-12 cm   i: *Impagidinium parvireticulatum* 1172D-10R-2W, 10-12 cm j: *Kisselovia insolens* 1172D-13R-2W, 90-92 cm k: *Kisselovia ornata* U1356A-101R-1W, 10-12 cm   l: *Kisselovia* sp. A 1172D-13R-2W, 40-42 cm   m: *Membranophoridium perforatum* 1172D-11R-6W, 104-106 cm n: *Membranophoridium perforatum* 1172D-12R-7W, 65-67 cm o: *Paucisphaeridium* sp. 1172D-14R-1W, 60-62 cm.

Scale bar = 25 μm
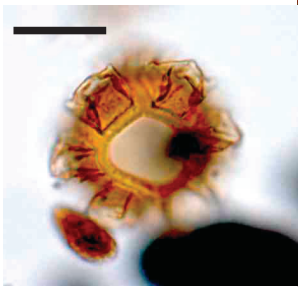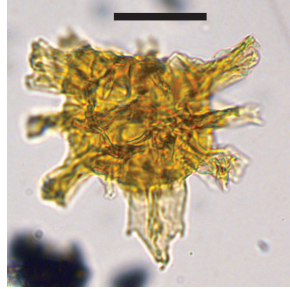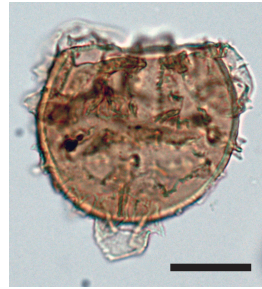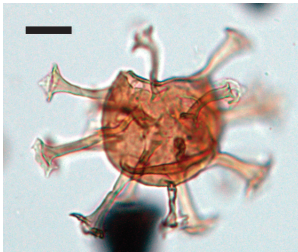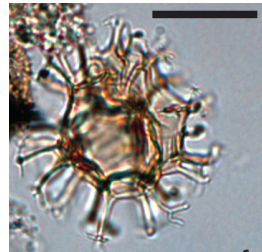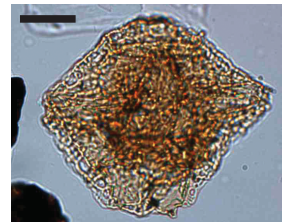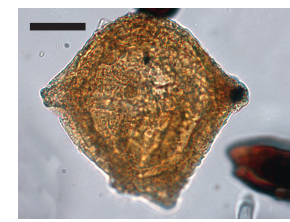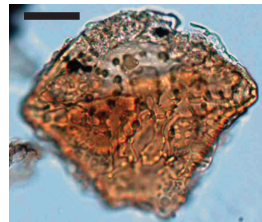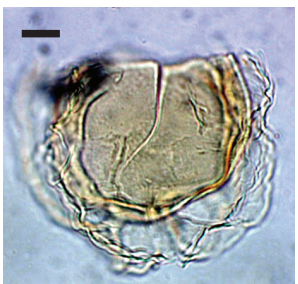
Plate V. a: *Octodinium askiniae* 1172D-10R-2W, 10-12 cm b: *Palaeocystodinium golzowense* U1356A- 104R-1W, 50-52 cm c: *Palaeoperidinium pyrophorum* 1172D-24R-5W, 38-40 cm d: 1172D-24R-5W, 20-22 cm e: *Paucisphaeridium* sp. 1172D-14R-2W, 150-152 cm f: *Paucisphaeridium* sp. 1172D-14R-2W, 30-32 cm g: *Pentadinium laticinctum* U1356A-100R-1W, 10-13 cm h: *Phthanoperidinium alectrolohpum* U1356A-100R-1W, 104-107 cm   i: *Phthanoperidinium stockmansii* 1172D-9R-6W, 40-42 cm   j: *Phthanoperidinium stockmansii* 1172D-9R-6W, 40-42 cm k: *Polysphaeridium subtile* 1172D-12R-4W, 10-12 cm   l: *Rhombodinium subtile* U1356A-105R-4W, 99-101 cm m: *Samlandia delicata* U1356A-100R-1W, 104-107 cm.

Scale bar = 25 μm
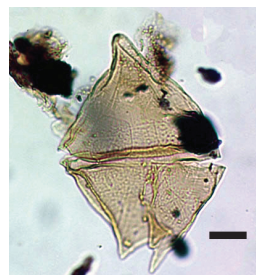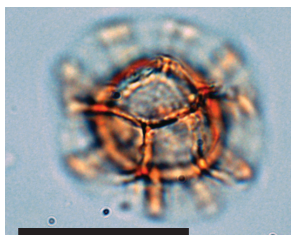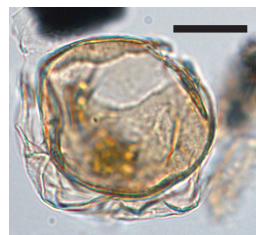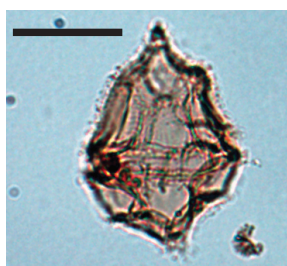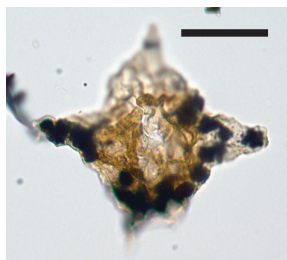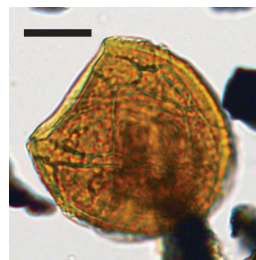
Plate VI. a: *Samlandia septata* U1356A-104R-6W, 70-72 cm   b: *Schematophora obscura* 1172D-13R-3W, 105-107 cm c: *Schematophora obscura* U1356A-102R-1W, 30-32 cm d: *Senegalinium* sp. 1172D-15R-7W, 10-12 cm e: *Senegalinium* sp. 1172D-12R-4W, 10-12 cm f: *Senegalinium* sp. 1172D-15R-7W, 10-12 cm g: *Spinidinium macmurdoense* 1172D-13R-7W, 35-37 cm h: *Spinidinium schellenbergii* U1356A-98R-1W, 100-103 cm i: *Spinidinium schellenbergii* 1172D-6R-2W, 38-40 cm j: *Spiniferites* sp. B Brinkhuis et al., 2003 U1356A-101R-1W, 80-82 cm k: *Tectatodinium pellitum* U1356A-99R-2W, 0-3 cm l: *Thalassiphora pelagica* 1172D-12R-2W, 10-12 cm   m: *Vozzhennikovia apertura* 1172D-18R-1W, 40-42 cm   n: *Vozzhennikovia apertura* 1172D-14R-4W, 10-12 cm o: *Vozzhennikovia* cf. *roehliae* 1172D-13R-3W, 135-137 cm.

Scale bar = 25 μm

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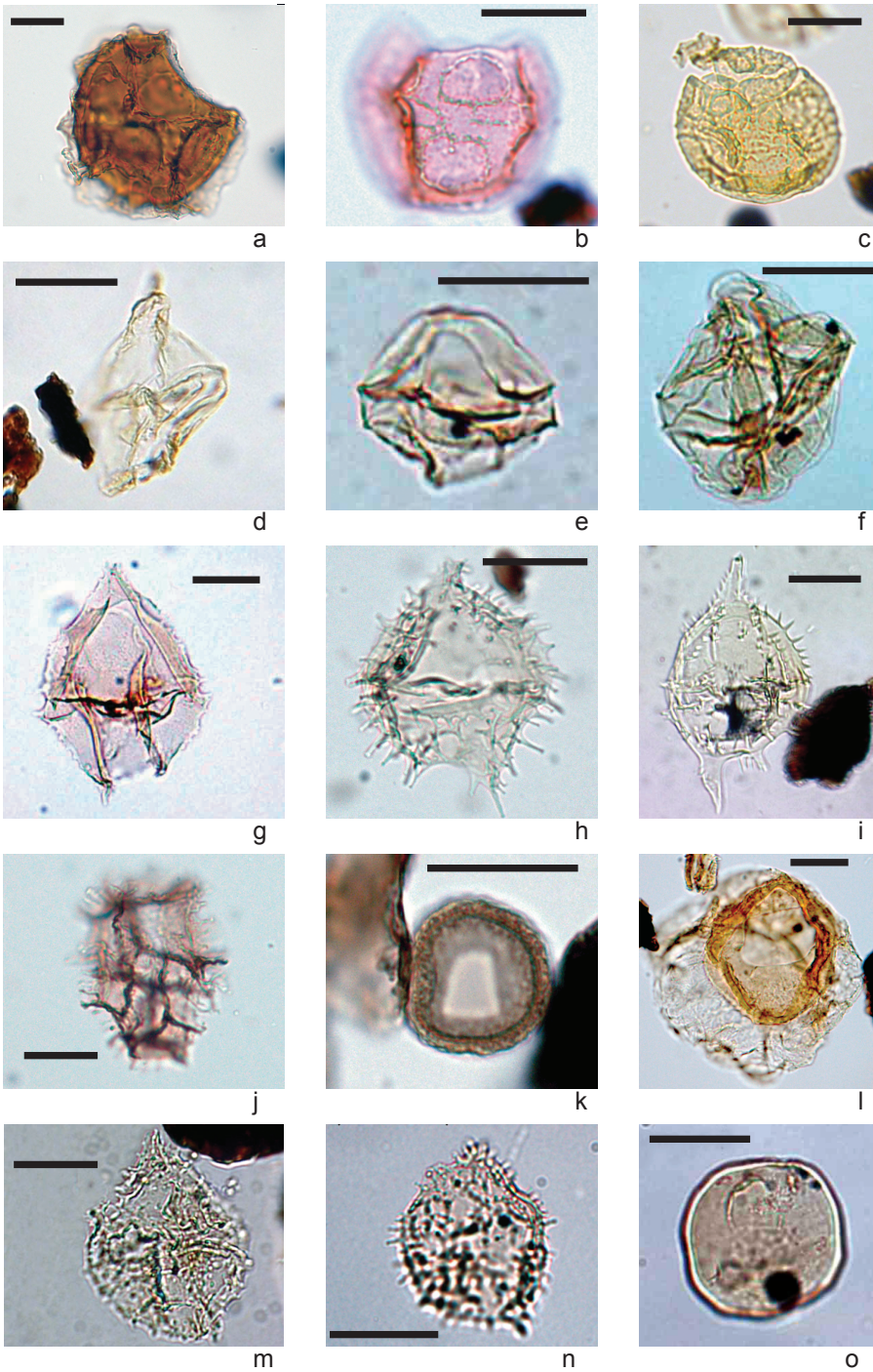
Plate VII. a: *Vozzhennikovia* cf. *roehliae* 1172D-13R-3W, 135-137 cm b: *Vozzhennikovia stickleyae* 1172D-14R-4W, 10-12 cm c: *Wetzeliella astra* 1172D-13R-1W, 40-42 cm d: *Wetzeliella samlandica* 1172D-13R-2W, 40-42 cm e: *Wetzeliella samlandica* 1172D-13R-2W, 40-42 cm f: *Wilsonidium echinosuturatum* 1172D-5R-4W, 38-40 cm g: *Wilsonidium ornatum* 1172D-12R-5W, 40-42 cm h, i: dinocyst gen. et sp. indet. U1356A-104R-6W, 70-72 cm j: *Apectodinium homomorphum* 1172D-15R-5W, 100-102 cm k: *Apectodinium homomorphum* 1172D-15R-5W, 100-102 cm l: *Cordosphaeridium fibrospinosum* 1172D-15R-5W, 10-12 cm m: *Cordosphaeridium* sp. A U1356A-103R-1W, 90-92 cm n: *Elytrocysta brevis* 1172D-14R-2W, 115-117 cm o: *Hystrichosphaeridium tubiferum* 1172D-15R-5W, 100-102 cm.

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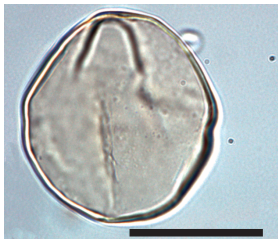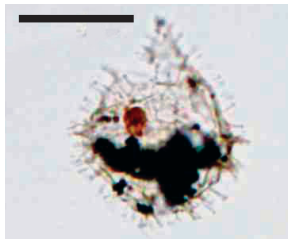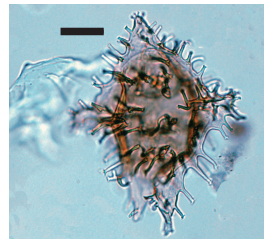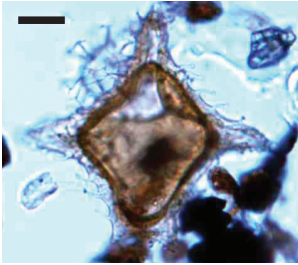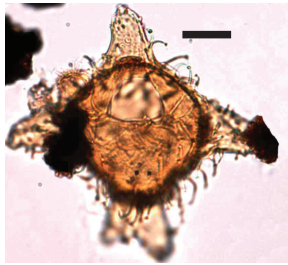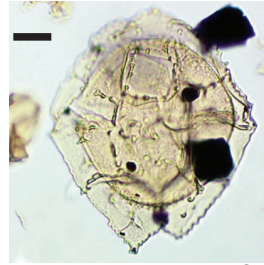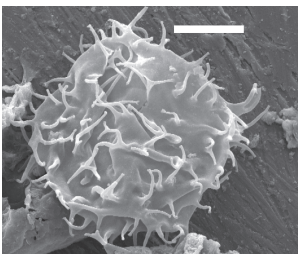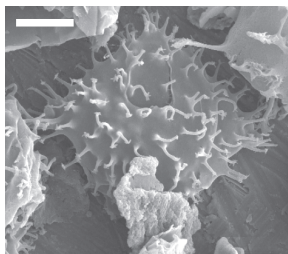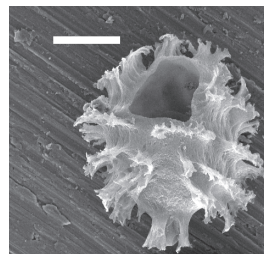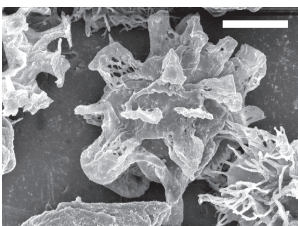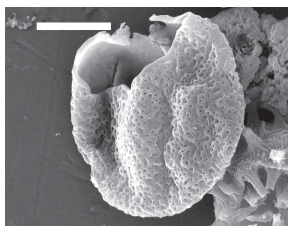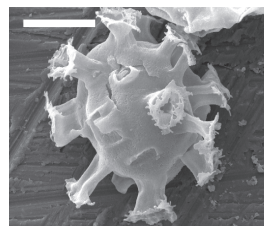
Plate VIII. a: *Impagidinium maculatum* 1172D-14R-4W, 10-12 cm b: *Kisselevia insolens* U1356A-101R-1W, 120-121 cm c: *Kisselevia* sp. A 1172D-13R-2W, 40-42 cm d: *Paucisphaeridium* sp. 1172D-14R-1W, 140-142 cm e: *Paucisphaeridium* sp. 1172D-14R-1W, 140-142 cm f: *Senegalinium* sp. 1172D-14R-4W, 10-12 cm g: *Senegalinium* sp. 1172D-14R-2W, 115-117 cm h: *Senegalinium* sp. 1172D-14R-2W, 115-117 cm i: *Spinidinium macmurdoense* 1172D-14R-4W, 10-12 cm j: *Vozzhennikovia apertura* 1172D-14R-1W, 10-12 cm k: *Vozzhennikovia apertura* 1172D-14R-1W, 10-12 cm l. *Wetzeliella samlandica* 1172D-13R-2W, 40-42 cm m: *Wetzeliella samlandica* 1172D-13R-2W, 40-42 cm n: dinocyst gen. et sp. indet. U1356A-105R-1W, 75-77 cm o: dinocyst gen. et sp. indet. U1356A-104R- 6W, 70-72 cm.

Scale bar = 25 μm

CHAPTER 2

Early Paleogene temperature evolution
of the southwest Pacific Ocean

Dracodinium waipawaense (Wilson, 1967b) Costa and Downie, 1979
More or less rhomboidal pericyst with the cingular horns longer than the apical and
antapical horns. The margin is generally smooth but sometimes serrated near the apical
and antapical horns. Left antapical horn when developed is very small. Periphragm is
punctate. The periarcheopyle extends into the apical pericoel.

Relative to the present day, meridional temperature gradients in the early Eocene age (~56–53 million years ago (Ma)) were unusually low, with slightly warmer equatorial regions (Pearson et al., 2007) but with much warmer subtropical Arctic (Sluijs et al., 2006) and mid-latitude (Sluijs et al., 2007b) climates. By the end of the Eocene epoch (34 Ma), the first major Antarctic ice sheets had appeared (Barker et al., 2007; Zachos et al., 1992), suggesting that major cooling had taken place. Yet the global transition into this icehouse climate remains poorly constrained, as only a few temperature records are available portraying the Cenozoic climatic evolution of the high southern latitudes. Here we present a uniquely continuous and chronostratigraphically well-calibrated TEX_{86} record of sea surface temperature (SST) from an ocean sediment core in the East Tasman Plateau (paleolatitude ~65°S). We show that southwest Pacific SSTs rose above present-day tropical values (to 34°C) during the early Eocene age (53 Ma) and had gradually decreased to about 21°C by the early late Eocene (36 Ma). Our results imply that there was almost no latitudinal SST gradient between sub-equatorial and sub-polar regions during the early Eocene (~55–50 Ma). Thereafter, the latitudinal gradient markedly increased. In theory, if Eocene cooling was largely driven by a decrease in atmospheric greenhouse gas concentration (Zachos et al., 2008), additional processes are required to explain the relative stability of tropical SSTs given that there was more significant cooling at higher latitudes.

The Paleogene temperature evolution of the Antarctic margin, particularly the Pacific sector, is still poorly resolved. One difficulty with obtaining relevant records close to the Antarctic continent is the general absence of biogenic carbonate in most marine facies, which hampers traditional $\delta^{18}O$ and/or Mg/Ca-based reconstructions of the subpolar temperature evolution. In the absence of biogenic carbonates, organic sea-surface-temperature proxies such as the tetraether index of lipids consisting of 86 carbon atoms (TEX_{86}; Schouten et al., 2002) and the alkenone unsaturation index ($U^{K'}_{37}$; Müller, 1998) are required for reconstructing high-latitude climatic evolution (Liu et al., 2009; Sluijs et al., 2006).

We apply TEX_{86} and $U^{K'}_{37}$ on a stratigraphically continuous sedimentary section from the southwest Pacific Ocean, drilled by the Ocean Drilling Program (ODP) Leg 189 Site 1172, paleolatitude 65°S (Exon et al., 2004; Fig. 1). A full methodological description is available in the Supplements. The record contains an expanded succession of marginal marine sediments from the lower Paleocene epoch to the upper Eocene (~64–36 Ma), with tight chronostratigraphic control, including magnetostratigraphy (Stickley et al., 2004a; Supplements; Fig. S2). The presence of typical transantarctic organic-walled dinoflagellate cysts in the Tasmanian region indicates an Antarctic-derived northward-flowing Tasman Current throughout the Paleogene, which is verified by experiments based on a general circulation model (Huber et al., 2004; Fig. 1). This Antarctic influence at the East Tasman Plateau (ETP) persisted until at least the early late Eocene (~35.5 Ma), when deepening of the Tasmanian Gateway lead to a reorganization of the Tasman and proto-Leeuwin ocean currents (Stickley et al., 2004b).
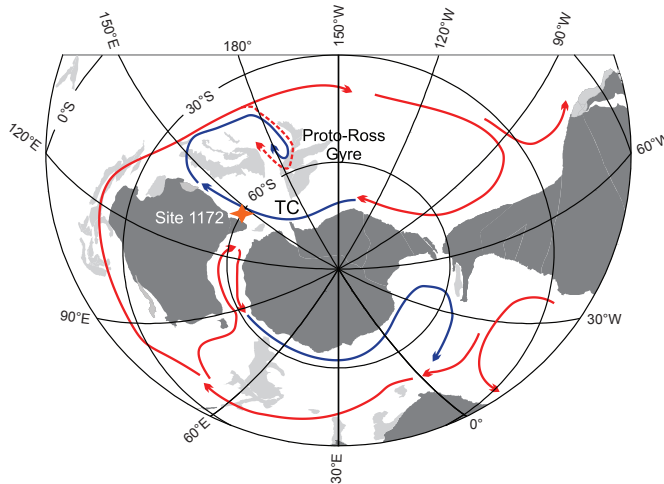
Figure 1. Paleogeographic reconstruction for the South Pacific Ocean at early–middle Eocene times. Surface circulation (Huber et al., 2004) indicates the Antarctic-derived Tasman Current (TC) over the East Tasman Plateau. Paleogeographic charts obtained from the Ocean Drilling Stratigraphic Network (ODSN); after Hay et al. (1999). The dashed red arrow around New Zealand indicates potential mixing of low-latitude surface waters (from the East Australian Current) with the TC.

According to the oldest part of the record, TEX_{86}-derived SSTs at the ETP gradually decreased from ~25°C around 63 Ma to a minimum of ~20°C around 58 Ma (Fig. 2a). During the late Paleocene and early Eocene, Tasmanian SSTs gradually rose to tropical values of ~34°C during the early Eocene climatic optimum (EECO; Zachos et al., 2008), between 53 and 49 Ma (Fig. 2a). A gradual cooling trend throughout the middle Eocene (starting at the termination of the EECO ~49 Ma) arrived at temperatures of ~23°C 42 Ma, which is still relatively warm. Subsequently, an interruption of the cooling trend occurred at the middle Eocene climatic optimum (MECO; ~40 Ma; Bohaty and Zachos 2003), followed by a relatively rapid SST decrease to ~21°C in the early late Eocene (Fig. 2a). The late middle and late Eocene TEX_{86}-based SSTs are supported by $U_{37}^{K'}$ SST estimates derived from the same samples (Supplements; Fig. S3). Both SST estimates also compare well with those for other late Eocene (Southern Ocean) sites (Liu et al., 2009). Unfortunately, sediments from the ETP older than the MECO did not contain alkenones for $U_{37}^{K'}$ SST reconstructions.

The middle Eocene SSTs correspond closely to those from sections in New Zealand (Burgess et al., 2008; Hollis et al., 2009), according to records based on TEX_{86} (Fig. 2a), Mg/Ca and $\delta^{18}O$, indicating regional consistency of our reconstructed SSTs. Also, trends in our Tasmanian SST record are remarkably similar to those in the global stack of benthic foraminiferal oxygen isotopes (Zachos et al., 2008; Fig. 2b), which we updated and augmented with recently published data (Supplements). This correspondence between the two records (Supplements; Fig. S4) indicates that the regional SSTs co-varied with the SSTs where 'global' deep water was sourced. It has previously been suggested that the Southern Ocean was the main region of deep-water formation during the Paleogene (Thomas et al., 2003). In contrast, absolute SST estimates from the Tasmanian region are much higher than those inferred from the benthic foraminiferal oxygen isotopes (Fig. 2). Part of this discrepancy might be due to seasonality, with TEX_{86} being slightly skewed towards summer temperatures and benthic foraminiferal $\delta^{18}O$ towards winter temperatures (Supplements). Another possibility is that deep-water formation occurred in areas that were cooler than the Tasmanian sector.

Sea surface temperature reconstructions based on bivalve-shell oxygen isotopes from Seymour Island on the Antarctic shelf, for example, yield much lower SSTs (Ivany et al., 2008). It is possible that the Antarctic margin was more susceptible to winter cooling than the open ocean, or that portions of the coast along the Antarctic sector were somehow isolated from the southern edges of the Southern Ocean gyres. Another possibility is that the aragonite bivalve shells integrate temperature over a greater portion of the year. Regardless, the large SST difference between the Weddell Sea and the ETP would suggest a relatively steep gradient within a few degrees of latitude. Antarctica, being a polar continent, would most likely have experienced extremes in temperature, in particular having cool winters. Such conditions might have been recorded in the bivalves from the Weddell Sea but not in the more distal ETP. In turn, deep-water formation might have been restricted to the Antarctic shelf areas, such as the Weddell Sea.
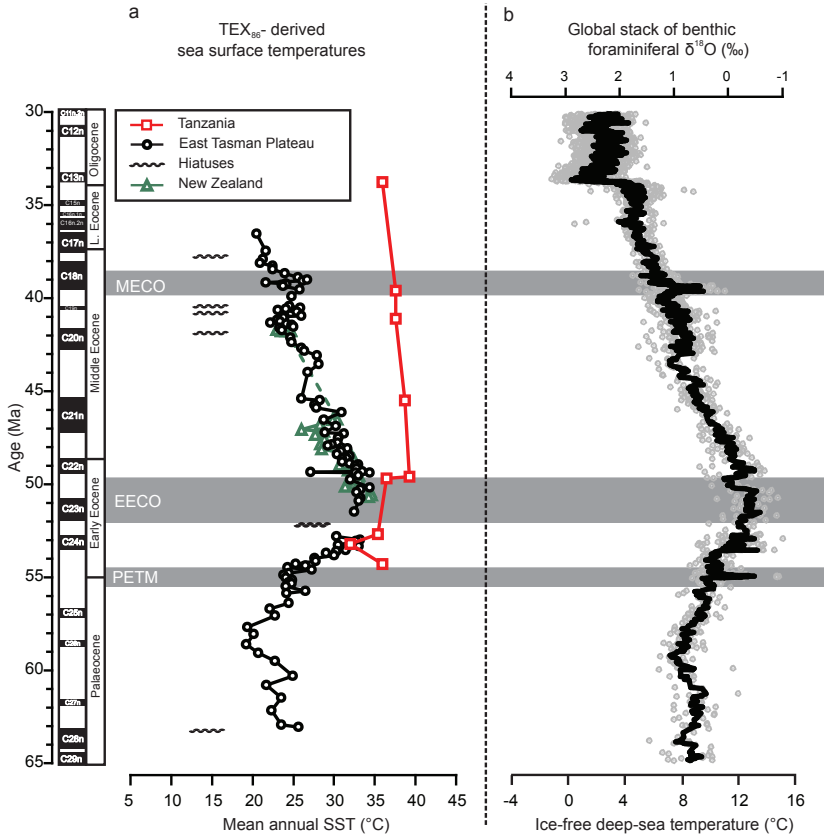
Figure 2. a, TEX$_{86}$ SST reconstructions from ODP Site 1172, New Zealand (Burgess et al., 2008; Hollis et al., 2009) and Tanzania (Pearson et al., 2007) (all according to the same calibration; see Supplements). The black wiggly lines are short (~100 kyr; Röhl et al., 2004a) and longer hiatuses at Site 1172. b, Global stack of benthic foraminiferal oxygen isotopes (grey data; Supplements). The temperature scale assumes ice-free conditions ($\delta^{18}O_{SMOW}$ = -1.2‰, where $\delta^{18}O_{SMOW}$ = ($^{18}O/^{16}O)_{sample}$ / ($^{18}O/^{16}O)_{SMOW}$ - 1; SMOW, standard mean ocean water), and indicates deep-sea temperatures. The black solid line reflects a five-point running average. PETM, Paleocene–Eocene thermal maximum; EECO, Early Eocene Climatic Optimum; MECO, Middle Eocene Climatic Optimum.

Planktonic foraminiferal $\delta^{18}O$ analyses from equatorial regions previously indicated that Paleogene low-latitude SSTs were the same, or even lower, than those of today (Barron 1987), a problem that puzzled paleoclimate scientists for decades. The oxygen isotopic composition of planktonic foraminiferal tests in porous carbonate-rich pelagic facies were later found to be partially altered owing to recrystallization primarily during early diagenesis (Pearson et al., 2001; Schrag et al., 1995). In contrast, carbonate-poor and clay-rich facies typically found on the continental margins contain calcite shells without major diagenetic overprint (Pearson et al., 2007). For the Eocene, such well-preserved planktonic foraminifera indicate near equatorial SSTs that were greater than those of the present day, and

agree with TEX_{86}-derived SSTs (Pearson et al., 2001, 2007). Another observation from well-preserved foraminifera and TEX_{86} is that (sub-) equatorial SSTs were remarkably stable throughout the Eocene (Pearson et al., 2007; Fig. 2). Stable low-latitude SSTs concomitant with high-latitude Eocene cooling thus suggests that there were increasing SST gradients during the Eocene. Although SST trends are often reconstructed using multi-proxy studies, the difference in absolute SSTs between various proxy reconstructions can be considerably large (Hollis et al., 2009; Huber 2008; Liu et al., 2009), even when measured on the same sediments. Despite the fact that multi-proxy approaches are generally encouraged in paleoclimate studies, exclusion of such inter-proxy biases in latitudinal gradient reconstructions requires single-proxy SST records from around the world. Traditional calcite-based SST reconstructions are less suitable for this because calcite is only sparsely available in high-latitude sediments. The organic TEX_{86} and $U^{K'}_{37}$ SST proxies, however, can be used independently of latitude and are, hence, suitable for single-proxy SST gradient reconstructions. Moreover, they do not require critical assumptions about ancient seawater chemistry, unlike $\delta^{18}O$ and Mg/Ca.

We compiled Eocene TEX_{86} and $U^{K'}_{37}$ SST reconstructions from a suite of sedimentary records from localities worldwide and noted increased middle Eocene latitudinal SST gradients in both hemispheres (Fig. 3), relative to the early Eocene. These SST gradients are in general agreement with those found for terrestrial mean annual temperatures, based on early–middle Eocene fossil leaves (Greenwood and Wing 1995). Adding the bivalve-based SST reconstructions from Seymour Island (Ivany et al., 2008) to our organic proxy data suggests a strong gradient between 60° and 70°S, which contrasts with the small gradient between 60°S and the equator (Fig. 3). A part of this large Southern Ocean SST gradient might be due to biases between organic and calcite proxies. A large part, however, may realistically reflect the influence of the cool Antarctic interior, which cooled the Antarctic shelf. In contrast to the continental South Pole, the Arctic region is an oceanic basin. Instead of amplifying the seasonal cycle, the Arctic Ocean probably moderated seasonal extremes in the northern high-latitude greenhouse. Hence, Paleogene latitudinal temperature gradients, like those of today, would have exhibited a high degree of asymmetry between the two hemispheres.

It has been suggested that the general warmth that characterized early Paleogene climates was forced by high atmospheric greenhouse gas concentrations (Zachos et al., 2008). Concomitantly, the absence of polar ice sheets eliminated ice–albedo feedbacks in the Paleogene greenhouse. The middle–late Eocene global cooling has been related to long-term atmospheric CO_2 decline, eventually resulting in the onset of major Antarctic glaciation around the Eocene/Oligocene boundary (Zachos et al., 2008). Our results imply that meridional temperature gradients markedly increased together with deep-sea cooling (Fig. 2; Zachos et al., 2008). Although high-latitudes cooled, tropical temperatures seem to have remained fairly stable throughout the Eocene (Fig. 3; Pearson et al., 2007). This observation raises questions concerning the precise role of decreasing atmospheric greenhouse gas concentrations in cooling the Eocene poles, as in theory (Huber and Sloan 2001) they should have cooled tropical regions as well. The role of potential high-latitude climate feedbacks

involving, for example, differences in cloud/water vapour distribution (Abbot and Tziperman 2008) might have been much more instrumental in the middle Eocene climatic deterioration than previously thought. Another potential positive-feedback mechanism for high-latitude cooling would be ice–albedo feedback. However, the presence of substantial middle Eocene continental ice is still equivocal given the general warmth and overall absence of conclusive physical evidence.
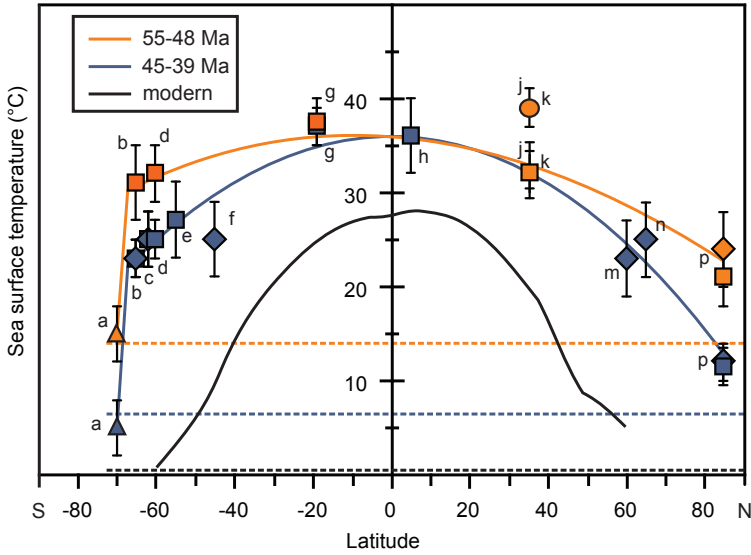
Figure 3. Bivalve-shell $\delta^{18}O$ (triangles), TEX_{86} (squares) and $U^{K'}_{37}$ (diamonds) SST reconstructions for the early (orange) and mid-middle (blue) Eocene. Data are from Seymour Island (Ivany et al., 2008) (a), the East Tasman Plateau (b), Deep Sea Drilling Project (DSDP) Site 277 (Liu et al., 2009) (c), New Zealand (Burgess et al., 2008; Hollis et al., 2009) (d), DSDP Site 511 (Liu et al., 2009) (e), ODP Site 1090 (Liu et al., 2009) (f), Tanzania (Pearson et al., 2007) (g), ODP Site 925 (Liu et al., 2009) (h), New Jersey (Sluijs et al., 2007b) (j, k; circle represents peak PETM SSTs (Sluijs et al., 2007b)), ODP Site 336 (Liu et al., 2009) (m), ODP Site 913 (Liu et al., 2009) (n) and the Arctic Ocean (Sangiorgi et al., 2008; Sluijs et al., 2006; Weller and Stein 2008) (p) (Supplements; Fig. S1). Error bars indicate the range of variation. Gradients represent second-order polynomials, excluding bivalve-shell data. Black and dashed lines represent the present-day zonally averaged latitudinal temperature gradient (Shea et al., 1992) and age-specific deep-sea temperatures, respectively (Fig. 2b; Zachos et al., 2008).

Supplements to Chapter 2

1. Site description, lithology and depositional setting

The East Tasman Plateau (ETP) is located at 45° Southern latitude (Slat.), in the Southwest Pacific. ODP Site 1172 is situated on the flat western side of the (now) submerged ETP, at ~2620 m water depth (Exon et al., 2001). Paleogene sediments were drilled in two holes, 1172A and D, which penetrated ~766 meters of Quaternary to upper Maastrichtian sediments (Stickley et al., 2004a). Paleogene sediments at ODP Site 1172 are olive grey, with some variations to greyish brown and to greenish grey, with abundant bioturbation. Late Paleocene and early Eocene sediments are darker grey, and turn lighter further up section. In the studied section, carbonate content in the sediments is minimal (Exon et al., 2001; Röhl et al., 2004a).

During the early Paleogene, the ETP was at 65°Slat., much closer to Antarctica than it is today (Fig. S1a; Exon et al., 2004). Northward rifting of the ETP (including the Australian continent) started close to the middle- late Eocene boundary (~35.5 Ma) with deepening of the Tasmanian Gateway (Stickley et al., 2004b) and this caused submersion of the ETP. Tectonic movements of the ETP prior to the latest Eocene were confined to slight eastward rifting from the Tasmanian continent (Exon et al., 2004). Prior to the submersion of the ETP, the depositional setting at ODP Site 1172 is interpreted as a shallow marine, tranquil shelf (Exon et al., 2004). During the upper middle Eocene, orbitally forced climatic cycles superimposed on third order sea level changes were identified in ETP sediments (Röhl et al., 2004a). These sea level cycles caused hiatuses in the record (Röhl et al., 2004a), correlated to sea level low stands. It is unknown whether these sea level cycles also influenced the ETP paleo- environments prior to 43 Ma. However, the complete magnetic profile (Stickley et al., 2004a) indicates that hiatuses are relatively small (~100kyr). Microfossil analyses in combination with surface-ocean General Circulation Models (GCMs) indicate that the ETP is under influence of northward flowing surface currents (Huber et al., 2004), bringing Antarctic- derived surface waters over the ETP (Fig. S1b). Therefore, the recorded climate signal at ODP Site 1172 is at least in part Antarctic-derived.

2. Revised age model of ODP Leg 189 Site 1172

For the age model, tie points are based solely on magnetic reversals and PETM isotope events (Bijl et al., 2007). The depths of the magnetic reversals were obtained from Fuller and Touchard (2004) and Stickley et al. (2004a). High-resolution organic walled dinoflagellate cyst (dinocyst) chronology comparison between ODP Site 1172 and Site 1171 resulted in the identification of a hiatus spanning from mid-Subchron C24n to mid- Subchron C23n, which resulted in a revision of the published magnetostratigraphy for ODP Site 1172 (Fuller and Touchard 2004). The PETM was previously interpreted to be missing from the record (Röhl et al., 2004b), however we identified the PETM slightly further up section (Bijl et al., 2007; Chapter 4). The identification of the PETM led to a revised interpretation of the paleomagnetic data of Fuller and Touchard (2004). All these re-interpretations resulted in an adapted age

model for ODP Site 1172 (Table S1 and Fig. S2). To line the deep sea oxygen isotope data up with the data of Site 1172, the onset of the PETM in the deep sea oxygen isotope data was set at 55 Ma, although an older age of the PETM had recently been suggested (Westerhold et al., 2007). Ages for the magnetic reversals were derived from the Gradstein et al. timescale (2004), except for the Subchrons C24n and C25n, which were from Westerhold et al. (2007).

a

b
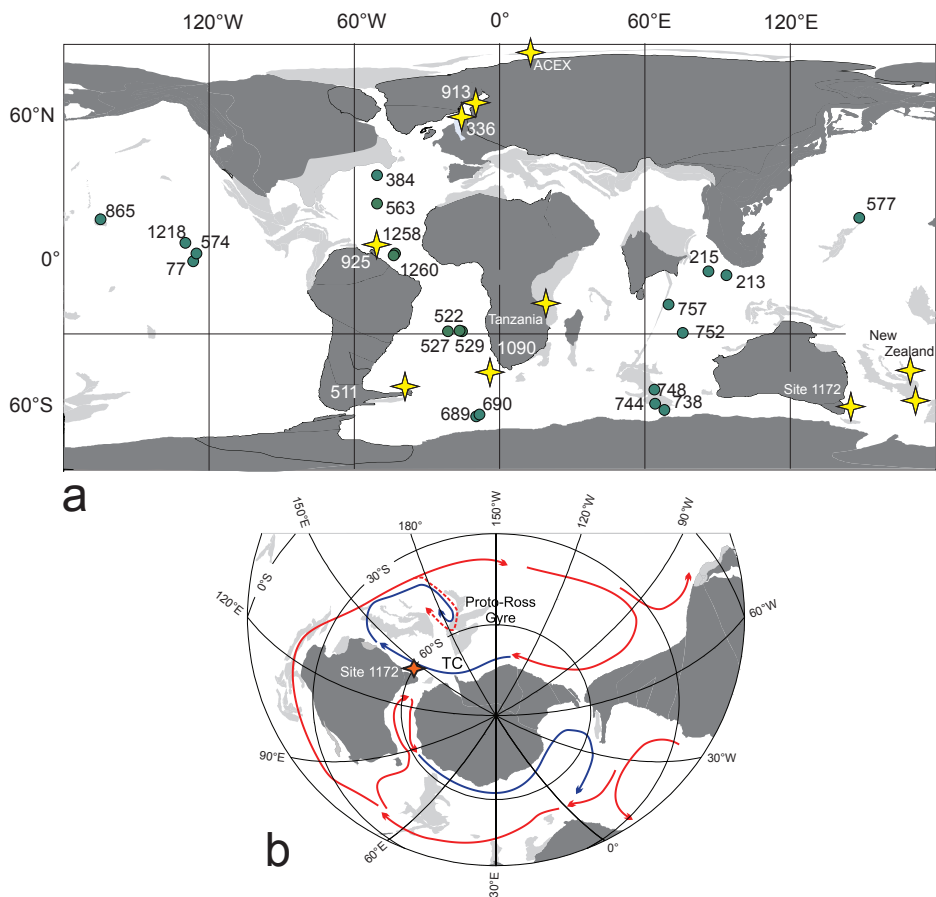
Figure S1. a): Site locations of the records used in this study. The stars represent those sites with TEX$_{86}$ and/or U$_{37}^{K'}$ SST reconstructions, circles represent records in the herein updated global stack of benthic foraminiferal oxygen isotopes. The numbers in the graph refer to the site codes as assigned by the different ocean drilling programs DSDP, ODP and IODP. The paleogeographic reconstruction represents early- middle Eocene times (~50 Ma). Paleogeographic chart was obtained from the Ocean Drilling Stratigraphic Network, after Hay et al. (1999). b): South Pacific paleogeographic reconstruction, with GCM- derived ocean current model. Paleogeographic chart was obtained from the Ocean Drilling Stratigraphic Network, after Hay et al. (1999). Reconstructed ocean currents from Huber et al. (2004). TC: Tasman Current.
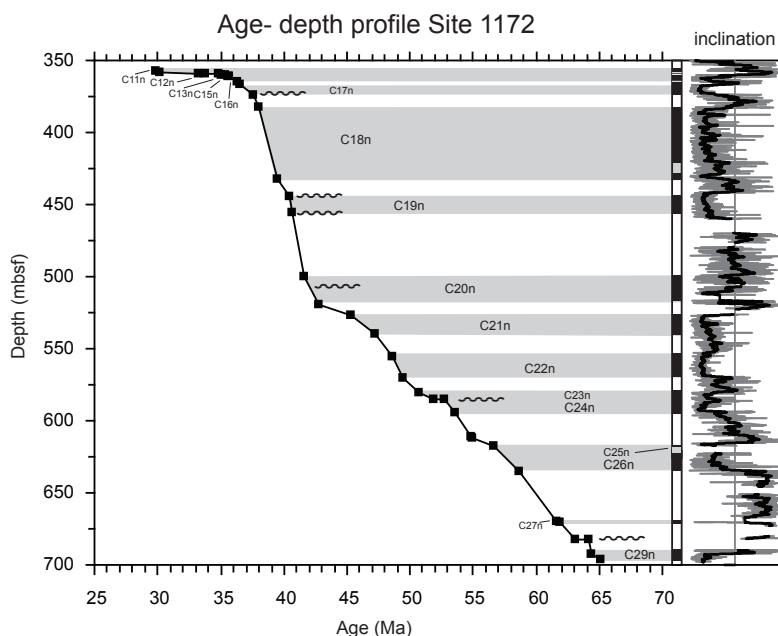
Figure S2. Age- depth profile for ODP Site 1172 with the revised age model. Also plotted are the inclination data, obtained from the Ocean Drilling Program database (www-odp.tamu.edu). The black line in the inclination curve represents a 25-point running mean through the inclination data. Note the hiatus in the early Paleocene (Schellenberg et al., 2004), the increase in sedimentation rates after the PETM, the small hiatus in the late early Eocene and the drop in sedimentation rates during the late Eocene (Stickley et al., 2004a, b; Sluijs et al., 2003). For more information on the late Eocene to early Oligocene stratigraphy see Sluijs et al. (2003).

3. Methods: TEX$_{86}$ and U$_{37}^{K'}$

Organic compounds were extracted from powdered and freeze- dried sediments with dichloromethane (DCM)/ methanol (MeOH) (9:1) by using accelerated solvent extraction technique (Dionex). Excess solvent was removed using rotary evaporation with vacuum. The total extracts were separated in polar and apolar fractions over an activated Al$_2$O$_3$ column using Hexane:DCM (1:1) and DCM:MeOH (1:1), respectively. The polar fraction was then dissolved in a 99:1 hexane/propanol solvent, and filtered using a 0.45 μm, 4 mm diameter polytetrafluoroethylene (PTFE) filter, before being analyzed using a high performance liquid chromatography/ atmospheric pressure positive ion chemical ionization mass spectrometry (HPLC/ APCI-MS). HPLC/APCIMS analyses were done according to Schouten et al. (2007a) using an Agilent 1100 series LC/MSD SL and separation and a Prevail Cyano column (2.1 x 150 mm, 36mm; Alltech), maintained at 30°C. The glycerol dialkyl glycerol tetraethers (GDGTs) were eluted using a changing mixture of hexane and propanol as follows, 99% hexane: 1% propanol for 5 minutes, then alinear gradient to 1.8% propanol in 45 minutes. Flow rate was 0.2 ml per minute. Single ion monitoring was set to scan the 5 [M+H]+ ions of the GDGTs with a dwell time of 237 ms for each ion.

We applied the core top calibration equation from Kim et al. (2008) to translate TEX_{86} values to an estimation of mean annual sea surface temperature:

$$T = -10.78 + 56.2 \times TEX_{86} \tag{2.1}$$

About half of the analyzed samples were run in duplicate, to estimate the reproducibility. In all these samples, the reproducibility is better than 0.5°C, and in most cases better than 0.25°C. Recently, Liu et al. (2009) used in their calibration the same core top data set as Kim et al. (2008), but included data that Kim et al. (2008) discarded. However, Kim et al. discarded core top data from cold (Arctic and Antarctic) or hypersaline (Red Sea) regions for a reason: in those special settings, crenarchaeota appear to behave differently to SST (see e.g., Turich et al., 2007). Kim et al. (2008) postulated that the near invariance of high-latitudinal TEX_{86} index values to SST was due to changed membrane adaptation of low temperature crenarchaeotal communities. The good correlation in trends between our TEX_{86} SSTs and benthic foraminiferal $\delta^{18}O$ suggests that TEX_{86} confidently records ancient SST trends and the correlation between polar SSTs and TEX_{86} index values in ancient greenhouse climates seems not to be invariant like at present day conditions (Fig. S3).

Therefore, present day TEX_{86} issues at polar regions appear to be specific for the present day polar climates, and did not affect the relation between TEX_{86} and SST in ancient polar greenhouse climates. In fact, the high-latitude SSTs inferred in our study (Fig. 2) are obviously well above modern day high-latitude SST range. Therefore, addition of polar SST data points to the core top calibration (as was done by Liu et al. (2009)) is not valid for our purpose, but only skews the calibration towards conditions that were not even present in the Paleogene greenhouse. Furthermore, incubation studies show that at the high temperature end (i.e., above modern day temperature range) there is a continued linear relation between TEX_{86} and SST at least until 40°C (Schouten et al., 2007b). Hence there is a prior no reason to assume a decreased sensitivity of TEX_{86} to SST at above modern day temperatures. Thus, while Liu et al. (2009) may have provided a good TEX_{86} calibration for icehouse conditions with low polar temperatures, in ancient greenhouse climates we prefer the linear calibration equation by Kim et al. (2008) instead of the reciprocal calibration of Liu et al. (2009).

To analyze the alkenones, the apolar fractions were separated in 3 fractions by Al_2O_3 column chromatography using Hexane:DCM (9:1, v/v); Hexane:DCM (1:1, v/v; the ketone fraction with the alkenones) and DCM. The ketone fractions were analyzed for alkenones by gas chromatography (GC) using an Agilent 6890 equipped with an on column injector. Separation was achieved using a 50m CPSil 5CB column (0.32 mm internal diameter, 0.12 micrometer film thickness). The values calculated from the C37:2 and C37:3 alkenones were converted into temperature values by applying the core-top calibration of Müller et al. (1998):

$$T = (U_{37}^{K\prime} - 0.033)/0.044 \tag{2.2}$$

The reproducibility of the analysis is better than 0.5°C. Selected fractions were analyzed by GC/mass spectrometry (GC/MS) using a TRACE DSQ to confirm the presence of the alkenones and the absence of co-eluting compounds. Alkenones could only be detected in sufficient quantities in late Eocene sediments.

Type	Event	Reversal	Depth (mbsf)	Age (Ma)
m	T	C11n.2n	357.300	29.853
m	O	C11n.2n	358.600	30.217
m	T	C13n	358.800	33.270
m	O	C13n	358.900	33.740
m	T	C15n	359.700	34.780
m	O	C15n	359.900	35.040
m	T	C16n.1n	360.100	35.400
m	O	C16n.1n	361.300	35.570
m	T	C16n.2n	361.500	35.710
m	O	C16n.2n	365.000	36.280
m	T	C17n	367.100	36.510
m	O	C17n	373.900	37.550
m	T	C18n.1n	382.100	38.032
m	O	C18n.2n	432.000	39.460
m	T	C19n	444.000	40.440
m	O	C19n	456.000	40.670
m	T	C20n	500.000	41.590
m	O	C20n	519.000	42.770
m	T	C21n	526.400	45.350
m	O	C21n	540.000	47.240
m	T	C22n	555.900	48.600
m	O	C22n	570.000	49.430
m	T	C23n	581.000	50.730
m	O	C23n	585.000	51.900
m	T	C24n	585.010	52.730
m	O	C24n	594.200	53.530
i	T	PETM	610.800	54.850
i	O	PETM	611.800	55.000
m	T	C25n	618.000	56.650
m	O	C26n	634.800	58.700
m	T	C27n	670.000	61.650
m	O	C27n	670.750	61.983
m	T	C28n	685.000	63.104
m	O	C28n	685.010	64.128
m	T	C29n	692.600	64.432
m	O	C29n	695.990	65.118
m	T	C30n	695.991	65.861
m	O	C30n	709.100	67.696
m	T	C31n	713.300	67.809
m	O	C31n	736.600	68.732

Table S1. Ages and depths for the magnetic reversals and the PETM as used in the age model. m: magnetic reversal, i: isotope event, O: onset, T: termination.

4. The updated global stack of benthic foraminiferal oxygen isotopes

The previously published global stack of benthic foraminiferal oxygen isotopes (Zachos et al., 2008) (see Fig. S1 for site locations) had been modified for the purpose of this paper. First of all, the entire dataset had been converted to the timescale of Gradstein et al. (2004). Subsequently, data from Site 748B and 1218 had been adjusted to the latest age model for the MECO (Bohaty et al., 2009). Isotope data from Demerara Rise (Site 1258 and 1260; Sexton et al., 2006) were added, after adjustment to the Gradstein timescale (Gradstein et al., 2004). The MECO data of Site 1260 were previously plotted on a predated MECO time scale, and were adjusted according to Bohaty et al. (2009). Figure 2 shows the updated global stack of benthic foraminiferal oxygen isotopes, with a 5 point running mean through the entire dataset. Especially the Paleocene part of the stack is notoriously scattered.
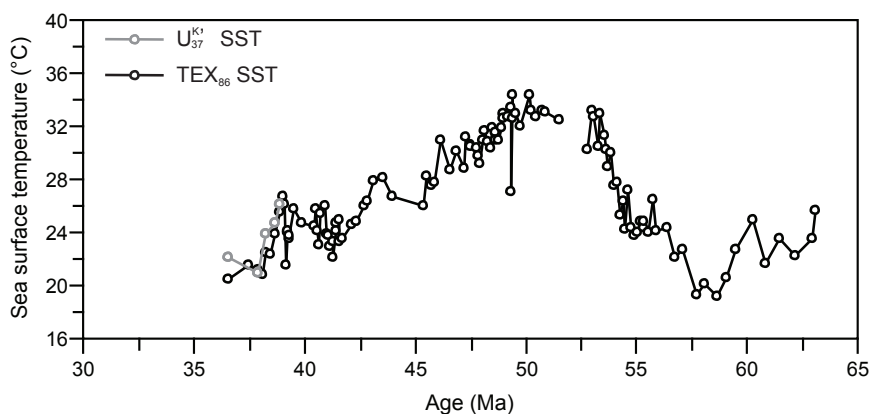
Figure S3. The Tasmanian TEX_{86} (black) and $U^{K'}_{37}$ (red) SST record for the early Paleogene.

5. Correlation of benthic foraminiferal $\delta^{18}O$ and Tasmanian SSTs

In order to make a more detailed comparison between Tasmanian SSTs and the global stack of benthic foraminiferal oxygen isotopes, we re-sampled the global stack. Using Analyseries software (Paillard et al., 1996), we interpolated the oxygen isotope data for each data point of TEX_{86}, in order to obtain contemporaneous Tasmanian SST and 'global' benthic foraminiferal oxygen isotope values. Correlation between ETP SSTs and the resampled updated benthic $\delta^{18}O$ stack is good for the entire interval between 64 and 36 Ma (R^2=0.64; Fig. S4). R^2 values are even higher if Paleocene and early Eocene values are separated from middle and late Eocene values (R^2 =0.76 and 0.74, respectively). In these two time intervals, SSTs correlate with a different slope to benthic $\delta^{18}O$. This could imply regional SST warming in the Southwest Pacific. Alternatively, it could indicate an increase in the benthic $\delta^{18}O$ stack values through an increase in the basin-to-basin $\delta^{18}O$ gradient (Fig. S3), changes in deep water source area, or the build up of continental ice. The high R^2 values indicate that the Southwest Pacific SST trend mirrored that of the deep ocean, and suggests

that the locus of deep-water formation may have been in the Southwest Pacific. While overall trends in both records match closely, absolute deep-sea temperatures are much cooler than southwest Pacific SSTs throughout the Paleogene. One explanation for the large discrepancy in absolute temperatures in both records might be seasonality.
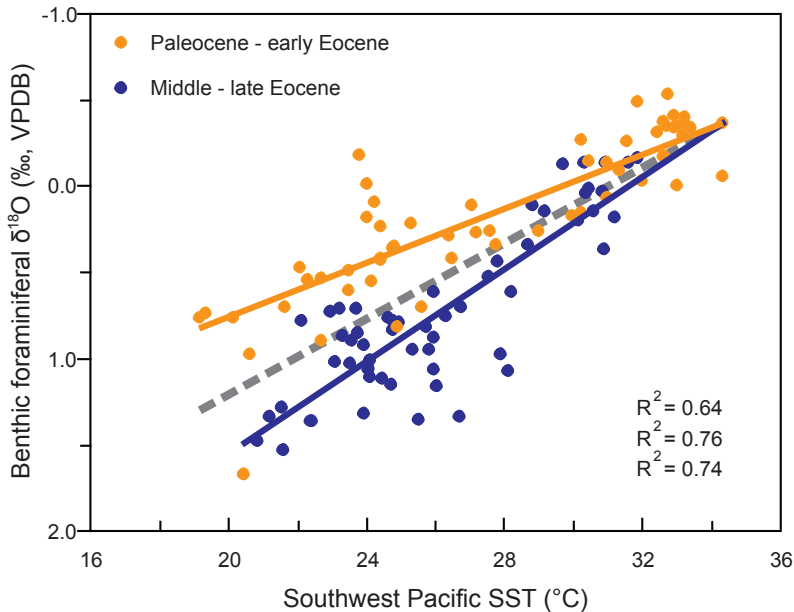
Figure S4. Correlation between Tasmanian SSTs (x-axis) and the re-sampled updated global stack of benthic foraminiferal oxygen isotopes (y-axis) is good (grey line; $R^2=0.64$). Conspicuously, the correlation coefficient for middle Eocene- late Eocene data is different from that for the Paleocene- early Eocene.

While generally the TEX$_{86}$ SST proxy reflects mean annual SST (Kim et al., 2008; Schouten et al., 2002) high-latitude TEX$_{86}$ SST reconstructions might be skewed towards summer, because that is the main period of organic carbon export and thus of fluxes of crenarchaeotal membrane lipids to the sea floor (Sluijs et al., 2006, 2008a). Although ODP Site 1172 was not within the polar circle it may be that the main season of lipid export at ~65°Slat. was during summer. In turn, deep-water formation at high-latitudes mainly occurs during winter, when surface waters reach maximum densities, thereby skewing benthic δ^{18}O-based temperature estimates towards winter SSTs of the source regions.

CHAPTER 3

Environmental forcings of Paleogene Southern Ocean dinoflagellate biogeography

Arachnodinium antarcticum Wilson and Clowes, 1982
Cyst has a smooth ovoidal central body and faintly striate, locally perforate processes, finely denticulate on the upper margin. Antapical process is hollow, typically broad and somewhat rectangular in cross section. Paratabulation indicated by process distribution and pattern of trabecular network; parasutural features lacking.

Despite warm polar climates and low meridional temperature gradients, a number of different high-latitude plankton assemblages were, to varying extents, dominated by endemic species during most of the Paleogene. To better understand the evolution of Paleogene plankton endemism in the high southern latitudes, we investigate the spatio-temporal distribution of the fossil remains of dinoflagellates, i.e., organic-walled cysts (dinocysts), and their response to changes in regional sea surface temperature (SST). We show that Paleocene and early Eocene (~65–50 Ma) Southern Ocean dinocyst assemblages were largely cosmopolitan in nature but that a distinct switch from cosmopolitan-dominated to endemic-dominated assemblages (the so-called "transantarctic flora") occurred around the early-middle Eocene boundary (~50 Ma). The spatial distribution and relative abundance patterns of this transantarctic flora correspond well with surface water circulation patterns as reconstructed through general circulation model experiments throughout the Eocene. We quantitatively compare dinocyst assemblages with previously published TEX_{86}– based SST reconstructions through the early and middle Eocene from a key locality in the southwest Pacific Ocean, ODP Leg 189 Site 1172 on the East Tasman Plateau. We conclude that the middle Eocene onset of the proliferation of the transantarctic flora is not linearly correlated with regional SST records and that only after the transantarctic flora became fully established later in the middle Eocene, possibly triggered by large-scale changes in surface-ocean nutrient availability, were abundances of endemic dinocysts modulated by regional SST variations.
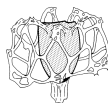
1. Introduction

Early Paleogene temperatures and atmospheric CO_2 concentrations were significantly higher than today (Chapter 2; Pearson and Palmer 2000; Sexton et al., 2006; Sluijs et al., 2006; Zachos et al., 2008). This long-term global warm period peaked during the late early Eocene (~52– 50 Ma; Sexton et al., 2006; Zachos et al., 2001), with high (>30°C) southern high-latitude SSTs and minimal latitudinal temperature gradients (Chapter 2). The subsequent long-term global cooling through the middle and late Eocene (Chapter 2; Sexton et al., 2006; Zachos et al., 2008) culminated in the establishment of large Antarctic ice sheets around the time of the Eocene-Oligocene boundary (~34 Ma; Barker et al., 2007; Coxall et al., 2005; DeConto and Pollard 2003a, b; Miller et al., 1987).
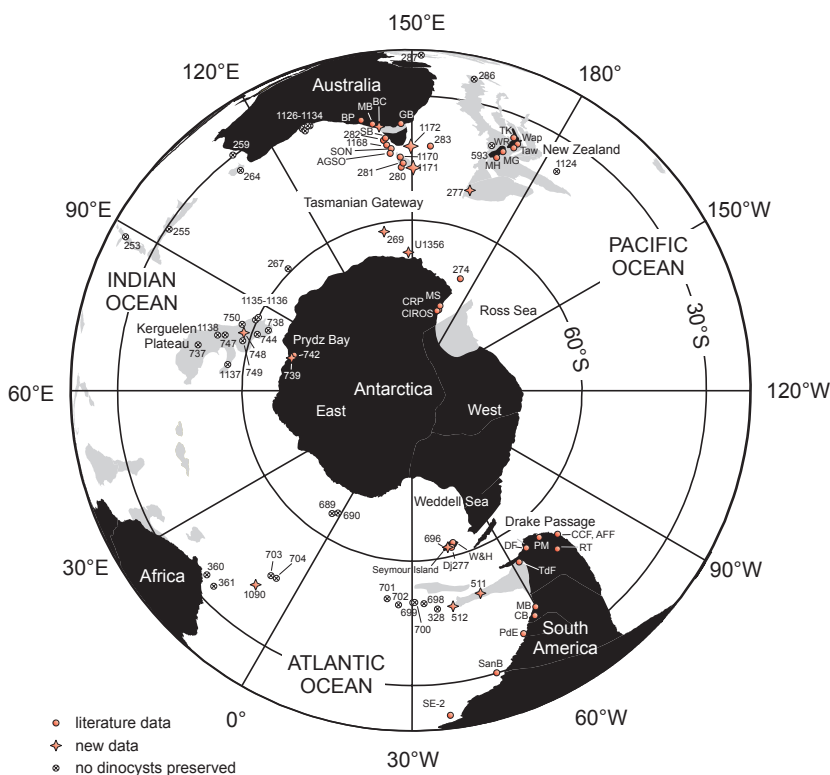
Figure 1. Location map for present-day positions of evaluated sites. Numbers 253–593 refer to Deep Sea Drilling Project (DSDP) sites; numbers 689–1172 refer to Ocean Drilling Program (ODP) sites. IODP Site U1356 refers to a site drilled by the Integrated Ocean Drilling Program. For code legend, see Table S1. The geographic map was derived from the Ocean Drilling Stratigraphic Network (ODSN).
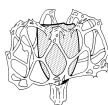
The warm climates and low latitudinal temperature gradients prior to the onset of major Antarctic glaciation facilitated the poleward migration of many biota (Adams et al., 1990; Eberle et al., 2009; Sluijs et al., 2006). Surprisingly, however, Southern Ocean fossil assemblages are distinctly endemic during long intervals of the early Paleogene and become increasingly endemic over time; besides molluscs (Zinsmeister, 1979), this phenomenon has been observed in numerous microfossil groups such as dinocysts (Wrenn and Beckman, 1982; Wrenn and Hart 1988) radiolaria (Lazarus and Caulet, 1993; Lazarus et al., 2008), calcareous nannoplankton and planktonic foraminifera (Nelson and Cooke, 2001; Villa et al., 2008), and diatoms (Harwood, 1991; Stickley et al., 2004b). Dinocysts are abundant in marginal marine sediments, notably at high-latitudes. The early Paleogene Antarctic endemic dinocyst community is taxonomically well documented from outcrops and sediment cores around the Antarctic margin, and it is often referred to as the "transantarctic flora" (cf. Wrenn and Beckman, 1982). However, the evolution and driving forces behind this Paleogene Southern Ocean endemism are largely unknown, although SST is presumed to have been involved (Huber et al., 2004). So far, a lack of reliable SST reconstructions has hampered the evaluation of this hypothesis, and it is only through recently published geochemical data (Chapter 2) that such a test has become possible.

In this study, we aim to elucidate the role of oceanic surface circulation, SST and other environmental changes as a potential selective mechanism for Eocene Antarctic endemism among dinoflagellates. First, we provide an overview of Paleogene dinocyst paleobiogeography, for which we integrate newly generated palynological data with those previously published. Additionally, we statistically examine the relation of SST (using the TEX_{86} paleothermometry of Chapter 2) and ocean-wide productivity (using the thallium isotope records of Nielsen et al., 2009) to dinocyst assemblage changes in the southwest Pacific Ocean (Ocean Drilling Program Site 1172).

2. Material and Methods

2.1. Southern Ocean sites

We generated palynological data on sediments recovered at Ocean Drilling Program (ODP) and Deep Sea Drilling Project (DSDP) Sites 269, 277, 511, 512, 696, 739, 748 and 1090, as well as from the Brown's Creek section in southeast Australia (Fig. 1). Higher-resolution time series data were obtained from ODP Sites 1172 and 1171, and from outcrop samples from Blanche Point along the Southern Australian Margin. Furthermore, we reviewed published information on Paleogene (~65–25 Ma) Southern Ocean dinocysts from Integrated Ocean Drilling Program (IODP), DSDP and ODP sites, the "Cape Roberts Project" (CRP 1, 2a and 3) and "Cenozoic Investigations of the Ross Sea 1" (CIROS-1) drilling expeditions. In addition, studies of outcrops that flank the Southern Ocean were considered from the southern Australian margin, McMurdo Sound, New Zealand, Tierra del Fuego, Santa Cruz Province, Brunswick Peninsula, and Seymour Island. An inventory of all sites and outcrops considered (including references) is listed in Table S1 at the end of this chapter. An overview on the present-day locations of these sites is provided in Fig. 1.
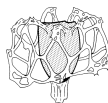
The age control for these records is based on calcareous nannoplankton stratigraphy and, to a lesser extent, on diatom and/or palynomorph stratigraphies, radiometric dating, and magnetostratigraphy. The approximate tectonic configurations of the continents used in Figs. 2 and 3 were reconstructed using the Ocean Drilling Stratigraphic Network (ODSN) database: http://www.odsn.de (after Hay et al. (1999)). The semi-continuous succession recovered at ODP Site 1172 has been magnetostratigraphic calibrated (Stickley et al., 2004a) and extends from the Quaternary to the Maastrichtian (latest Cretaceous). Notably, it represents a reference section for Paleogene Antarctic endemism. ODP Site 1172 is located on the submerged western side of the East Tasman Plateau (ETP) at 45°S and a water depth of 2,620 m (Fig. 1; see Exon et al. (2004) for details). During the early Paleogene, the ETP was located at 65°S, much closer to Antarctica than it is today (Exon et al., 2004). Recovery at Hole 1172A includes Paleogene sediments down to the mid- middle Eocene, while Hole 1172D extends the record from the mid- middle Eocene to the Maastrichtian. Calcite concentrations increase from ~40 Ma onward, which has been linked to a relative sea level rise following basin subsidence (Röhl et al., 2004a). Paleocene and Eocene sediments were recovered from between 350 and 700 meters below seafloor (mbsf; Exon et al., 2003; Stickley et al., 2004a). This part of the section consists of brownish to greenish grey, bioturbated siltstones, that are poor in calcium carbonate, but yield abundant organic matter, including dinocysts (Exon et al., 2004).

2.2. Southern Ocean surface-current configurations

According to modeling experiments and biogeographical data, in the early Paleogene the region was under the influence of the Antarctic-derived, northward flowing Tasman Current (Huber et al., 2004). This current was part of a large gyre system in the South Pacific Ocean, the Proto-Ross Gyre (Stickley et al., 2004b). The tectonic positioning of the continents that prescribe this current system was stable at least until ~35.5 Ma, when the northward movement of Australia from Antarctica deepened the Tasmanian Gateway and much later allowed for the Antarctic Circumpolar Current to establish (Stickley et al., 2004b).

2.3. Palynological processing and taxonomy

Processing sediment samples for palynology followed the standardized methods described, e.g., by Sluijs et al. (2003). Briefly, this involves treatment of dried sediment samples with ~30% HCl and ~38% HF. No bleaching and heavy-liquid separation was applied. A 15 micrometer nylon mesh was used for sieving. Residues were mounted on slides for microscope analysis. Approximately 200 dinocysts were counted per sample and identified to the species level at 500× magnification. An inventory of dinocyst species encountered in this study is listed in Table S2 at the end of this chapter; dinocyst nomenclature and taxonomy, unless stated otherwise, are based on the work of Fensome and Williams (2004).

2.4. Dinocysts: endemic versus nonendemic taxa

Dinocysts are the remains of unicellular eukaryotic plankton (dinoflagellates) that dwell as motile stages in surface oceans (e.g., Fensome et al., 1993). Dinoflagellates occur in virtually all marine settings, but are most abundant and diverse in shelfal environments (e.g., Dale 1996; Pross and Schmiedl 2002; Pross
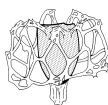
and Brinkhuis 2005; Sluijs et al., 2005). The fossil record of dinoflagellates is predominantly based on their resistant organic-walled cysts (dinocysts). The dinocyst record, in turn, is particularly good in shelfal environments where relatively high sedimentation rates lead to a reduced oxygen exposure of the cysts. In contrast, deeper environments with well-ventilated bottom waters exhibit generally lower sedimentation rates and high oxygen concentrations, and in these settings dinocysts oxidize and do not preserve.

Species described by…	Now assigned to transantarctic flora
…prior to Wrenn and Hart (1988)	*Alterbidinium distinctum*
	Arachnodinium antarcticum
	Deflandrea antarctica
	Deflandrea cygniformis
	Deflandrea granulata
	Deflandrea oebisfeldensis
	Spinidinium macmurdoense
	Vozzhennikovia apertura
…Wrenn and Hart (1988)	*Spinidinium colemanii*
	Octodinium askiniae
	Hystrichosphaeridium truswelliae: now non-endemic (Sluijs and Brinkhuis, 2009)
	Impletosphaeridium clavus
	Deflandrea webbii
	Spinidinium luciae
…Wilson (1988)	*Membranophoridium perforatum*
…Goodman and Ford (1983)	*Phthanoperidinium* sp. A
	Phthanoperidinium sp. B
…Stover and Williams (1987)	*Gippslandia extensa*
…Mao and Mohr (1995)	*Phthanoperidinium antarcticum*
…Stover and Williams (1995)	(now) *Enneadocysta dictyostila*
…Levy and Harwood (2000)	*Vozzhennikovia netrona*
…Brinkhuis et al. (2003a)	*Deflandrea* sp. A
…Clowes and Wilson (2006)	*Corrudinium regulare*
…Sluijs et al. (2009b)	*Vozzhennikovia stickleyae*
	Vozzhennikovia roehliae
	Spinidinium schellenbergii
	Moria zachosii
Bipolar species assigned by:	**Bipolar species**
Warnaar (2006)	*Phthanoperidinium stockmansii*
Wrenn and Hart (1988)	*Spinidinium macmurdoense*

Table 1. Overview of dinocysts herein assigned to the transantarctic flora and the bipolar species.
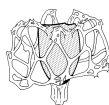
Dinocyst species are considered to be "endemic to the Southern Ocean" if their empirically established distribution patterns are restricted to this region (i.e., latitudes south of 45°S). The Eocene endemic dinocysts in the Southern Ocean were originally referred to as the "transantarctic flora," a name derived from their presumed affinity to the inferred Trans-Antarctic Seaway (Wrenn and Beckman 1982). Later studies reported elements of the transantarctic flora from other Southern Ocean sectors (Brinkhuis et al., 2003a; Clowes and Wilson 2006; Levy and Harwood 2000; Macphail and Truswell 2004; Mao and Mohr 1995; Sluijs et al., 2009b; Wrenn and Hart 1988). Some of these reports also described new species, but did not place those within the transantarctic flora. Other species occur in high abundances only in both north and south polar regions, and may therefore be placed in a "bipolar" group (e.g., Wrenn and Hart 1988). For this study, we include the bipolar species in our endemic dinocysts category herein after. Moreover, some taxa that are allegedly endemic to the Southern Ocean are morphologically similar to some cosmopolitan forms. For example, the supposedly endemic species *Dracodinium waipawaense* (Wilson 1967b) and the Northern Hemisphere *D. varielongitudum* (Williams and Downie 1966) have the same stratigraphic range and are taxonomically identical, if one allows for subtle morphological variations in shape, size and number of processes (spines) on the cysts. Hence, it can be debated whether species such as *D. waipawaense* are truly endemic or whether the slight morphological differences are the result of eco-phenotypic variability within a single dinocyst species. Because of these uncertainties, we did not consider such species to be part of the transantarctic flora. On the other hand, the Eocene species *Enneadocysta dictyostila* (Menéndez 1965; Stover and Williams 1995, emend. Fensome et al., 2006) can be classified as endemic, despite its resemblance to the Northern Hemisphere species *Areosphaeridium diktyoplokum* (Klumpp 1953). The two species can be clearly separated based on differing (para-) tabulation, a fundamental structural difference that is highly unlikely to occur within an individual species or genus (Goodman and Ford 1983); see discussions by Fensome et al. (2006). An overview of the dinocyst species we assign herein to the transantarctic flora is provided in Table 1.

2.5. Canonical correspondence analyses

We assessed the potential correspondence of the dinocyst data from ODP Site 1172 with four environmental variables reconstructed from the same site as well as other records using Canonical Correspondence Analyses. With this statistical method, quadratic relations between the first and higher axes have been found to cause arch effects in the ordination (ter Braak 1986). To avoid these effects, we applied detrended canonical correspondence analyses (DCCA; see ter Braak 1986). We performed two DCCA analyses, one on the entire investigated record and the second only using middle Eocene data. The TEX_{86}-based SST record of ODP Site 1172 from (Chapter 2) was added as the first environmental variable. This SST record comprises ~120 samples for the Paleocene-Eocene (average sample resolution: 250 kyr). As second variable, we used the thallium isotope record from the Pacific Ocean (Nielsen et al., 2009). We averaged two coeval records of thallium isotopes and re-sampled the data to the (lower-resolution) TEX_{86} SST data set. The thallium isotope record is interpreted to reflect ocean-wide Pacific Ocean productivity trends (Nielsen et al., 2009). Third, the peridinioid/ gonyaulacoid dinocyst ratio may serve as a measure for
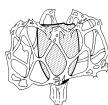
the trophic state of the surface waters (Sluijs et al., 2005). Finally, we added the percentage of endemic dinocysts. We used only those samples that were analyzed for both TEX_{86} and palynology, rather than interpolating the TEX_{86} data. The DCCA analysis on the entire data set comprises 59 samples. We performed a separate DCCA analysis for the middle Eocene using the same environmental variables, for which the number of samples is 21.

3. Results

3.1. Dinocyst biogeography through time

Distribution maps of the compiled circum-Antarctic dinocyst records for seven time slices spanning ~56 Ma to 36 Ma are presented in Figs. 2a–2g. The distribution patterns for the two oldest time slices (Fig. 2a, 56–54 Ma; Fig. 2b, 53–51 Ma) have limited numbers of dinocyst-bearing sites restricted to the southwest Pacific Ocean and Drake Passage. This limited information shows that endemics are present, but do not numerically dominate the dinocyst assemblages, not even on the Antarctic Margin. During the early middle Eocene, endemic taxa are dominant in the areas east of the Tasman Gateway and Drake Passage (Fig. 2c, 50–48 Ma). From the middle Eocene onward, endemic dinoflagellates consistently and strongly dominate the assemblages along the western boundaries of the Drake Passage and the Tasman Gateway as well as at Prydz Bay (Figs. 2d, 2e, and 2f; 47 Ma to 39 Ma), at least until the mid to late Eocene (Fig. 2g; 38–36 Ma). Many undoubtedly Oligocene and younger records in the south- west Pacific (not presented here; e.g., ODP Leg 28 (Kemp 1975)) and Atlantic (Guerstein et al., 2008) yield some endemic dinocysts. These are most likely reworked from Eocene sediments as a result of Oligocene glacial activity. We have compiled a composite of the generalized distribution of endemic dinocysts during the middle and late Eocene (47–36 Ma) and note that the sectors characterized by dominant endemism are restricted to the Antarctic margins and the eastern sides of the Tasman and Drake conduits (Fig. 3). The Australo-Antarctic Gulf area is characterized by overall low abundances of endemics throughout the Paleogene (Fig. 3).

Figure 2 (next pages). Circum-Antarctic geographical distribution maps showing the spatio-temporal evolution of dinocyst endemism for seven different time slices. Color shading of Sites yielding dinocysts indicates relative abundance of transantarctic flora, from red (0%) to blue (100%). (a) Thanetian-Ypresian boundary (56–54 Ma), (b) mid-Ypresian (53–51 Ma), (c) Ypresian- Lutetian boundary (50–48 Ma), (d) early Lutetian (47–45 Ma), (e) late Lutetian (44–42 Ma), (f) Lutetian-Bartonian boundary (41–39 Ma), and (g) Bartonian- Priabonian boundary (38–36 Ma). Maps were derived from the Ocean Drilling Stratigraphic Network (ODSN). Black areas indicate (continental) blocks that are mostly subaerial. Note that several blocks shown in black were partly submerged during the Paleogene (e.g., Ross Sea, southern Australian margin, and parts of Argentina). Shaded areas indicate mostly submerged continental blocks (e.g., Brown et al., 2006). For locality codes, see Table S1.
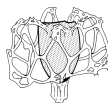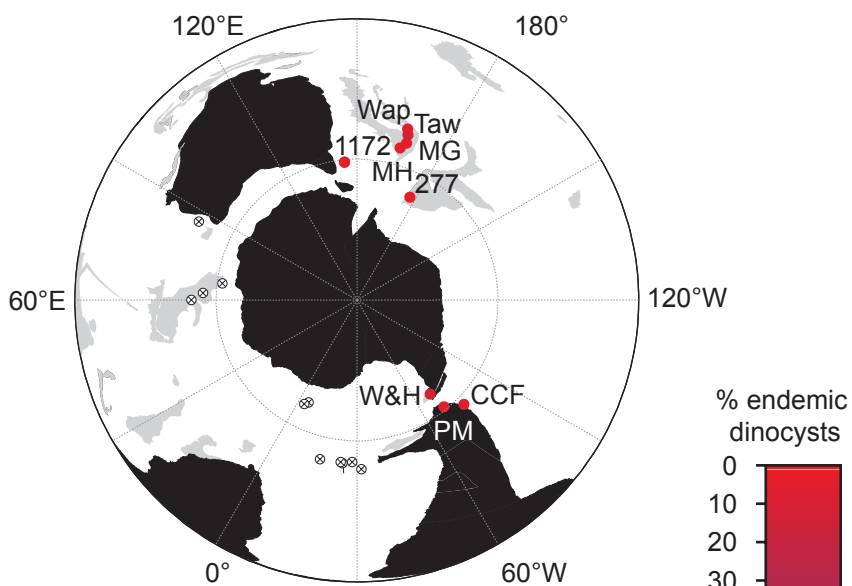
a

late Paleocene - early Eocene (56-54 Ma)
Thanetian - Ypresian
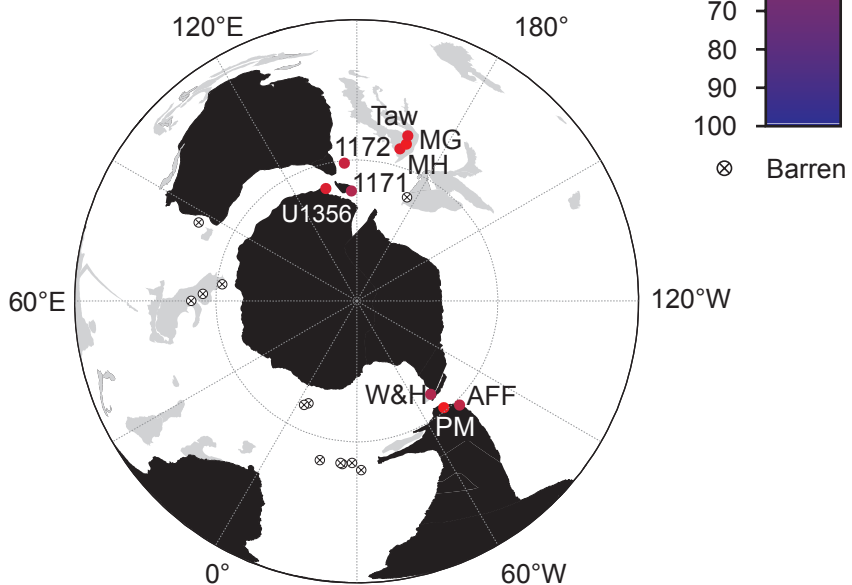
b

early Eocene (53-51 Ma)
Ypresian
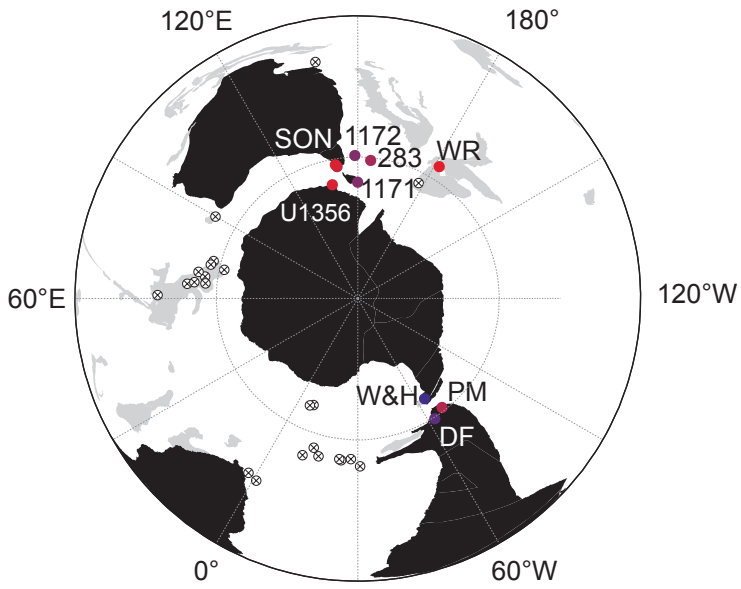
c   early - middle Eocene Boundary (50-48 Ma)
Ypresian - Lutetian
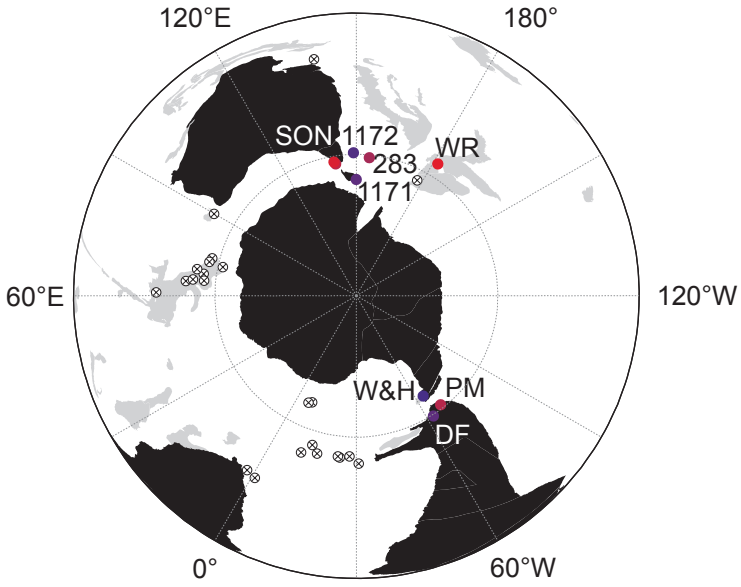
d   middle Eocene (47-45 Ma)
early Lutetian

e   middle Eocene (44-42 Ma)
late Lutetian

f   middle Eocene (41-39 Ma)
Lutetian - Bartonian

% endemic
dinocysts

%

Barren

middle - late Eocene (38-36 Ma)
Bartonian - Priabonian

g
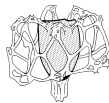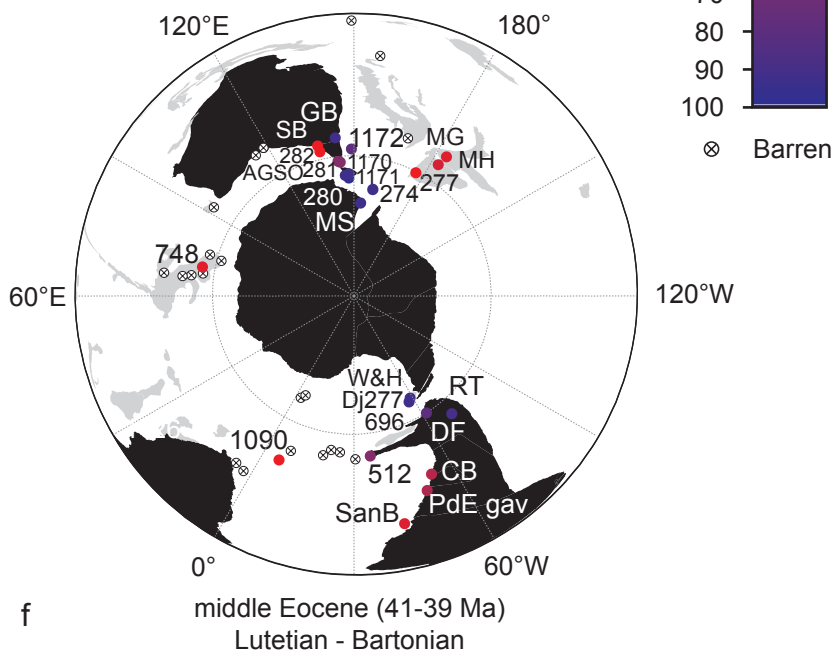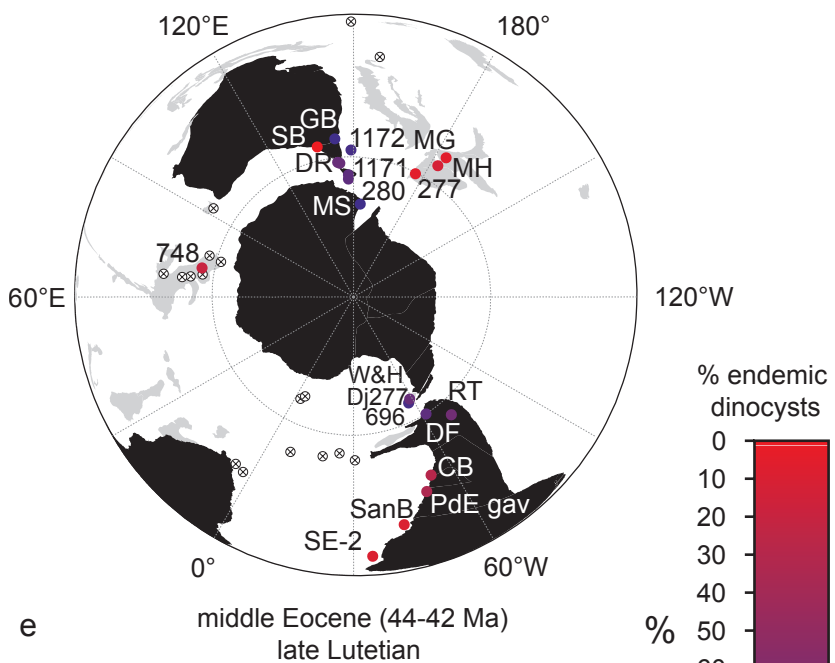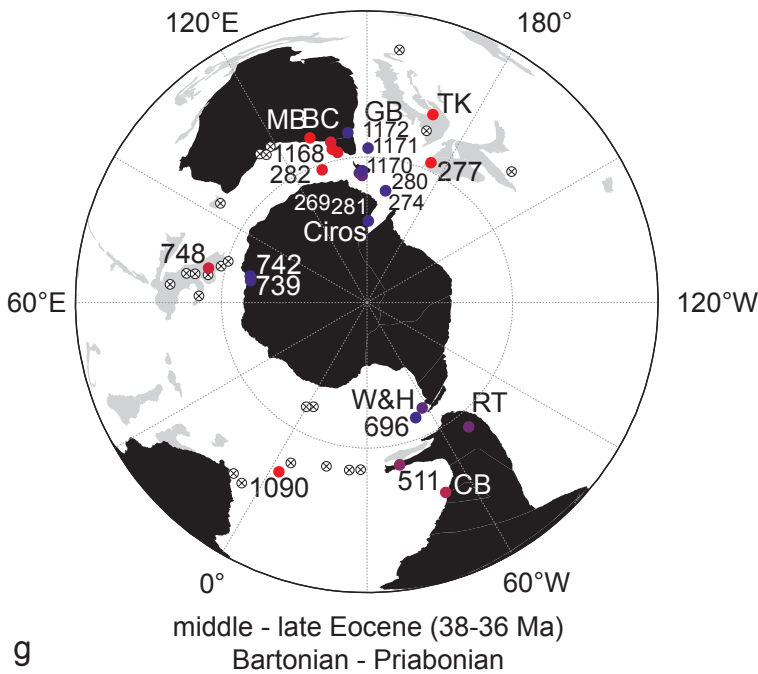
3.2. Statistical analysis

With our DCCA analyses, four canonical axes are established for the four environmental variables. The first axis explains 29% of the dinocyst data, the second axis another 4%. Fig. 4 shows the dinocyst scores for the first two axes with the environmental vectors. The SST vector points to the northeast, whereas the other vectors (i.e., thallium isotopes, % peridinioid dinocysts and % endemic dinocysts) are approximately perpendicular to the temperature vectors. In Fig. 4, the dinocyst groups are divided into endemic (blue triangles) and nonendemic (red circles). The endemic taxa are clearly separated from the nonendemic ones, with high scores on both axes for nonendemic and low scores for endemic dinocysts. The few exceptions include *Membranophoridium perforatum* (Mper) and *Hystrichosphaeridium truswelliae* (Htru). The latter has traditionally been interpreted as being endemic to the Southern Ocean. However, recently *H. truswelliae* was recorded in Paleocene- Eocene strata from New Jersey (northeast North America), clearly disputing its endemic signature (Sluijs and Brinkhuis 2009). 'Outliers' within the nonendemic taxa include the cosmopolitan *Enneadocysta* species (Enne), *Phthanoperidinium comatum* (Pcom) and *Selenopemphix* spp. (Sele). Comparing the scores of the first two axes to the environmental variables, we note that the SST variations correspond with the sample scores of the second axis. This axis explains only 4% of the dinocyst variation (Fig. 5). Patterns in the sample scores of the first axis correspond with the abundance of endemic dinocysts (Fig. 5).
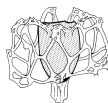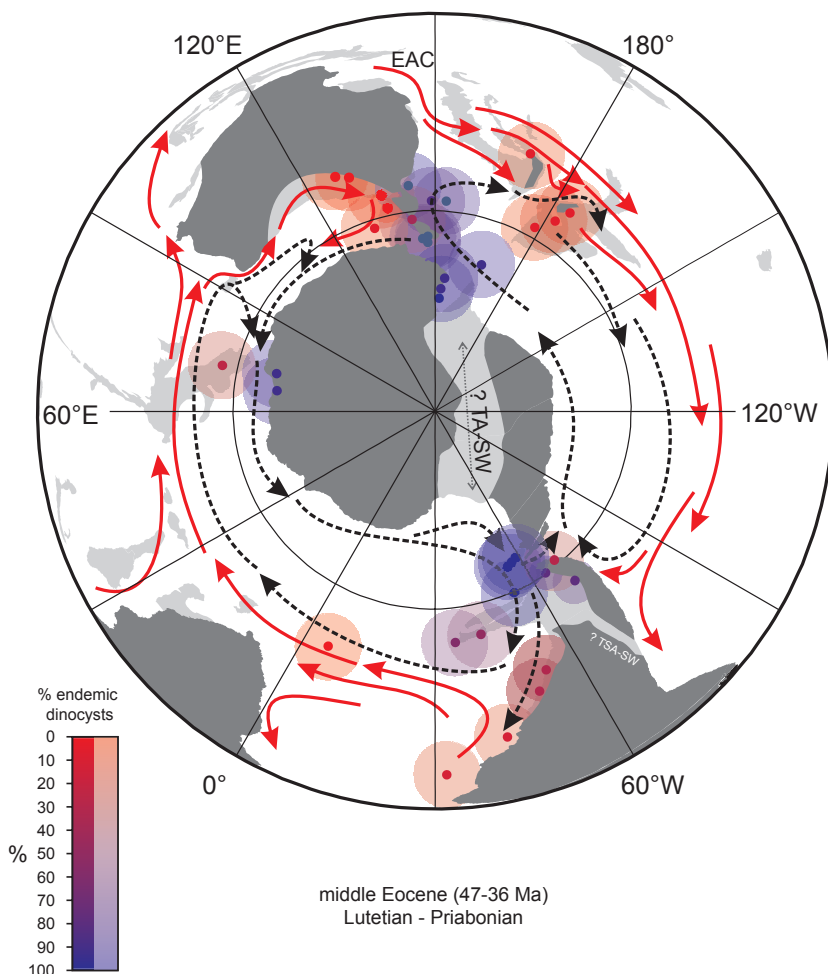
Figure 3. Generalized middle to late Eocene (47–36 Ma) dinocyst geographical distribution map overlain by the ocean circulation pattern inferred from the GCM experiments of (Huber et al., 2004). Maps derived from the Ocean Drilling Stratigraphic Network (ODSN). Shaded areas indicate mostly submerged continental blocks (Brown et al., 2006). Abbreviations: TA-SW, Trans-Antarctic Seaway (hypothetical; see Wrenn and Beckman (1982)); TSA-SW, Trans-South American Seaway (hypothetical; see Kohn et al. (2004) and Lagabrielle et al. (2009)); EAC, East Australian Current (Huber et al., 2004).

Thus, while spatial gradients in % endemic dinocysts appear to be a good indicator of the location of ocean currents, SST appears to have had little effect on the biogeography of endemic dinocysts over the complete record. However, during the interval of long-term cooling which began after the Early Eocene Climate Optimum (EECO; Chapter 2), abundance changes in endemic dinocysts do now covary with SST. For instance, during the Middle Eocene Climatic Optimum (MECO), a short-lived warming event at ~40 Ma (Bohaty et al., 2009), endemic dinocyst abundances fall again to below 50% (Fig. 5).

Figure 4. Detrended Canonical Correspondence Analysis (DCCA) plot, showing first and second axis scores for dinocyst species and vectors for the environmental variables. Note the clear distinction between cosmopolitan and low-latitude to mid-latitude species (red circles) in the northeastern side of the plot versus endemic or bipolar species (blue triangles) in the southwestern side. Cosmopolitan and low-latitude to midlatitude groups: APIC, gonyaulacoid dinocyst complex with apical archeopyle; Pcom, *Phthanoperidinium comatum*; GlAr, *Glaphyrocysta/Areoligera* complex; Wet, Wetzelielloid dinocysts; GONY, Gonyodomid dinocysts; Oper, *Operculodinium* complex; Pgol, *Palaeocystodinium golzowense*; FiCr, fibrous Cribroperidinioid dinocysts; SeAl, *Senegalinium/ Alterbidinium* complex; ElMi, *Elytrocysta/ Microdinium* spp.; Pauc, *Paucisphaeridium* complex; Impa, *Impagidinium* spp.; SpAc, *Spiniferites/ Achomosphaera* complex; Cere, *Cerebrocysta* spp.; Enne, *Enneadocysta* spp. (pars). Endemic and bipolar groups: Sele, *Selenopemphix* complex; Creg, *Corrudinium regulare*; Edic, *Enneadocysta dictyostila*; Oask, *Octodinium askiniae*; Aant, *Arachnodinium antarcticum*; VoSp, *Vozzhennikovia/ Spinidinium* complex; Psto, *Phthanoperidinium stockmansii* complex; Dant, *Deflandrea antarctica* complex; Htru, *Hystrichosphaeridium truswelliae*; Mper, *Membranophoridium perforatum*. Vectors: SST, TEX$_{86}$-derived SST from ODP Site 1172; Thal, [205]Tl thallium isotope record from (Nielsen et al., 2009); PERI, percentage of peridinioid dinocysts at ODP Site 1172; ENDE, percentage endemic/bipolar dinocysts at ODP Site 1172.

After the MECO, the cooling trend continues and endemic dinocysts re-dominate the assemblages (Fig. 5). To test the relation between dinocyst endemism and SST in the post- EECO interval of endemic dinocyst dominance, we performed another DCCA analysis involving only those data within the middle Eocene interval. In this interval, the variation explained by SST is ~20% (Fig. 6).

4. Discussion

4.1. Comparison with other fossil groups

While endemism is recorded within several fossil groups in the Eocene, a direct comparison with the dinocyst record of the southwest Pacific is hampered by a paucity of corresponding records around the Antarctic margin older than ~45 Ma. For instance, although diatomaceous (biosiliceous) sediments of early Cretaceous age and K-T boundary age occur in the western Antarctic margin (Gersonde and Harwood 1990; Harwood 1988; Harwood and Gersonde 1990), these appear to be a rare exception since most known biosiliceous records in Antarctic marginal (shelf) areas are mid Eocene in age (i.e., ODP Site 1172 Stickley et al., 2004a) or commonly much younger (e.g., Lazarus et al., 2008). Calcareous microfossil data from the southern high-latitudes predating 45 Ma are similarly scarce. Nevertheless, the few existing records of microfossils other than dinocysts from that time indeed show an increase in endemism beginning in the mid middle Eocene, as has been shown for radiolarians (Lazarus et al., 2008) and calcareous nannoplankton (Villa et al., 2008). Calcareous nannoplankton also exhibits a decrease in endemism during the MECO (Villa et al., 2008).

4.2. Potential driving forces: sea surface temperature, endemism and statistics

The establishment of dominant Antarctic endemic dinocyst assemblages characterizes the early middle Eocene transition (~50 Ma). However, SSTs did not decrease abruptly at that time (Fig. 5). This implies that the increase in abundance of endemic dinocysts was not linearly related to SST. In contrast, during the middle Eocene, abundances of endemic dinocysts are modulated by SST changes (Fig. 6). What then triggered endemic dinocysts to proliferate shortly after ~50 Ma? As stated above, the peak warmth of the EECO at ~50 Ma also marks the transition into the onset of high-latitude and global cooling that, despite an interruption by at least one short warming interval (i.e., the MECO, at ~40 Ma; Fig. 5), culminated in the onset of Antarctic glaciation during the Eocene-Oligocene transition (e.g., Zachos et al., 1992). It may be possible that the endemic dinoflagellate taxa, when they appeared, were capable to sustain in warm subtropical conditions during the EECO, but subsequently evolved to be more specialized to cooler surface waters during middle Eocene cooling. The sudden shift to dominant endemic dinocysts suggests the crossing of an environmental threshold at which cosmopolitan taxa were outcompeted by the endemic taxa. Because the DCCA statistical method identifies linear correspondence between assemblages and environmental data, this threshold may still have been SST.
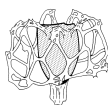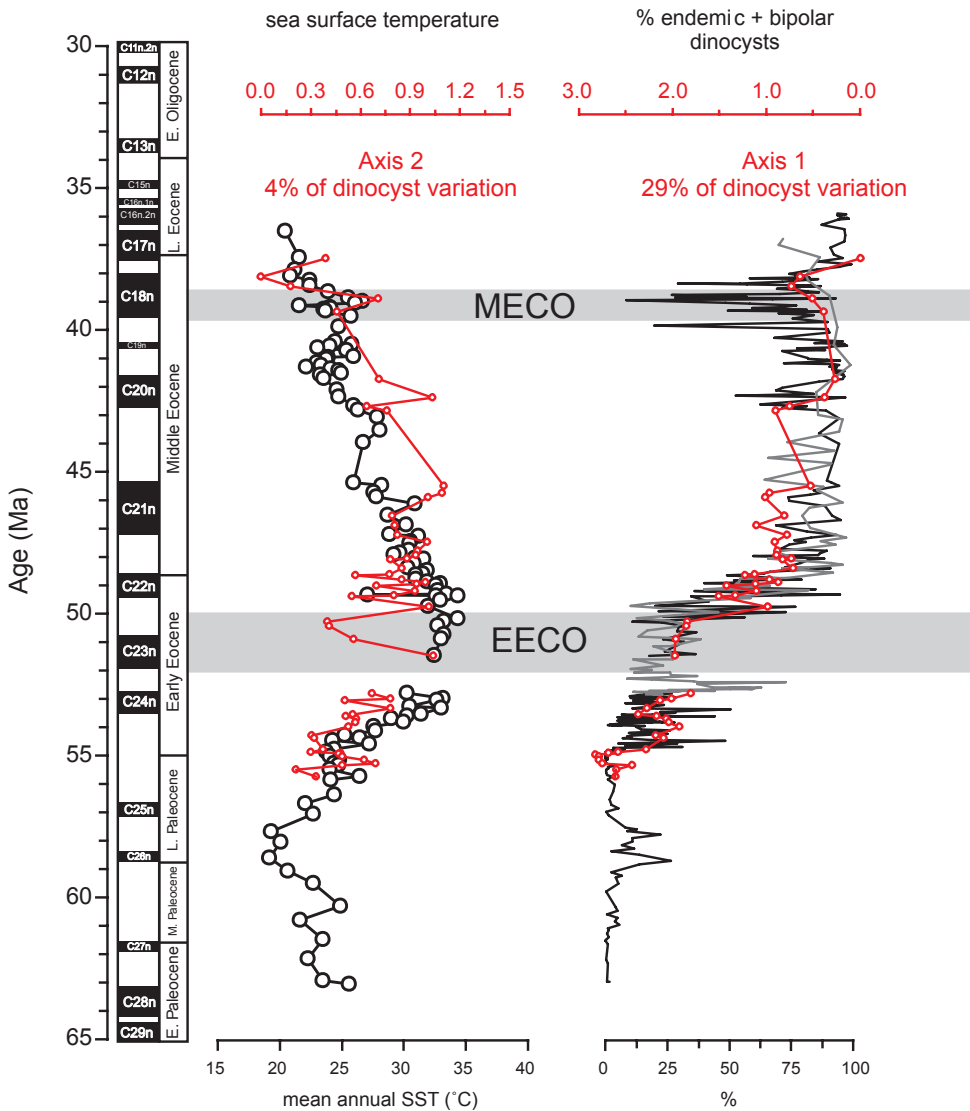
Figure 5. (left) TEX$_{86}$-based SST curve from ODP Site 1172 from Chapter 2 overlain by the second axis scores of the DCCA analysis. Note that the second axis explains only 4% of the total dinocyst assemblage variation. (right) Endemic dinocyst abundances at ODP Sites 1172 (black) and 1171 (grey) overlain by the first axis scores.

The mechanism described above may provide an explanation for the proliferation of endemic dinocysts just after 50 Ma, but it fails to explain the presence, in low abundance, of endemic dinocysts during Paleocene and early Eocene times. The most abundant endemic dinocyst groups, i.e., *Deflandrea antarctica*, *Phthanoperidinium stockmansii* group and *Vozzhennikovia/Spinidinium* spp., are

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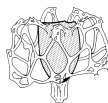
already present in the Paleocene. Apparently, Paleocene– early Eocene surface waters were not as favourable for endemic species as they were in the middle Eocene. Hence, along with SST changes, other physio-chemical parameters of the surface waters must have changed between the Paleocene and middle Eocene that stimulated Southern Ocean endemism to proliferate just after the early middle Eocene boundary. A possible explanation may lie in the diatom record at ODP Site 1172, where opal-A is preserved from approximately 47 Ma (Stickley et al., 2004a, b) or 45 Ma (Chapter 2) onward (Fig. 7). In the middle Eocene, global cooling enhanced latitudinal gradients (Chapter 2). This cooling stimulated upwelling, bringing silica-enriched waters that characterized Eocene oceans (e.g., McGowran 1989), particularly the Pacific (e.g., Moore 2008), to the surface for utilization by the diatoms. This change apparently brought about significant diatom production and preservation to the region at ODP Site 1172, and coincides with a global evolutionary turnover event within the diatoms and globally enhanced preservation of opal-A associated with global cooling (Baldauf and Barron 1990; Barron and Baldauf 1989; McGowran 1989).
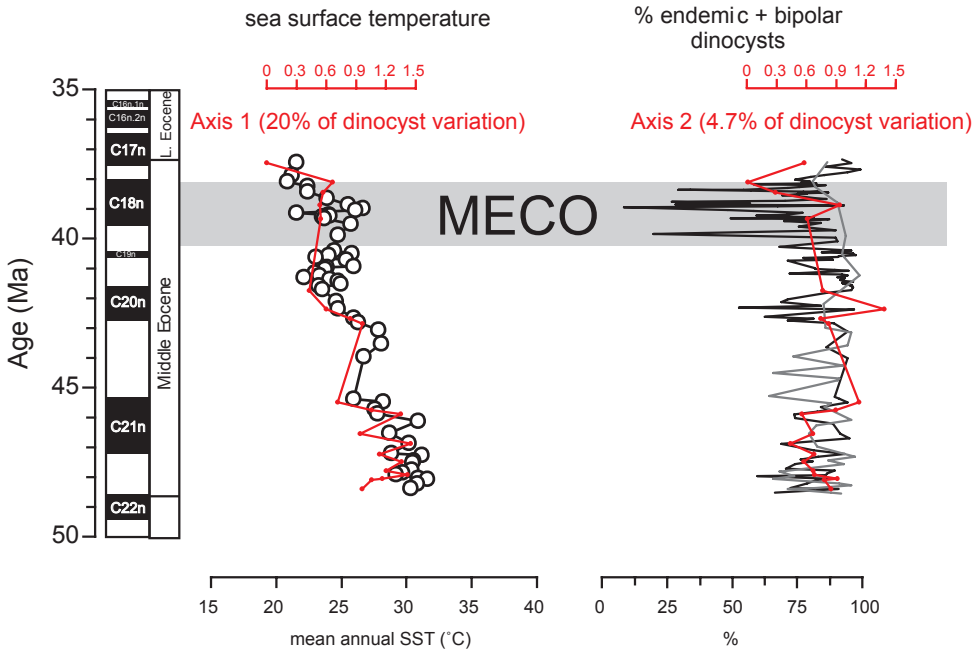
Figure 6. (left) Middle Eocene TEX$_{86}$-based SST curve from ODP Site 1172 (Chapter 2) overlain by the first axis scores of the middle Eocene DCCA experiment (in red; 20% of the dinocyst variation). The second axis scores correspond to (right) the % endemic dinocyst curve but explain only 4.7% of the variation in the dinocyst assemblage.

Increased weathering may have led to an additional nutrient supply to the oceans including both Si and the diatom bio-limiting nutrient Fe (Boyd et al., 2000). It appears plausible that the endemic dinocysts that proliferated during the earliest middle Eocene benefited from the same circulation changes and eutrophication phase

that stimulated diatom productivity at that time. Further evidence for changes in ocean-wide surface water fertility comes from thallium isotopes measured on ferromanganese crusts from the Pacific Ocean (Fig. 7; Nielsen et al., 2009). A shift in these isotopes together with a coeval shift in sulphur isotopes suggests a major ocean-wide increase in marine organic carbon export from 55 to 45 Ma. At the same time, deep-water carbon isotopic values between oceanic basins start to diverge, suggesting (among other possibilities) an increase in the biological pump leading to increased water mass aging gradients (Fig. 7; Sexton et al., 2006). The fertilization of the oceanic basins is, as suggested above, likely attributed to increased upwelling and weathering associated with global cooling and ever-increasing latitudinal gradients.
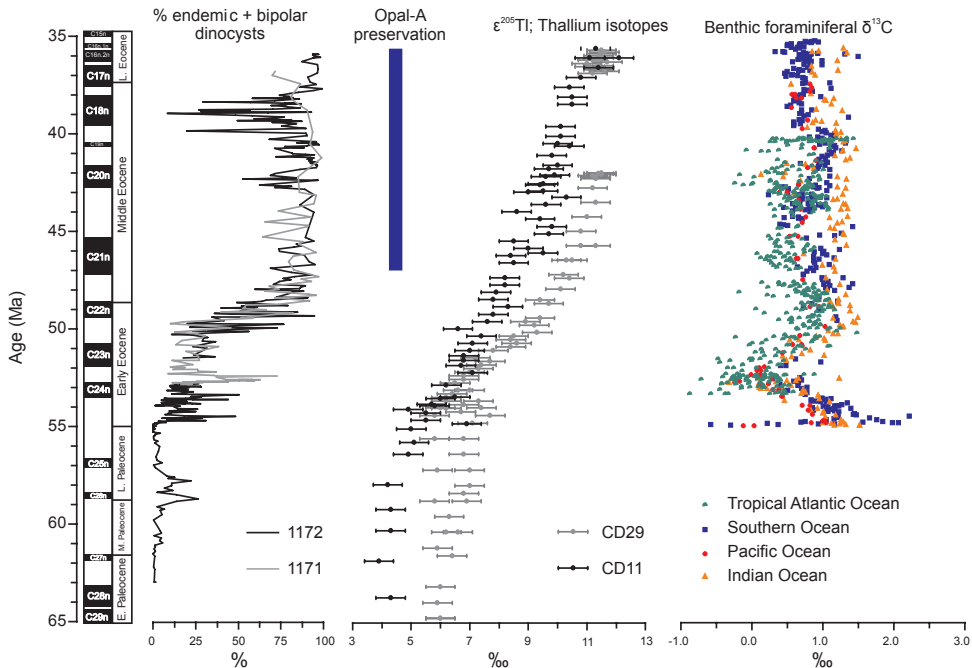
Figure 7. The relative abundance of endemic dinocysts at ODP Site 1172 (black) and Site 1171 (grey) show a major increase at 50 Ma. Opal-A preservation at ODP Site 1172 commenced between 45 Ma (Lazarus et al., 2008) and 47 Ma (Stickley et al., 2004a) probably as a result of increased latitudinal gradients stimulating upwelling and productivity (see text). Thallium isotope records from the Pacific Ocean, measured on two ferromanganese crusts labelled "CD11" (black) and "CD29" (grey), obtained from (Nielsen et al., 2009), show a major increase from 55 to 45 Ma, which has been shown to suggest a permanent increase in marine organic carbon productivity. Benthic foraminiferal carbon isotope data from the tropical Atlantic Ocean (half circles), Southern Ocean (squares), Pacific Ocean (circles), and Indian Ocean (triangles) diverge from ~50 Ma upward, indicating stronger carbon isotope gradients between oceanic basins, which may suggest increased carbon pumping (data from Sexton et al. (2006)).
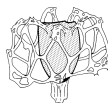
5. Conclusions

New data and reviews of previous reports allow a considerably improved spatio-temporal reconstruction of Paleogene circum-Antarctic dinocyst distribution patterns. Dinocysts can be divided into Antarctic endemic, bipolar, middle- to low-latitude, and cosmopolitan groups. Their spatial distribution patterns match ocean current patterns generated through GCM experiments (Huber et al., 2004) during the middle to late Eocene. Despite open connections between the Southern Ocean and adjacent mid-latitude oceans, (Exon et al., 2004), circum-Antarctic dinocyst assemblages switched from a cosmopolitan and low-latitude-derived dinoflagellate cyst assemblage to an endemic-dominated assemblage in the early middle Eocene. By numerically evaluating the relation between Southern Ocean dinocyst assemblages and the SST evolution at ODP Site 1172, we identify that SST was not the primary driving factor in the establishment of dominantly endemic dinocyst assemblages. We conjecture that ocean-wide surface water ocean fertility (Nielsen et al., 2009; Sexton et al., 2006) brought about at post-EECO cooling (Chapter 2; Sexton et al., 2006), helped to stimulate diatom production and 'switch on' opal preservation in the area and caused a rapid reorganization of high-latitude dinocyst assemblages. Once the endemic community became the dominant component of the dinocyst assemblages in the early middle Eocene southwest Pacific, the influence of SST variations on the dinocyst assemblage was stronger than during the Paleocene and early Eocene.
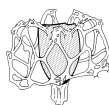
Supplements to Chapter 3

Table S1: Site list

	Leg	Site	type	location	Long	Lat	depth	references	# of samples	bottom age (Ma)	% endemic dinocysts
Thanetian-Ypresian Boundary 56-54 Ma											
dsdp	29	277	drill	campbell plateau	166.11	-52.13	1214	Kennett and Houtz, 1975; Hollis et al., 1997	2	56	5
odp	189	1172	drill	east tasman plateau	149.56	-43.58	2622	Exon et al., 2001/2004; Sluijs et al., 2003; Brinkhuis et al., 2003a; Stickley et al., 2004a	>10	66	5
chorillo chico fm.		CCF	outcr	brunswick peninsula	-71.33	-53.17	N/A	Quattrocchio and Sarjeant 2003 Quattrocchio, 2009	>10	62	5
muzzle group		MG	outcr	new zealand	174.00	-42.00	N/A	Strong et al, 1995	8	66	5
moeraki-hampden		MH	outcr	new zealand	45.00	-45.00	N/A	Wilson, 1985; Crouch 2001	>10	56	5
peninsula mitre		PM	outcr	southeast tierra del fuego	-68.78	-54.67	N/A	Quattrocchio and Sarjeant, 2003	5	54	5
tawanui section		Taw	outcr	new zealand	169.53	-46.50	N/A	Crouch 2001	>10	57	5
		W&H	outcr	seymour island	-56.17	-64.17	N/A	Wrenn and Hart, 1988	>10	58	10
waipawa section		Wap	outcr	new zealand	175.50	-41.40	N/A	Wilson, 1988; Crouch, 2001	>10	56	15
dsdp	27	259	drill	off west australia	112.42	-29.37	4696	Veevers et al., 1974	N/A	N/A	barren
dsdp	36	328	drill	east of falkland plateau	-36.40	-49.49	5095	Barker et al., 1976; Gombos (p. 575); Harris (p. 761)	N/A	N/A	barren
odp	113	689	drill	maud rise	3.06	-64.31	2080	Barker et al., 1988/1990; Florindo and Roberts, 2005; Diester-Haass and Zahn, 2001	N/A	N/A	barren
odp	113	690	drill	maud rise	1.12	-65.10	2914	Barker et al., 1988/1990; Florindo and Roberts, 2005;	N/A	N/A	barren
odp	114	699	drill	georgia rise	-30.41	-51.33	3706	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	700	drill	georgia rise	-30.16	-51.31	3601	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	698	drill	georgia rise	-33.06	-51.28	2138	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	702	drill	mid atlantic ridge	-26.22	-50.57	3083	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	119	738	drill	kerguelen plateau	82.47	-62.43	2253	Barron et al., 1989	N/A	N/A	barren
odp	120	747	drill	kerguelen plateau	76.48	-54.49	1695	Wise et al., 1989/1992	N/A	N/A	barren
odp	120	748	drill	kerguelen plateau	78.59	-58.26	1288	Wise et al., 1989/1992; Mohr, 1990; Roberts et al., 2003	N/A	N/A	barren
Ypresian 53-51 Ma											
odp	189	1171	drill	south tasman rise	149.07	-48.30	2150	Exon et al., 2001/2004; Sluijs et al., 2003	>10	54	30
odp	189	1172	drill	east tasman plateau	149.56	-43.58	2622	Exon et al., 2001/2004; Sluijs et al., 2003; Brinkhuis et al., 2003a; Stickley et al., 2004a	>10	66	20
muzzle group		MG	outcr	new zealand	174.00	-42.00	N/A	Strong et al, 1995	5	66	5
moeraki-hampden		MH	outcr	new zealand	45.00	-45.00	N/A	Wilson, 1985; Crouch, thesis 2001;	>10	56	5
peninsula mitre		PM	outcr	sotheast tierra del fuego	-68.78	-54.67	N/A	Quattrocchio and Sarjeant, 2003	5	54	0
tawanui section		Taw	outcr	new zealand	169.53	-46.50	N/A	Crouch, thesis 2001	>10	57	5
		W&H	outcr	seymour island	-56.17	-64.17	N/A	Wrenn and Hart, 1988	>10	58	30
iodp	318	U1356	drill	wilkes land	136.05	-63.35	4003	Escutia et al., 2010	6	54	10
agua fresca form.		AFF	outcr	brunswick peninsula	-71.33	-53.17	N/A	Quattrocchio, 2009	>10	54	25
dsdp	27	259	drill	off west australia	112.42	-29.37	4696	Veevers et al., 1974	N/A	N/A	barren
dsdp	29	277	drill	campbell plateau	166.11	-52.13	1214	Kennett and Houtz, 1975; Hollis et al., 1997	N/A	N/A	barren
dsdp	36	328	drill	east of falkland plateau	-36.40	-49.49	5095	Barker et al., 1976; Gombos (p. 575); Harris (p. 761)	N/A	N/A	barren
odp	113	689	drill	maud rise	3.06	-64.31	2080	Barker et al., 1988/1990; Florindo and Roberts, 2005; Diester-Haass and Zahn, 2001	N/A	N/A	barren
odp	113	690	drill	maud rise	1.12	-65.10	2914	Barker et al., 1988/1990; Florindo and Roberts, 2005;	N/A	N/A	barren
odp	114	698	drill	georgia rise	-33.06	-51.28	2138	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	699	drill	georgia rise	-30.41	-51.33	3706	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	700	drill	georgia rise	-30.16	-51.31	3601	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	698	drill	georgia rise	-33.06	-51.28	2138	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	702	drill	mid atlantic ridge	-26.22	-50.57	3083	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	119	738	drill	kerguelen plateau	82.47	-62.43	2253	Barron et al., 1989	N/A	N/A	barren
odp	120	747	drill	kerguelen plateau	76.48	-54.49	1695	Wise et al., 1989/1992	N/A	N/A	barren
odp	120	748	drill	kerguelen plateau	78.59	-58.26	1288	Wise et al., 1989/1992; Mohr, 1990; Roberts et al., 2003	N/A	N/A	barren
Ypresian-Lutetian Boundary (50-48 Ma)											
dsdp	29	283	drill	tasman sea	154.17	-43.55	4729	Kennett and Houtz 1975; Haskell and Wilson (p. 723); Warnaar, 2006	>10	65	40
odp	189	1171	drill	south tasman rise	149.07	-48.30	2150	Exon et al., 2001/2004; Sluijs et al., 2003	>10	54	60
odp	189	1172	drill	east tasman plateau	149.56	-43.58	2622	Exon et al., 2001/2004; Sluijs et al., 2003; Brinkhuis et al., 2003a; Stickley et al., 2004a	>10	66	60
sonne 36, 22sl		SON	grav	tasman margin (west)	144.25	-43.37	3726	Truswell, 1997	4	50	10
sonne 36, 30kl		SON	grav	tasman margin (west)	145.42	-43.47	1755	Truswell, 1997	4	50	25
la despedida fm.		DF	outcr	northeast tierra del fuego	-68.83	-53.88	N/A	Guerstein et al., 2008	4	48	45
península mitre		PM	outcr	sotheast tierra del fuego	-68.78	-54.67	N/A	Quattrocchio and Sarjeant, 2003	5	54	30
		W&H	outcr	seymour island	-56.17	-64.17	N/A	Wrenn and Hart, 1988	>10	58	90
Waipara river		WR	outcr	New Zealand	172.00	-43.04	N/A	Hollis et al., 2009	>10	50	5
iodp	318	U1356	drill	wilkes land	136.05	-63.35	4003	Escutia et al., 2010	6	54	10
dsdp	26	255	drill	indian ocean (east)	93.44	-31.08	1144	Davies et al., 1974	N/A	N/A	barren
dsdp	28	264	drill	offshore sw australia	112.03	-34.58	2873	Hayes et al., 1975	N/A	N/A	barren
dsdp	29	277	drill	campbell plateau	166.11	-52.13	1214	Kennett and Houtz, 1975; Hollis et al., 1997	N/A	N/A	barren
dsdp	30	287	drill	coral sea basin	153.16	-15.55	4653	Andrews et al., 1975	N/A	N/A	barren
dsdp	36	328	drill	east of falkland plateau	-36.40	-49.49	5095	Barker et al., 1976; Gombos (p. 575); Harris (p. 761)	N/A	N/A	barren
dsdp	40	360	drill	off south africa/agulhas	18.06	-35.51	2949	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
dsdp	40	361	drill	off south africa/agulhas	15.27	-35.04	4549	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
odp	113	689	drill	maud rise	3.06	-64.31	2080	Barker et al., 1988/1990; Florindo and Roberts, 2005; Diester-Haass and Zahn, 2001	N/A	N/A	barren
odp	113	690	drill	maud rise	1.12	-65.10	2914	Barker et al., 1988/1990; Florindo and Roberts, 2005; Diester-Haass and Zahn, 2001	N/A	N/A	barren
odp	114	698	drill	georgia rise	-33.06	-51.28	2138	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	699	drill	georgia rise	-30.41	-51.33	3706	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	700	drill	georgia rise	-30.16	-51.31	3601	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	701	drill	mid atlantic ridge	-23.13	-51.59	4637	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	702	drill	mid atlantic ridge	-26.22	-50.57	3083	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	703	drill	meteor rise	7.54	-47.03	1796	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	119	737	drill	kerguelen-heard plateau	73.02	-50.14	564	Barron et al., 1989; Wei and Thierstein, 1991	N/A	N/A	barren
odp	119	738	drill	kerguelen plateau	82.47	-62.43	2253	Barron et al., 1989	N/A	N/A	barren
odp	120	747	drill	kerguelen plateau	76.48	-54.49	1695	Wise et al., 1989/1992	N/A	N/A	barren
odp	120	748	drill	kerguelen plateau	78.59	-58.26	1288	Wise et al., 1989/1992; Mohr, 1990; Roberts et al., 2003	N/A	N/A	barren
odp	120	749	drill	kerguelen plateau	76.24	-58.43	1071	Wise et al., 1989/1992	N/A	N/A	barren
odp	120	750	drill	kerguelen plateau	81.14	-57.35	2030	Wise et al., 1989/1992	N/A	N/A	barren
odp	183	1135	drill	kerguelen plateau (s)	84.16	-59.42	1567	Coffin et al., 2000; 2003	N/A	N/A	barren
odp	183	1136	drill	kerguelen plateau (s)	84.50	-59.39	1930	Coffin et al., 2000; 2003	N/A	N/A	barren
odp	183	1138	drill	kerguelen plateau	75.59	-53.33	1141	Coffin et al., 2000; 2003; Bohaty and Zachos, 2003	N/A	N/A	barren
mid-Lutetian (47-45 Ma)											
dsdp	29	283	drill	tasman sea	154.17	-43.55	4729	Kennett and Houtz, 1975; Hollis et al., 1997		65	40
odp	189	1171	drill	south tasman rise	149.07	-48.30	2150	Exon et al., 2001/2004; Sluijs et al., 2003	>10	54	80
odp	189	1172	drill	east tasman plateau	149.56	-43.58	2622	Exon et al., 2001/2004; Sluijs et al., 2003; Brinkhuis et al., 2003a; Stickley et al., 2004a	>10	66	85

Name	Leg	Site	Method	Location	Long	Lat	Depth	Reference			
sonne 36, 22sl		SON	grav	tasman margin (west)	144.25	-43.37	3726	Truswell, 1997	4	50	10
sonne 36, 30kl		SON	grav	tasman margin (west)	145.42	-43.47	1755	Truswell, 1997	4	50	25
la despedida fm.		DF	outcr	northeast tierra del fuego	-68.83	-53.88	N/A	Guerstein et al., 2008	6	48	75
península mitre		PM	outcr	sotheast tierra del fuego	-68.78	-54.67	N/A	Quattrocchio and Sarjeant, 2003	3	54	30
		W&H	outcr	seymour island	-56.17	-64.17	N/A	Wrenn and Hart, 1988	>10	58	90
Waipara river		WR	outcr	New Zealand	172.00	-43.04	N/A	Hollis et al., 2009	>10	50	5
dsdp	26	255	drill	indian ocean (east)	93.44	-31.08	1144	Davies et al., 1974	N/A	N/A	barren
dsdp	28	264	drill	offshore sw australia	112.03	-34.58	2873	Hayes et al., 1975	N/A	N/A	barren
dsdp	29	277	drill	campbell plateau	166.11	-52.13	1214	Kennett and Houtz, 1975; Hollis et al., 1997	N/A	N/A	barren
dsdp	30	287	drill	coral sea basin	153.16	-15.55	4653	Andrews et al., 1975	N/A	N/A	barren
dsdp	36	328	drill	east of falkland plateau	-36.40	-49.49	5095	Barker et al., 1976; Gombos (p. 575); Harris (p. 761)	N/A	N/A	barren
dsdp	40	360	drill	off south africa/agulhas	18.06	-35.51	2949	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
dsdp	40	361	drill	off south africa/agulhas	15.27	-35.04	4549	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
odp	113	689	drill	maud rise	3.06	-64.31	2080	Barker et al., 1988/1990; Florindo and Roberts, 2005; Diester-Haass and Zahn, 2001	N/A	N/A	barren
odp	113	690	drill	maud rise	1.12	-65.10	2914	Barker et al., 1988/1990; Florindo and Roberts, 2005; Diester-Haass and Zahn, 2001	N/A	N/A	barren
odp	114	698	drill	georgia rise	-33.06	-51.28	2138	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	699	drill	georgia rise	-30.41	-51.33	3706	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	700	drill	georgia rise	-30.16	-51.31	3601	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	701	drill	mid atlantic ridge	-23.13	-51.59	4637	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	702	drill	mid atlantic ridge	-26.22	-50.57	3083	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	703	drill	meteor rise	7.54	-47.03	1796	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	120	747	drill	kerguelen plateau	76.48	-54.49	1695	Wise et al., 1989/1992	N/A	N/A	barren
odp	120	748	drill	kerguelen plateau	78.59	-58.26	1288	Wise et al., 1989/1992; Mohr, 1990; Roberts et al., 2003	N/A	N/A	barren
odp	120	749	drill	kerguelen plateau	76.24	-58.43	1071	Wise et al., 1989/1992	N/A	N/A	barren
odp	120	750	drill	kerguelen plateau	81.14	-57.35	2030	Wise et al., 1989/1992	N/A	N/A	barren
odp	183	1135	drill	kerguelen plateau (s)	84.16	-59.42	1567	Coffin et al., 2000; 2003	N/A	N/A	barren
odp	183	1136	drill	kerguelen plateau (s)	84.50	-59.39	1930	Coffin et al., 2000; 2003	N/A	N/A	barren
odp	183	1138	drill	kerguelen plateau	75.59	-53.33	1141	Coffin, Frey et al, 2000; 2003; Bohaty and Zachos, 2003	N/A	N/A	barren
mid-Lutetian (44-42 Ma)											
dsdp	29	277	drill	campbell plateau	166.11	-52.13	1214	Kennett and Houtz, 1975; Hollis et al., 1997	3	56	0
dsdp	29	280	drill	tasmania	147.14	-48.57	4176	Kennett and Houtz, 1975; Haskell and Wilson (p. 723);	6	49	80
dsdp	29	281	drill	tasmania	147.46	-47.60	1591	Kennett and Houtz, 1975;Haskell and Wilson (p. 723); Perch-Nielsen (p. 875); Crouch & Hollis 1996	2	44	90
odp	113	696	drill	weddell sea	-42.56	-61.51	650	Barker et al., 1988/1990; Mao and Mohr; 1995;	>10	43	90
odp	120	748	drill	kerguelen plateau	78.59	-58.26	1288	Wise et al., 1989/1992; Mohr, 1990; Roberts et al., 2003	>10	92	20
odp	189	1171	drill	south tasman rise	149.07	-48.30	2150	Exon et al., 2001/2004; Sluijs et al., 2003	>10	54	90
odp	189	1172	drill	east tasman plateau	149.56	-43.58	2622	Exon et al., 2001/2004; Sluijs et al., 2003; Brinkhuis et al., 2003a; Stickley et al., 2004a	>10	66	80
agso 147, dr14		AGZO	dred	tasman margin (west)	144.19	-45.02	4100	Truswell, 1997	1	45	75
agso 147, dr18		AGZO	dred	south tasman rise	144.57	-46.47	3450	Truswell, 1997	1	45	75
agso 147, gco1		AGZO	grav	tasman margin (west)	144.11	-44.10	4264	Truswell, 1997	1	45	75
gaviotin		PdE	drill	off uruguay	-54.75	-36.62	N/A	Daners et al., 2004; Guerstein and Daners, in press	3	44	15
pejerrey		CB	drill	offshore argentina	-55.50	-38.83	N/A	Archangelsky et al.,1997; Guerstein and Daners, in press	5	44	30
la despedida fm.		DF	outcr	northest tierra del fuego	-68.83	-53.88	N/A	Guerstein et al., 2008	6	48	75
la meseta fm.		Dj277	outcr	seymour island	58.45	-64.10	N/A	Cocozza and Clarke, 1992	10	45	75
		GB		gippsland basin	136.70	-38.30	N/A	Holdgate et al., 1995; 2000; 2003; Macphail, 2001	3	45	90
muzzle group		MG	outcr	new zealand	174.00	-42.00	N/A	Strong et al., 1995	4	66	75
moeraki-hampden		MH	outcr	new zealand	45.00	-45.00	N/A	Wilson, 1985; Burgess, 2008	>10	43	14
		MS	drill	mcmurdo sound	-166.50	-78.08	N/A	Wilson, 1967a; Cranwell et al., 1960; Cranwell 1964	4	44	90
rio turbio		RT	core	southwest santa cruz	-72.30	-51.50	N/A	Archangelsky, 1969; Guerstein, pers. obs.	8	46	70
santos b.		SanB	drill	off brazil	-45.00	-35.00	N/A	Griggs, 1981 in: Wrenn and Hart, 1988	unknown	45	10
clam-1		SB	drill	sorell basin	144.15	-40.50	N/A	(Esso) Boreham et al., 2002	4	45	0
cape sorell-1		SB	drill	sorell basin	145.02	-42.08	94	Amoco, 1982	4	45	0
sergipe basin		SE-2	core	sergipe basin, offshore brazil	-37.80	-11.20	N/A	Ferreira, 2004 in Guerstein and Daners, in press	>10	44	5
		W&H	outcr	seymour island	-56.17	-64.17	N/A	Wrenn and Hart, 1988	>10	58	90
dsdp	26	253	drill	ninetyeast ridge	87.22	-24.53	1962	Davies et al., 1974	N/A	N/A	barren
dsdp	28	264	drill	offshore sw australia	112.03	-34.58	2873	Hayes et al., 1975	N/A	N/A	barren
dsdp	30	286	drill	south new hebrides trench	166.22	-16.32	4465	Andrews et al., 1975	N/A	N/A	barren
dsdp	30	289	drill	Ontong-Java Plateau	158.30	0.29	2206	Andrews et al., 1975	N/A	N/A	barren
dsdp	40	360	drill	off south africa/agulhas	18.06	-35.51	2949	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
dsdp	40	361	drill	off south africa/agulhas	15.27	-35.04	4549	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
odp	113	689	drill	maud rise	3.06	-64.31	2080	Barker et al., 1988/1990; Florindo and Roberts, 2005; Diester-Haass and Zahn, 2001	N/A	N/A	barren
odp	114	698	drill	georgia rise	-33.06	-51.28	2138	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	702	drill	mid atlantic ridge	-26.22	-50.57	3083	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	703	drill	meteor rise	7.54	-47.03	1796	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	119	737	drill	kerguelen-heard plateau	73.02	-50.14	564	Barron et al., 1989; Wei and Thierstein, 1991	N/A	N/A	barren
odp	119	738	drill	kerguelen plateau	82.47	-62.43	2253	Barron et al., 1989	N/A	N/A	barren
odp	120	747	drill	kerguelen plateau	76.48	-54.49	1695	Wise et al., 1989/1992	N/A	N/A	barren
odp	120	749	drill	kerguelen plateau	76.24	-58.43	1071	Wise et al., 1989/1992	N/A	N/A	barren
odp	183	1135	drill	kerguelen plateau (s)	84.16	-59.42	1567	Coffin et al., 2000; 2003	N/A	N/A	barren
odp	183	1136	drill	kerguelen plateau (s)	84.50	-59.39	1930	Coffin et al., 2000; 2003	N/A	N/A	barren
Bartonian (41-39 Ma)											
dsdp	29	277	drill	campbell plateau	166.11	-52.13	1214	Kennett and Houtz, 1975; Hollis et al., 1997	1	56	0
dsdp	29	280	drill	tasmania	147.14	-48.57	4176	Kennett and Houtz, 1975; Haskell and Wilson (p. 723); Perch-Nielsen (p. 875); Crouch & Hollis 1996	6	49	95
	71	512	drill	north of weddell sea	-40.52	-49.52	1846	Ludwig et al., 1983; Goodman & Ford (p. 859);	10	42	60
odp	113	696	drill	weddell sea	-42.56	-61.51	650	Barker et al., 1988/1990; Mao and Mohr; 1995;	>10	43	95
odp	120	748	drill	kerguelen plateau	78.59	-58.26	1288	Wise et al., 1989/1992; Mohr, 1990; Roberts et al., 2003	>10	92	5
odp	177	1090	drill	agulhas ridge	8.54	-42.55	3710	Gersonde et al., 1999; 2001; Latimer and Filippelli, 2002;	>10	42	0
odp	189	1170	drill	south tasman rise	146.03	-47.09	2704	Exon et al., 2001/2004; Sluijs et al., 2003	>10	41	90
odp	189	1171	drill	south tasman rise	149.07	-48.30	2150	Exon et al., 2001/2004; Sluijs et al., 2003	>10	54	93
odp	189	1172	drill	east tasman plateau	149.56	-43.58	2622	Exon et al., 2001/2004; Sluijs et al., 2003; Brinkhuis et al., 2003a; Stickley et al., 2004a	>10	66	80
gaviotin		PdE	drill	off uruguay	-54.75	-36.62	N/A	Daners et al., 2004; Guerstein and Daners, in press	3	44	35
pejerrey		CB	drill	offshore argentina	-55.50	-38.83	N/A	Archangelsky et al.,1997; Guerstein and Daners, in press	5	44	30
la despedida fm.		DF	outcr	northest tierra del fuego	-68.83	-53.88	N/A	Guerstein et al., 2008	6	48	75
la meseta fm.		Dj277	outcr	seymour island	58.45	-64.10	N/A	Cocozza and Clarke, 1992	10	45	90
		GB		gippsland basin	136.70	-38.30	N/A	Holdgate et al., 1995; 2000; 2003; Macphail, 2001	3	45	90
muzzle group		MG	outcr	new zealand	174.00	-42.00	N/A	Strong et al., 1995	4	66	5
Hampden section		MH	outcr	New zealand	45.00	-45.00	N/A	Burgess et al., 2008	>10	42	15
		MS	outcr	mcmurdo sound	-166.50	-78.08	N/A	Wilson, 1967a; Cranwell et al., 1960; Cranwell 1964	4	44	95
rio turbio		RT	core	southwest santa cruz	-72.30	-51.50	N/A	Archangelsky, 1969; Guerstein, pers. obs.	7	45	35
santos b.		SanB	drill	off brazil	-45.00	-35.00	N/A	Griggs, 1981 in: Wrenn and Hart, 1988	unknown	45	10
clam-1		SB	drill	sorell basin	144.15	-40.50	102	(Esso) Boreham et al., 2002	4	45	0
cape sorell-1		SB	drill	sorell basin	145.02	-42.08	94	Amoco, 1982	4	45	0
agso 147, dr14		AGZO	grav	tasman margin (west)	144.11	-44.10	4264	Truswell, 1997	1	45	60

agso 147, dr18		AGZO	dred	tasman margin (west)	144.19	-45.02	4100	Truswell, 1997	1	45	60
agso 147, gco1		AGZO	dred	south tasman rise	144.57	-46.47	3450	Truswell, 1997	1	45	60
		W&H	outcr	seymour island	-56.17	-64.17	N/A	Wrenn and Hart, 1988	>10	58	95
dsdp	26	255	drill	indian ocean (east)	93.44	-31.08	1144	Davies et al., 1974	N/A	N/A	barren
dsdp	28	264	drill	offshore sw australia	112.03	-34.58	2873	Hayes et al., 1975	N/A	N/A	barren
dsdp	28	274	drill	ross sea (nw)	173.26	-68.60	3305	Hayes et al., 1975; Kemp, 1975	N/A	N/A	barren
dsdp	29	281	drill	tasmania	147.46	-47.60	1591	Kennett and Houtz, 1975; Haskell and Wilson (p. 723); Perch-Nielsen (p. 875); Crouch & Hollis 1996	N/A	N/A	barren
dsdp	29	282	drill	tasmania	143.29	-42.15	4202	Kennett and Houtz, 1975; Haskell and Wilson (p. 723)	N/A	N/A	barren
dsdp	30	286	drill	south new hebrides trench	166.22	-16.32	4465	Andrews et al., 1975	N/A	N/A	barren
dsdp	30	289	drill	Ontong-Java Plateau	158.30	0.29	2206	Andrews et al., 1975	N/A	N/A	barren
dsdp	36	328	drill	east of falkland plateau	-36.40	-49.49	5095	Barker et al., 1976; Gombos (p. 575); Harris (p. 761)	N/A	N/A	barren
dsdp	40	360	drill	off south africa/agulhas	18.06	-35.51	2949	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
dsdp	40	361	drill	off south africa/agulhas	15.27	-35.04	4549	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
dsdp	90	593	drill	tasman sea	167.40	-40.30	1068	Kennett et al., 1986	N/A	N/A	barren
odp	113	689	drill	maud rise	3.06	-64.31	2080	Barker et al., 1988/1990; Florindo and Roberts, 2005;	N/A	N/A	barren
odp	113	690	drill	maud rise	1.12	-65.10	2914	Barker et al., 1988/1990; Florindo and Roberts, 2005;	N/A	N/A	barren
odp	114	698	drill	georgia rise	-33.06	-51.28	2138	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	699	drill	georgia rise	-30.41	-51.33	3706	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	701	drill	mid atlantic ridge	-23.13	-51.59	4637	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	702	drill	mid atlantic ridge	-26.22	-50.57	3083	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	114	703	drill	meteor rise	7.54	-47.03	1796	Ciesielski et al., 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	119	738	drill	kerguelen plateau	82.47	-62.43	2253	Barron et al., 1989	N/A	N/A	barren
odp	120	747	drill	kerguelen plateau	76.48	-54.49	1695	Wise et al., 1989/1992	N/A	N/A	barren
odp	120	749	drill	kerguelen plateau	76.24	-58.43	1071	Wise et al., 1989/1992	N/A	N/A	barren
odp	182	1126	drill	eyre terrace	128.04	-33.31	784	Feary et al., 2000; Li et al., 2003a, 2003b	N/A	N/A	barren
odp	182	1128	drill	great australian bight	127.70	-34.50	3874	Feary et al., 2000; Li et al., 2003a, 2003b	N/A	N/A	barren
odp	183	1135	drill	kerguelen plateau (s)	84.16	-59.42	1567	Coffin et al., 2000; 2003	N/A	N/A	barren
odp	183	1136	drill	kerguelen plateau (s)	84.50	-59.39	1930	Coffin et al., 2000; 2003	N/A	N/A	barren
odp	183	1138	drill	kerguelen plateau	75.59	-53.33	1141	Coffin et al., 2000; 2003; Bohaty and Zachos, 2003	N/A	N/A	barren

Bartonian-Priabonian Boundary (38-36 Ma)

ciros	1	CIROS	drill	ross sea	164.30	-77.05	200	Barrett, 1989; Hannah, 1994; 1997; Coccioni and Galeotti, 1997; Hannah et al., 1997; Wilson et al., 1998	10	37	90
dsdp	28	269	drill	antarctic margin	140.04	-61.41	4282	Hayes et al., 1975	>10	37	5
dsdp	28	274	drill	ross sea (nw)	173.26	-68.60	3305	Hayes et al., 1975; Kemp, 1975	>10	40	90
dsdp	29	277	drill	campbell plateau	166.11	-52.13	1214	Unpublished data; Bijl, Houben Sluijs, Brinkhuis, pers. obs.	>10	56	0
dsdp	29	280	drill	tasmania	147.14	-48.57	4176	Kennett and Houtz, 1975; Haskell and Wilson (p. 723); Perch-Nielsen (p. 875); Crouch & Hollis 1996	6	49	90
dsdp	29	281	drill	tasmania	147.46	-47.60	1591	Kennett and Houtz, 1975; Haskell and Wilson (p. 723); Perch-Nielsen (p. 875); Crouch & Hollis 1996	8	44	90
dsdp	29	282	drill	tasmania	143.29	-42.15	4202	Kennett and Houtz, 1975; Haskell and Wilson (p. 723)	>10	40	10
dsdp	29	283	drill	tasman sea	154.17	-43.55	4729	Kennett and Houtz, 1975; Haskell and Wilson (p. 723)	>10	65	80
dsdp	71	511	drill	north of weddell sea	-46.58	-51.00	2589	Ludwig et al., 1983; Goodman & Ford (p. 859); Wise, 1983	3	37	50
odp	113	696	drill	weddell sea	-42.56	-61.51	650	Barker et al., 1988/1990; Mao and Mohr, 1995;	>10	43	90
odp	119	739	drill	prydz bay	75.05	-67.17	412	Barron et al., 1989; Truswell, 1991; Wei and Thierstein, 1991	>10	37	80
odp	119	742	drill	prydz bay	75.24	-67.33	416	Barron et al., 1989; Truswell, 1991	>10	37	90
odp	120	748	drill	kerguelen plateau	78.59	-58.26	1288	Wise et al., 1989/1992; Mohr, 1990; Roberts et al., 2003	>10	92	18
odp	177	1090	drill	agulhas ridge	8.54	-42.55	3710	Gersonde et al., 1999; 2001; Latimer and Filippelli, 2002;	>10	42	0
odp	189	1168	drill	tasman margin (west)	144.25	-42.37	2463	Exon et al., 2001/2004; Sluijs et al., 2003; Brinkhuis et al., 2003b	>10	37	10
odp	189	1170	drill	south tasman rise	146.03	-47.09	2704	Exon et al., 2001/2004; Sluijs et al., 2003	>10	41	90
odp	189	1171	drill	south tasman rise	149.07	-48.30	2150	Exon et al., 2001/2004; Sluijs et al., 2003	>10	54	76
odp	189	1172	drill	east tasman plateau	149.56	-43.58	2622	Exon et al., 2001/2004; Sluijs et al., 2003; Brinkhuis et al., 2003a; Stickley et al., 2004a	>10	66	93
pejerrey	pejerrey	CB	drill	offshore argentina	-55.50	-38.83	N/A	Archangelsky et al.,1997	7	44	5
brown's creek		BC	outcr	south australian margin	138.65	-34.59	N/A	Cookson and Eisenack, 1965; Shafik, 1996; Shafik and Idnurm, 1997	2	37	0
		GB		gippsland basin	136.70	-38.30	N/A	Holdgate et al., 1995; 2000; 2003; Macphail, 2001	10	45	90
		MB	drill	murray basin	141.00	-34.00	N/A	Martin, 1991; 1993	4	37	0
rio turbio		RT	core	southwest santa cruz	-72.30	-51.50	N/A	Archangelsky, 1969; Guerstein, pers. obs.	7	45	35
te kuiti group		TK	outcr	new zealand	174.00	-37.00	N/A	Pocknall, 1991; Edbrooke et al., 1998	6	37	10
		W&H	outcr	seymour island	-56.17	-64.17	N/A	Wrenn and Hart, 1988	>10	58	80
dsdp	26	253	drill	ninetyeast ridge	87.22	-24.53	1962	Davies et al., 1974	N/A	N/A	barren
dsdp	28	264	drill	offshore sw australia	112.03	-34.58	2873	Hayes et al., 1975	N/A	N/A	barren
dsdp	28	267	drill	antarctic margin	104.29	-59.16	4522	Hayes et al., 1975	N/A	N/A	barren
dsdp	29	277	drill	campbell plateau	166.11	-52.13	1214	Kennett and Houtz, 1975; Hollis et al., 1997	N/A	N/A	barren
dsdp	30	286	drill	south new hebrides trench	166.22	-16.32	4465	Andrews et al., 1975	N/A	N/A	barren
dsdp	40	360	drill	off south africa/agulhas	18.06	-35.51	2949	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
dsdp	40	361	drill	off south africa/agulhas	15.27	-35.04	4549	Bolli et al., 1978; McLachlan & Pieterse (p.857)	N/A	N/A	barren
dsdp	90	593	drill	tasman sea	167.40	-40.30	1068	Kennett et al., 1986	N/A	N/A	barren
odp	113	689	drill	maud rise	3.06	-64.31	2080	Barker et al., 1988/1990; Florindo and Roberts, 2005; Diester-Haass and Zahn, 2001	N/A	N/A	barren
odp	113	690	drill	maud rise	1.12	-65.10	2914	Barker et al., 1988/1990; Florindo and Roberts, 2005; Diester-Haass and Zahn, 2001	N/A	N/A	barren
odp	114	699	drill	georgia rise	-30.41	-51.33	3706	Ciesielski et al 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	700	drill	georgia rise	-30.16	-51.31	3601	Ciesielski et al 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	701	drill	mid atlantic ridge	-23.13	-51.59	4637	Ciesielski et al 1988/1991	N/A	N/A	barren
odp	114	702	drill	mid atlantic ridge	-26.22	-50.57	3083	Ciesielski et al 1988/1991	N/A	N/A	barren
odp	114	703	drill	meteor rise	7.54	-47.03	1796	Ciesielski et al 1988/1991; Madile and Monechi, 1991	N/A	N/A	barren
odp	114	704	drill	meteor rise	7.25	-46.52	2532	Ciesielski et al., 1988/1991	N/A	N/A	barren
odp	119	737	drill	kerguelen-heard plateau	73.02	-50.14	564	Barron et al., 1989; Wei and Thierstein, 1991	N/A	N/A	barren
odp	119	738	drill	kerguelen plateau	82.47	-62.43	2253	Barron et al., 1989	N/A	N/A	barren
odp	119	744	drill	kerguelen plateau	80.35	-61.35	2307	Barron et al., 1989; Wei and Thierstein, 1991; Robert et al., 2003	N/A	N/A	barren
odp	120	747	drill	kerguelen plateau	76.48	-54.49	1695	Wise et al., 1989/1992	N/A	N/A	barren
odp	120	749	drill	kerguelen plateau	76.24	-58.43	1071	Wise et al., 1989/1992	N/A	N/A	barren
odp	181	1124	drill	rekohu drift	-176.53	-39.50	3978	Carter et al., 1999; 2002; McGonigal and di Stefano, 2002	N/A	N/A	barren
odp	182	1126	drill	eyre terrace	128.04	-33.31	784	Feary et al., 2000; Li et al., 2003a, 2003b	N/A	N/A	barren
odp	182	1128	drill	great australian bight	127.70	-34.50	3874	Feary et al., 2000; Li et al., 2003a, 2003b	N/A	N/A	barren
odp	182	1134	drill	eyre terrace	127.16	-33.32	701	Feary et al., 2000; Li et al., 2003a, 2003b	N/A	N/A	barren
odp	183	1137	drill	elan bank	68.06	-56.50	1005	Coffin et al., 2000; 2003	N/A	N/A	barren
odp	183	1138	drill	kerguelen plateau	75.59	-53.33	1141	Coffin et al., 2000; 2003; Bohaty and Zachos, 2003	N/A	N/A	barren

Table S2: Dinocyst species

taxon	type	first decibed from	
Achomosphaera ramulifera	cosmopolitan	Europe	
Adnatosphaeridium multispinosum	cosmopolitan	North Sea Basin	
Aiora fenestrata	endemic	South Australian Magin	
Alterbidinium acutulum	endemic SO	New Zealand	
Alterbidinium distinctum	endemic "TF"	McMurdo Sound	
Apectodinium homomorphum	(mid)-low Lat	South Australian Magin	
Apteodinium australiense	mid-low Lat	South Australian Magin	
Arachnodinium antarcticum	endemic "TF"	Antarctica	
Areoligera spp.	mid-low Lat	Europe	
Batiacasphaera cassicula	endemic SO	New Zealand	
Batiacasphaera compta	mid-low Lat	USA	
Batiacasphaera sp. A, B, C, D	endemic SO	East Tasman Plateau	Brinkhuis et al., 2003a, b
Batiacasphaera spp.	mid-low Lat	USA	
Brigantedinium pynei	Antarctic	Ross Sea	
Brigantedinium spp.	mid-low Lat	UK	
Caligodinium spp.	mid-low Lat	USA	
Carpatella spp.	mid-low Lat	Middle Europe	
Cassidium filosum	mid-low Lat	New Zealand	
Cassidium fragile	mid-low Lat	Australia	
Cerebrocysta bartonensis	mid-low Lat	North Sea Basin	
Cerebrocysta sp A, B, C	endemic SO	East Tasman Plateau	Brinkhuis et al., 2003a, b
Cerebrocysta spp.	mid-low Lat	North Sea Basin	
Cerodinium dartmoorium	mid-low Lat	South Australian Magin	
Cerodinium depressum	mid-low Lat	Northwest Europe	
Cerodinium sp. A	endemic SO	East Tasman Plateau	Brinkhuis et al., 2003a
Cerodinium speciosum	mid-low lat	Europe	
Cerodinium striatum	mid-low Lat	USA	
Charlesdowniea coleothrypta	mid-low Lat	UK	
Charlesdowniea edwardsii	mid-low Lat	New Zealand	
Cleistosphaeridium ancyreum	mid-low Lat	South Australian Magin	
Cleistosphaeridium diversispinosum	mid-low Lat	Europe	
Cleistosphaeridium placacanthum	mid-low Lat	South Australian Magin	
Cleistosphaeridium spp.	mid-low Lat	Middle Europe	
Cordosphaeridium fibrospinosum	mid-low Lat	Europe	
Cordosphaeridium minimum	mid-low Lat	Northwest Europe	
Corrudinium incompositum	mid-low Lat	USA	
Corrudinium obscurum	endemic SO	New Zealand	
Corrudinium regulare	endemic "TF"	Falkland Plateau	
Cribroperidinium spp.	cosmopolitan	UK	
Damassadinium crassimuratum	cosmopolitan	New Zealand	
Dapsilidinium spp.	mid-low Lat	North Sea Basin	
Deflandrea antarctica	endemic "TF"	McMurdo Sound	
Deflandrea convexa	endemic "TF"	New Zealand	
Deflandrea cygniformis	endemic "TF"	Southern Argentina	
Deflandrea granulata	mid-low Lat	Southern Argentina	
Deflandrea phosphoritica complex	mid-low Lat	Europe	
Deflandrea scabrata	endemic SO	New Zealand	
Deflandrea sp. A	endemic "TF"	East Tasman Plateau	Brinkhuis et al., 2003a
Dinopterygium sp. A	endemic SO	East Tasman Plateau	Brinkhuis et al., 2003a
Dinopterygium spp.	mid-low Lat	Europe	
Diphyes colligerum	cosmopolitan	South Australian Magin	
Distatodinium paradoxum	mid-low Lat	Europe	
Dracodinium waipawaense	endemic SO	New Zealand	
Elytrocysta brevis (Membranosphaera spp.)	mid-low Lat	North Sea Basin	
Enneadocysta dictyostila	endemic "TF"	McMurdo Sound	
Enneadocysta multicornuta	mid-low Lat	North Sea Basin	
Enneadocysta pectiniformis	mid-low Lat	North Sea Basin	
Eocladopyxis peniculata	mid-low Lat	Northwest Europe	
Gelatia inflata	bipolar	North Sea Basin	
Gippslandia extensa	endemic "TF"	Southeast Australia	
Glaphyrocysta delicata	mid-low Lat	New Zealand	
Glaphyrocysta divaricata	mid-low Lat	North Sea Basin	
Glaphyrocysta intricata	mid-low Lat	North Sea Basin	
Glaphyrocysta ordinata	mid-low Lat	North Sea Basin	
Glaphyrocysta pastielsii	mid-low Lat	South Australian Magin	
Glaphyrocysta retiintexta	mid-low Lat	South Australian Magin	
Hemiplacophora semilunifera	mid-low Lat	South Australian Magin	
Histiocysta spp.	mid-low Lat	Europe, USA	
Homotryblium tasmaniense	mid-low Lat	South Australian Magin	
Hystrichokolpoma bullatum	mid-low Lat	New Zealand	
Hystrichokolpoma pusillum	mid-low Lat	Europe	
Hystrichokolpoma rigaudiae	mid-low Lat	South Australian Magin	
Hystrichokolpoma spinosum	mid-low Lat	New Zealand	
Hystrichokolpoma truncatum	mid-low Lat	Europe	
Hystrichosphaeridium truswelliae	endemic "TF"	Seymour Island	
Hystrichosphaeridium tubiferum	mid-low Lat	Middle Europe	
Hystrichosphaeropsis spp.	mid-low Lat	Europe	
Hystrichostrogylon membraniphora	cosmopolitan	Europe	
Impagidinium aculeatum	mid-low Lat	Australia	
Impagidinium cassiculum	mid-low Lat	New Zealand	
Impagidinium crassimuratum	mid-low Lat	New Zealand	
Impagidinium dispertitum	mid-low Lat	South Australian Magin	
Impagidinium maculatum	mid-low Lat	South Australian Magin	
Impagidinium parvireticulatum	mid-low Lat	New Zealand	
Impagidinium velorum	mid-low Lat	North Sea Basin	
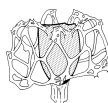
Impagidinium victorianum	mid-low Lat	South Australian Magin	
Impletosphaeridium clavus	endemic SO	Seymour Island	
Impletosphaeridium spp.	mid-low Lat	Northwest Europe	
Lejeunecysta cowiei	Antarctic	Ross Sea	
Lejeunecysta fallax	mid-low Lat	Northwest Europe	
Lejeunecysta hyalina	mid-low Lat	North Sea Basin	
Lejeunecysta spp.	mid-low Lat	Europe	
Lingulodinium bergmannii	mid-low Lat	Argentina	
Lingulodinium machaerophorum	cosmopolitan	South Australian Magin	
Lophocysta spp.	mid-low Lat	Norhtern Europe	
Membranophoridium perforatum	endemic SO	New Zealand	
Moria zachosii	endemic "TF"	East Tasman Plateau	
Nematosphaeropsis spp. (most)	cosmopolitan	South Australian Magin	
Octodinium askiniae	endemic "TF"	Seymour Island	
Operculodinium centrocarpum complex	cosmopolitan	South Australian Magin	
Operculodinium sp. A	endemic SO	East Tasman Plateau	Brinkhuis et al., 2003a
Operculodinium sp. B	endemic SO	East Tasman Plateau	Brinkhuis et al., 2003a
Palaeocystodinium australinum	mid-low Lat	South Australian Magin	
Palaeocystodinium golzowense	mid-low Lat	North Sea Basin	
Palaeoperidinium pyrophorum	mid-low Lat	Middle Europe	
Palaeotetradinium minisculum	mid-low Lat	North Sea Basin	
Paucisphaeridium sp. A	cosmopolitan	East Tasman Plateau	new species
Paucisphaeridium inversibuccinum	mid-low Lat	Europe	
Pentadinium laticinctum	mid-low Lat	North Sea Basin	
Phelodinium cranwelliae	Antarctic	Ross Sea	
Phelodinium harringtonnii	Antarctic	McMurdo Sound	
Phthanoperidinium alectrolophum	mid-low Lat	North Sea Basin	
Phthanoperidinium antarcticum	endemic "TF"	Scotia Sea, Antarctica	
Phthanoperidinium comatum	mid-low Lat	Northwest Europe	
Phthanoperidinium crenulatum	mid-low Lat	Europe	
Phthanoperidinium sp. A	endemic "TF"	Falkland Plateau	new species
Phthanoperidinium sp. B	endemic "TF"	Falkland Plateau	new species
Phthanoperidinium sp. C	endemic "TF"	Kerguelen, Antarctica	new species
Phthanoperidinium stockmansii	bipolar	Southern England	
Polysphaeridium subtile	mid-low Lat	Europe	
Pyxidinopsis spp. (most)	mid-low Lat	Northern Europe	
Pyxidinopsis waipawaensis group	endemic SO	New Zealand	
Rhombodinium rhomboideum	mid-low Lat	North Sea Basin	
Saeptodinium tasmaniense	endemic "TF"	Southeast Australia	
Samlandia chlamydophora	mid-low Lat	Europe	
Samlandia delicata	mid-low Lat	New Zealand	
Samlandia septata	mid-low Lat	New Zealand	
Schematophora obscura	endemic SO	New Zealand	
Schematophora speciosa	mid-low Lat	South Australian Magin	
Selenopemphix nephroides	mid-low Lat	Europe	
Selenopemphix prionata	Antarctic	McMurdo Sound	
Selenopemphix selenoides	mid-low Lat	Europe	
Senegalinium asymmetricum	mid-low Lat	McMurdo Sound	
Senegalinium bicavatum	mid-low Lat	Africa	
Senegalinium dilwynense	mid-low Lat	South Australian Magin	
Senegalinium spp. (pars)	mid-low Lat	Africa	
Spinidinium colemanii	mid-low Lat	Seymour Island	
Spinidinium luciae	endemic "TF"	Seymour Island	
Spinidinium macmurdoense	bipolar	McMurdo Sound	
Spinidinium schellenbergii	endemic "TF"	Tasman region	Sluijs et al., 2009b
Spiniferites ramosus complex	cosmopolitan	Middle Europe	
Spiniferites sp. A Brinkhuis et al., 2003a	endemic SO	East Tasman Plateau	Brinkhuis et al., 2003a
Stoveracysta kakanuiensis	endemic SO	New Zealand	
Stoveracysta ornata	cosmopolitan	South Australian Magin	
Tanyosphaeridium spp.	cosmopolitan	Europe	
Tectatodinium pellitum	mid-low Lat	Australia	
Thalassiphora delicata	mid-low Lat	North Sea Basin	
Thalassiphora pelagica	cosmopolitan	Europe	
Turbiosphaera filosa	mid-low Lat	McMurdo Sound	
Turbiosphaera sagena	mid-low Lat	McMurdo Sound	
Vozzhennikovia apertura	bipolar	McMurdo Sound	Levy and Harwood, 2000
Vozzhennikovia minuta	endemic SO	Weddell Sea	Sluijs et al., 2009b
Vozzhennikovia netrona	endemic "TF"	McMurdo Sound	
Vozzhennikovia roehliae	endemic "TF"	Tasman region	Sluijs et al., 2009b
Vozzhennikovia rotunda	bipolar	McMurdo Sound	
Vozzhennikovia stickleyae	endemic "TF"	Tasman region	Sluijs et al., 2009b
Wetzeliella articulata	mid-low Lat	Europe	
Wetzeliella astra	mid-low Lat	Europe	
Wetzeliella samlandica	mid-low Lat	Europe	
Wilsonidium echinosuturatum	endemic "TF"	New Zealand	
Wilsonidium ornatum	mid-low Lat	New Zealand	
Ynezidinium waipawaense	endemic SO	New Zealand	

Taxonomy based on those cited in Fensome et al., 2004, unless stated otherwise

Legend:	
Mid-low Lat	Species that originate from middle and low latitudes
Cosmopolitan	Species not particularly restricted to any latitude
Endemic SO	Species that are endemic to the Southern Ocean, but most likely represent regional dimorphism within dinoflagellate lineages
Endemic "TF"	Species included in the transantarctic flora
Antarctic	Dinoflagellate cyst species with heterotrophic affinities, and likely associated with sea ice conditions around Antarctica. This group replaces the transantarctic flora in the early Oligocene at numerous places.

CHAPTER 4

Southern Ocean warming, sea level and hydrological change during the Paleocene-Eocene Thermal Maximum

A brief (~150 kyr) period of global average surface warming marks the transition between the Paleocene and Eocene Epochs, ~56 million years ago. This so-called "Paleocene-Eocene Thermal Maximum" (PETM) is associated with the massive injection of ^{13}C-depleted carbon, reflected in a negative carbon isotope excursion (CIE). Biotic responses include a global abundance peak (acme) of the subtropical dinoflagellate *Apectodinium*. Here we identify the PETM in a marine sedimentary sequence deposited on the East Tasman Plateau at Ocean Drilling Program (ODP) Site 1172 and show, based on the organic paleothermometer TEX_{86}, that southwest Pacific sea surface temperatures increased from ~26°C to ~33°C during the PETM. Such temperatures before, during and after the PETM are >10°C warmer than predicted by paleoclimate model simulations for this latitude. In part, this discrepancy may be explained by potential seasonal biases in the TEX_{86} proxy in polar oceans. Additionally, the data suggest that not only Arctic, but also Antarctic temperatures may be underestimated in simulations of ancient greenhouse climates by current generation fully coupled climate models. An early influx of abundant *Apectodinium* confirms that environmental change preceded the CIE on a global scale. Organic dinoflagellate cyst assemblages suggest a local decrease in the amount of river run off reaching the core site during the PETM, possibly in concert with eustatic rise. Moreover, the assemblages suggest changes in seasonality of the regional hydrological system and storm activity. Finally, significant variation in dinoflagellate cyst assemblages during the PETM indicates that southwest Pacific climates varied significantly over time scales of $10^3 – 10^4$ years during this event, a finding comparable to similar studies of PETM successions from the New Jersey Shelf.

1. Introduction

Gradual widespread warming initiated in the late Paleocene (~59 Ma) and culminated in the Early Eocene Climatic Optimum (EECO; 52–50 Ma; e.g., Chapter 2; Adams et al., 1990; Zachos et al., 2001). Superimposed on this long-term warming trend, at least four "hyperthermals" occurred, which represent relatively brief (<200 kyr) intervals characterized by anomalously high temperatures (e.g., Bowen et al., 2006; Sluijs et al., 2007a). The Paleocene-Eocene Thermal Maximum (PETM) is the most prominent and best-studied hyperthermal and is marked by a negative carbon isotope excursion (CIE) in sedimentary components of 2.5–8‰, depending on analyzed substrate, location and completeness of the section (Kennett and Stott 1991; Koch et al., 1992; Schouten et al., 2007c). Moreover, massive dissolution of biogenic carbonates occurred in deep ocean basins (e.g., Zachos et al., 2005; Zeebe and Zachos 2007). The CIE and carbonate dissolution are consistent with geologically rapid, massive injections of ^{13}C-depleted carbon into the ocean-atmosphere system (Dickens et al., 1997; Panchuk et al., 2008; Zeebe et al., 2009), although the mechanism for such release remains controversial (Dickens et al., 1995; Kurtz et al., 2003; Svensen et al., 2004).

Stable oxygen isotope ($\delta^{18}O$) and Mg/Ca studies on planktonic foraminifera from deep-sea sediments indicate a 5–8°C surface warming during the PETM (Kennett and Stott 1991; Thomas et al., 2002; Zachos et al., 2003). Reconstruction of absolute sea surface temperatures (SST) from such sections has been problematic because of post-sedimentary recrystallization of planktonic foraminifera (Pearson et al., 2001; Schrag et al., 1995). Additionally, reduced pH may have dampened foraminifer $\delta^{18}O$ excursions, potentially resulting in too low estimates of PETM warming (Uchikawa and Zeebe 2010). More recently, the application of organic paleothermometers, such as TEX_{86} and MBT/CBT in marginal marine sequences provided estimates of absolute temperature evolution across the PETM and other hyperthermals in the Northern Hemisphere (e.g., Sluijs et al., 2006; Weijers et al., 2007a; Zachos et al., 2006). This work showed exceptionally high Arctic temperatures during this time interval, suggesting very low meridional temperature gradients (Sluijs et al., 2006). The marginal marine sections used in these studies have also revealed significant increases in river discharge and sediment input (e.g., Crouch et al., 2003; Giusberti et al., 2007; John et al., 2008; Sluijs et al., 2008a; for an overview see Sluijs et al., 2008b), changes in trophic level (e.g., Crouch et al., 2001; Gibbs et al., 2006; Speijer and Wagner 2002), as well as a globally recorded rise in sea level (Sluijs et al., 2008b). However, temperature and paleo-ecological data from marginal marine PETM sections from the Southern Hemisphere are rare (Crouch and Brinkhuis 2005; Crouch et al., 2001), and none are available from the southern high-latitudes, hampering thorough evaluation of climatic change in the sub-Antarctic realm.

We have generated geochemical and palynological data through upper Paleocene – lower Eocene sediments recovered during Ocean Drilling Program (ODP) Leg 189 at Site 1172 on the East Tasman Plateau, deposited at ~65°S paleolatitude (Exon et al., 2004; Fig. 1). Micropaleontological information from the southwest Pacific showed that this site was located within the Antarctic-derived, northward

flowing Tasman Current throughout the Paleogene, which is consistent with general circulation model experiments (Hollis et al., 2009; Huber et al., 2004). In an earlier study, based on initial shipboard samples we suggested that the PETM might not have been recovered at this site (Röhl et al., 2004b). Here, we identify a condensed PETM section on the basis of a negative CIE in organic matter within Subchron C24r. We perform TEX_{86}, dinoflagellate cyst assemblage analyses and X-ray fluorescence (XRF) core scanning in order to reconstruct paleoenvironmental conditions at southern high-latitudes across the PETM.
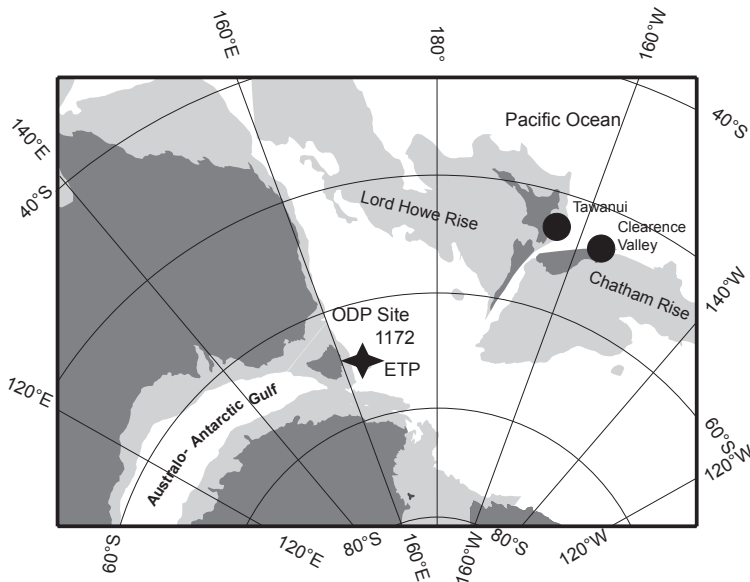
Figure 1. Tectonic reconstruction of the southwest Pacific region for the earliest Eocene, with East Antarctica held fixed, illustrating modern continents (dark grey), areas shallower than 3000m (light grey) and locations of ODP Site 1172 and other sites discussed in the text. ETP = East Tasman Plateau, STR = South Tasman Rise. The figure is modified from Cande and Stock (2004).

2. Material and methods

2.1. Material

Sediments of late Paleocene and early Eocene age at ODP Site 1172, Hole 1172D, consist mainly of organic-rich green and grey clay- and siltstones with low abundance of calcareous and siliceous microfossils, but high abundance of palynomorphs (notably dinocysts but also terrestrial pollen and spores). Glauconite and accessory minerals are recorded in varying abundance (Exon et al., 2001), with the glauconite grains being irregular and angular, which indicates that glauconite was formed in situ (based on thin sections, data not shown). Lithological and palynological information suggests an overall very shallow marine depositional setting with marked runoff from the nearby shores (Exon et al., 2001; Röhl et al., 2004b). Integrated dinoflagellate cyst and magnetostratigraphic studies identified Subchrons C25n, C24r

and C24n, with the top of Subchron C25n at 618.00 rmbsf and the onset of Subchron C24n at 594.2 rmbsf (Fuller and Touchard 2004; Stickley et al., 2004a). Average sedimentation rates implied by this age model are 5.7 m/Myr, when assuming a duration of 3.1 Ma for Subchron C24r (Westerhold et al., 2007).
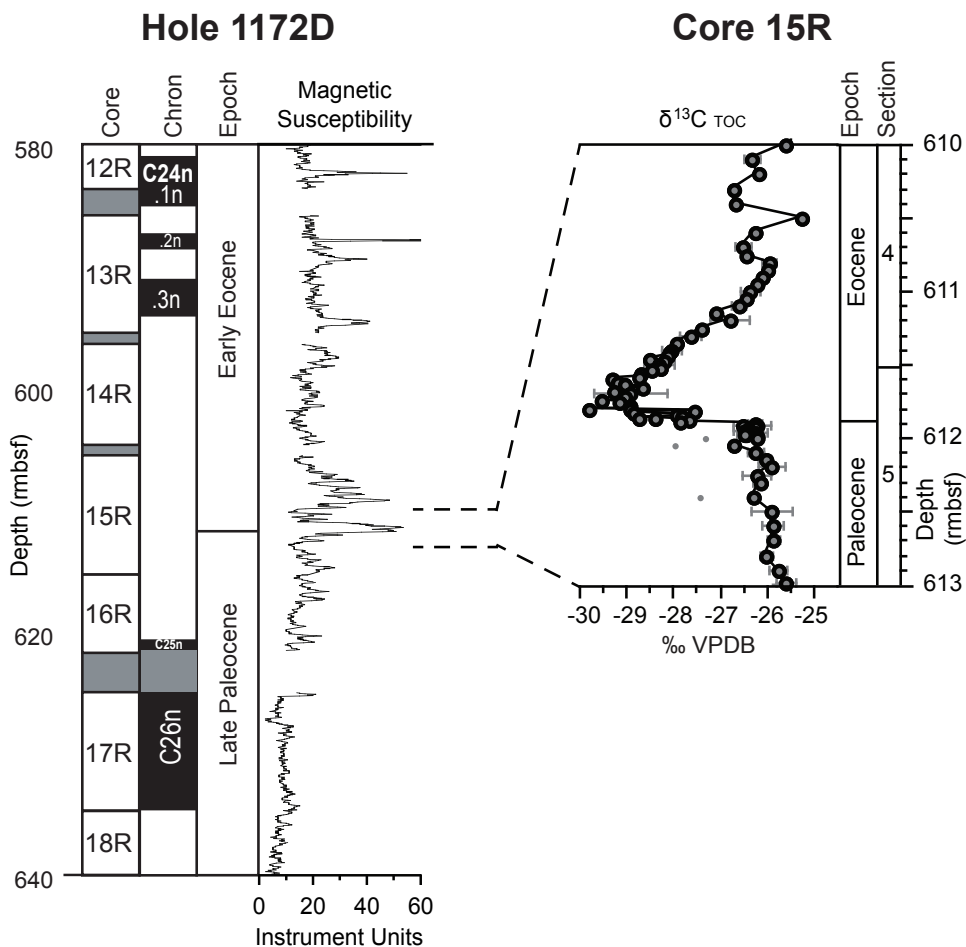

Figure 2. Core recovery and stratigraphic summary of the latest Paleocene to earliest Eocene at ODP Site 1172. Note that the core depths are in revised meters below sea floor (rmbsf; see text). Magnetostratigraphic interpretation is from Chapter 1, which differs from the interpretations published in the first post-cruise papers (Fuller and Touchard 2004; Röhl et al., 2004b) in the assignment of normal polarity intervals to Subchrons C25n and C26n. Magnetic susceptibility record represents shipboard data (Exon et al., 2001). Stable carbon isotope ($\delta^{13}C$) values of total organic carbon (TOC) across the PETM are expressed relative to the Vienna Pee Dee Belemnite standard. Error bars reflect duplicate based standard deviations and three grey data points are considered outliers because duplicate analyses indicated values consistent with surrounding samples.

2.2. Methods

The archive halves of Core 189-1172D-15R were subject to XRF Core Scanning. Subsequently, half-splits of these archive halves were sampled on a resolution of 1 to 2 cm, after which samples were freeze-dried. Splits of samples were then taken for palynology and organic geochemistry.

2.2.1. X-ray fluorescence (XRF) core scanning

We measured the elemental composition of sediments from Cores ODP Leg 189 Site 1172D-15R to -17R at the MARUM, Bremen University, Germany, using the XRF core scanner (Richter et al., 2006; Tjallingii et al., 2007). The XRF core scanner acquires bulk-sediment chemical data from split core surfaces. Although measured elemental intensities are predominantly proportional to concentration, they are also influenced by the energy level of the X-ray source, count time, and physical properties of the sediment (Röhl and Abrams 2000; Tjallingii et al., 2007). XRF data were collected every cm down-core over a 1 cm^2 area using 30 s count time. We used a generator setting of 20 kV and an X-ray current of 0.15 mA.

2.2.2. Palynology

Samples of 1–2 cm stratigraphic thickness were freeze-dried and a known amount of *Lycopodium* spores was added to ~4 g of material. Samples were then treated with 30% HCl and twice with 38% HF for carbonate and silicate removal, respectively. Residues were sieved using a 15-µm nylon mesh to remove small particles. To break up clumps of residue, the sample was placed in an ultrasonic bath for a maximum of 5 min, sieved again, and subsequently concentrated into 1 ml of glycerine water, of which 10 µl was mounted on microscope slides. Slides were counted for marine (e.g., dinocysts) and terrestrial palynomorphs (e.g., pollen and spores) to a minimum of 200 dinocysts. Marine and terrestrial palynomorph preservation was excellent for all samples. We generally follow dinocyst taxonomy of Fensome and Williams (2004), but follow Sluijs et al. (2009b) for the various spiny peridinioid taxa. Absolute quantitative numbers were counted using the relative abundance of *Lycopodium* spores (Stockmarr 1972).

2.2.3. Organic geochemistry

For stable carbon isotope analyses of total organic carbon ($\delta^{13}C_{TOC}$), freeze-dried samples were powdered, treated with 1M HCl to remove carbonate, centrifuged and the supernatant decanted, followed by two rinses with demineralised water and freeze-dried again. Residues were analyzed with a Fison NA 1500 CNS analyzer coupled to a Finnigan Delta Plus isotope ration mass spectrometer. Analytical precision and accuracy were determined by replicate analyses and by comparison with in-house standards, and were better than 0.1‰ and 0.1‰, respectively. For biomarker analyses, freeze-dried sediment samples (~3.5 g dry mass) were extracted with dichloromethane (DCM)/methanol (2:1) using accelerated solvent extraction (Dionex ACE). The extracts were separated by Al_2O_3 column chromatography using hexane/DCM (9:1, v/v) and DCM/methanol (1:1, v/v) to yield the apolar and polar fractions, respectively. The polar fractions were analyzed using high performance liquid chromatography/atmospheric pressure chemical ionization-mass spectrometry, according to Schouten et al. (2007a). Single ion monitoring was used to

quantify the abundance of the Glycerol Dialkyl Glycerol Tetraether (GDGT) lipids. The relative abundance of GDGTs was used to calculate TEX_{86} (Schouten et al., 2002). TEX_{86} is converted to mean annual SST by means of quasi-global core top calibrations. A new calibration with a logarithmic function was recently published (Kim et al., 2010), which is based on the currently available core-top data and thorough statistical analyses between GDGTs abundances and SST. An earlier calibration assumes a different logarithmic relation (Liu et al., 2009) that produces particularly different SSTs for high TEX_{86} values. We also determined the Branched and Isoprenoid Tetraether (BIT) index. This is a ratio between soil bacteria-derived and marine crenarchaeota-derived membrane lipids, and serves as a proxy for the relative amount of river transported soil organic matter versus marine organic matter (Hopmans et al., 2004).

2.2.4. Core depth adjustments

Based on correlations between physical properties data generated on core material and down hole logging, we have slightly changed the meters below sea floor (mbsf) depth of the core sediments in a recent paper (Chapter 2). We use revised mbsf (rmbsf) for these revised depths throughout. Relative to mbsf, Core 189-1172D-12R was shifted up by 0.36 m, -13R down by 1.87m, -14R down by 2.84 m, -15R down by 2.4 m, -16R down by 2.57m and -17R and -18R down by 2.66m.
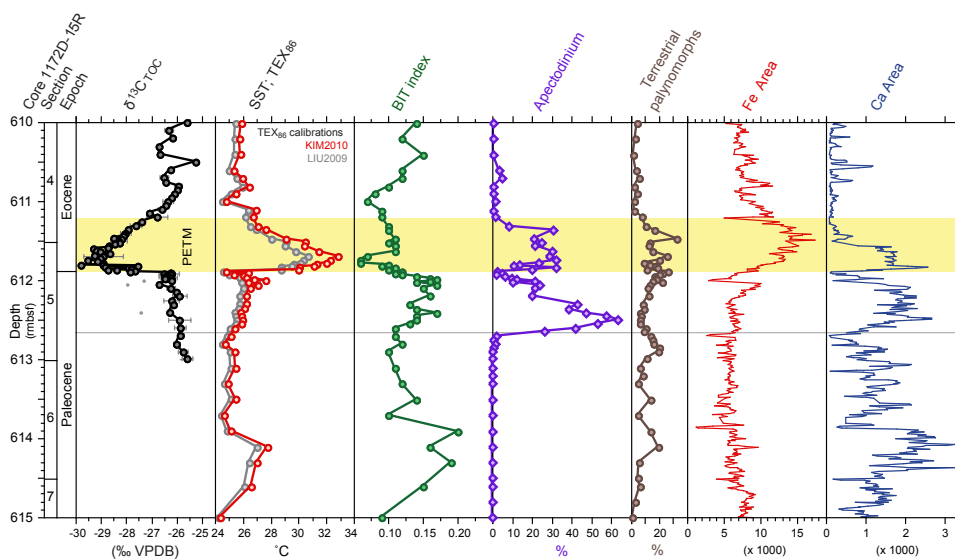

Figure 3. Organic geochemical, palynological and XRF results across the PETM of ODP Site 1172. From left to right: stable carbon isotope ($\delta^{13}C$) values of total organic carbon (TOC) (see caption at Fig. 2 for details); sea surface temperature (SST) based on TEX_{86} following the calibrations KIM2010 (Kim et al., 2010) and LIU2009 (Liu et al., 2009); BIT index; *Apectodinium* percentage of the dinocyst assemblage; abundance of terrestrially derived palynomorphs as a percentage of total pollen and dinocysts; and XRF intensity data for iron (Fe) and calcium (Ca).

3 Results

3.1. Stratigraphy

Between 611.89 and 611.86 rmbsf, the $\delta^{13}C_{TOC}$ curve shows a ~3‰ negative step from −26 to −29‰, followed by a ~20 cm interval of relatively stable values, and a subsequent exponential recovery that reaches background values between 611.2 and 611.0 rmbsf (Fig. 2). This excursion is located within Subchron C24r. Average sedimentation rates of 5.7 m/Myrs for this Subchron (see Material) imply that this excursion occurred ~1 Myr after the termination of Subchron C25n (Fuller and Touchard 2004; Chapter 2) and ~2 Myrs between this CIE and the onset of Subchron C24n. The orbitally based age model from Blake Nose (ODP Leg 171B) and Walvis Ridge (ODP Leg 208) also indicates ~1 Myr between the top of Subchron C25n and the PETM (Norris and Röhl 1999), and ~2 Myrs between the PETM and the onset of Subchron C24n (Westerhold et al., 2007). The onset of Eocene Thermal Maximum 2 occurred ~150 kyrs prior to the reversal of Subchron C24r to C24n.3n, which is inconsistent with the location of the recorded CIE. Rather, the overall stratigraphic position of the CIE implies the presence of the PETM in Core 1172D-15R. The thickness of the CIE at ODP Site 1172 is 65–90 cm, depending on the definition of its termination (Röhl et al., 2007). Assuming a 170 kyr duration of the CIE (Abdul Aziz et al., 2008; Röhl et al., 2007), this indicates average PETM sedimentation rates of ~0.4–0.5 cm/kyr, although sediment accumulation rates were likely highly variable in this pro-deltaic setting.

3.2. TEX$_{86}$ and BIT

Late Paleocene SSTs average ~26°C (1σ = 0.9) based on TEX$_{86}$, regardless of the applied calibration (Fig. 2). Concomitantly with the onset of the CIE, SSTs rise to average PETM values of ~31°C (1σ = 0.7) following KIM2010, or ~29°C for LIU2009 with peak values of almost 33°C and 31°C for the two calibrations, respectively, at 611.70 rmbsf. The magnitude of PETM warming was thus ~7°C with KIM2010 and 4°C with LIU2009. SSTs returned to pre-excursion values during the recovery of the CIE. The warming and the CIE are preceded by two samples with relatively low temperatures (~25°C). BIT values are low throughout the analyzed interval, indicating that TEX$_{86}$ values are not influenced by soil derived GDGTs (Weijers et al., 2006). The BIT record exhibits some scatter, but values during the PETM (0.09±0.02) are on average somewhat lower than before and after the PETM (0.13±0.03), suggesting increased marine production of isoprenoid GDGTs or a decreased supply of soil organic matter.

3.3. Palynology

Palynological assemblages are rich, well preserved and dominated by dinoflagellate cysts (dinocysts). Terrestrial pollen and spores are common to abundant throughout, with relatively high abundances within the PETM (Figs. 2, 3). Stratigraphically important dinocyst taxa include *Apectodinium* spp., *Eocladopyxis peniculata*, *Deflandrea* spp., *Melitasphaeridium pseudorecurvatum*, *Muratodinium fimbriatum* and the recently described species *Florentinia reichartii* (Sluijs and Brinkhuis 2009). In particular, the oldest abundant occurrence (>40% of the

assemblage) of *Apectodinium* in the southwest Pacific Ocean has been calibrated to the PETM (Crouch et al., 2001). At ODP Site 1172 however, the First Occurrence (FO) of abundant *Apectodinium* is at ~612.6 rmbsf, ~75 cm below the onset of the CIE (Fig. 3). *Apectodinium* abundances subsequently decrease to ~2%, followed by a second abundance maximum starting at the onset of the CIE. Along with *Apectodinium* spp., other quantitatively significant taxa in the assemblage mostly comprise cosmopolitan taxa such as *Senegalinium* spp., *Glaphyrocysta* spp., *Eocladopyxis peniculatum*, *Cordosphaeridium fibrospinosum*, *Thalassiphora* spp., *Kenleyia* spp., *Fibrocysta* spp. (and other members of the *Cordosphaeridium fibrospinosum* complex sensu Sluijs and Brinkhuis (2009)), *Diphyes colligerum*, *Paucisphaeridium*, *Deflandrea* (and a few related *Cerodinium*), *Membranosphaera* (often referred to as *Elytrocysta* in the Southern Ocean), and *Spiniferites* spp. *Hystrichosphaeridium truswelliae*, common in certain intervals, was long thought to have been endemic to the Antarctic realm, but was recently recorded in uppermost Paleocene and PETM sediments on the New Jersey Shelf (Sluijs and Brinkhuis 2009). In fact, PETM assemblages as a whole are strikingly similar to those reported from the New Jersey Shelf. Only few aspects of the assemblages are typical for the Antarctic Realm (e.g., Warnaar et al., 2009; Wrenn and Beckman 1982), including rare *Vozzhennikovia* spp., and temporally abundant *Pyxidinopsis* spp. *Senegalinium* spp. dominate assemblages from the base of the studied section (~615 rmbsf) up to ~613 rmbsf, an interval with very stable dinocyst assemblages with common *Pyxidinopsis*, *Spiniferites*, and *Deflandrea* spp. Assemblages are slightly richer above 613 rmbsf, with more abundant *Pyxidinopsis* spp. and common *H. truswelliae* and *Membranosphaera* spp. A peak in *Glaphyrocysta* spp. occurs around 613 rmbsf, directly followed by the first *Apectodinium* acme. Between 612.2 and 611.9 rmbsf, just below the onset of the CIE, successive transient abundances of *Glaphyrocysta*, *Deflandrea*, *Pyxidinopsis*, and *Operculodinium* spp., and *C. fibrospinosum* complex occur. A second acme of *Apectodinium* is recorded concomitant with the CIE. Within the CIE, transient abundances of *Glaphyrocysta* spp. and *Eocladopyxis peniculatum* occur. After the CIE, *Senegalinium* dominates assemblages again, while *Operculodinium* spp., *H. truswelliae* and *Membranosphaera* spp. are common.

3.4. XRF

Fe and Ca intensities exhibit a characteristic variability that can be directly attributed to lithology (Fig. 2). The sediments at this site are composed of clay- and siltstones with low abundance of $CaCO_3$ (<0.3%), between ~0.5–1% TOC, pyritized diatoms, glauconite, accessory minerals in changing abundance, including varying amounts of quartz (Exon et al., 2001). The dominant lithology is expressed as generally low Ca values in the XRF scans, but the Ca in these sediments is related to carbonate (Röhl et al., 2004a, b). Ca values are higher in the lower part of Core -15R, just three meters below the PETM (Fig. 2). These relatively higher carbonate contents, reflecting higher abundance of nannofossils, are in line with a shallow marine environment, but compared to the sediments below and above a relatively deeper depositional environment. The Ca intensities in this interval of Core -15R show regular fluctuations: about four cycles below the onset of the PETM, which may also be present in the BIT index, the TEX_{86} sea surface temperature data, and reversely in the percentage of terrestrial palynomorphs. Assuming an average sedimentation rate

of 5.7 m/Myrs these cycles could represent the low eccentricity frequency of the Milankovitch orbital band (100-kyr cycles). Ca values exhibit peak values during the warming of the PETM, followed by the lowest Ca values in the interval (611.56–610.57 rmbsf) (Fig. 2). Ca and Fe are often closely anticorrelated in the pelagic realm, e.g., at ODP Sites 690 and 1263, because abundances of these elements are both forced by carbonate export and preservation (Röhl et al., 2007). Although the Fe and Ca intensities are overall anti-correlated during the PETM at ODP Site 1172 (611.2–611.9 rmbsf) their relation is not as strong as in the deep sea. The Fe record exhibits maximum values (broad peak) in the upper part of the CIE, where the Ca values are lower and the terrestrial palynomorphs show maximum values. This indicates that the abundances of these elements are not controlled by carbonate dissolution, as expected for this shallow marine setting (Röhl et al., 2004a, b), but rather by the sediment supply from land and marine biogeochemical cycling. Indeed, it suggests that the relation of these two main elements Ca and Fe is driven by minor (Milankovitch-driven) variations in clay mineralogy in combination with very low carbonate contents. Ca values stay low above the PETM. In general, for the long-term Site 1172 record including Core -15R, Ca and Fe do not particularly anti-correlate on a larger scale. This is partly caused by the relatively high amount of silica in the sediments in addition to carbonate (Ca) and clay (Fe), which disturbed the perhaps expected anti-correlation of Ca and Fe.
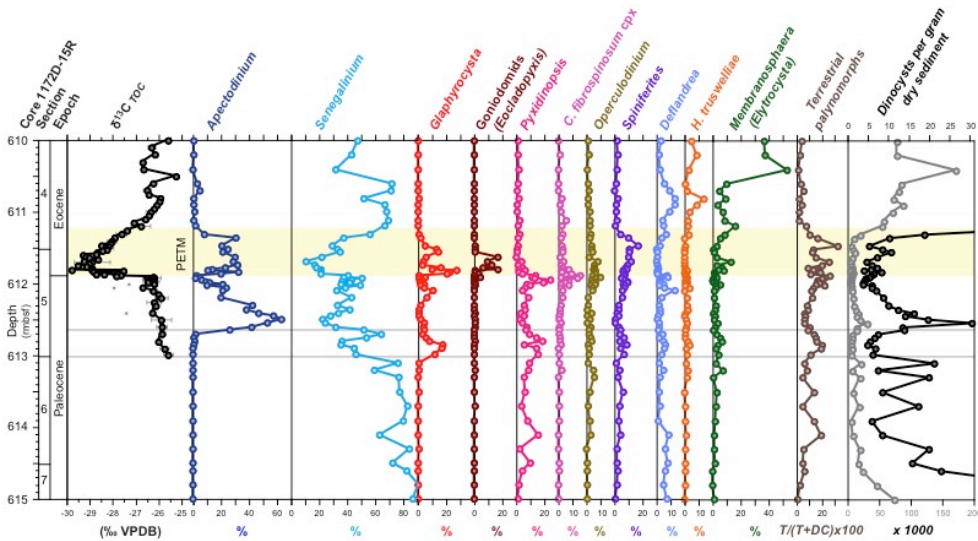

Figure 4. Palynological results across the PETM of ODP Site 1172. Dinocyst abundances are reflected as the percentage of the total dinocyst assemblage and the abundance of terrestrially derived palynomorphs as a percentage of total palynomorph sum. Goniodomids almost exclusively represent *Eocladopyxis peniculata*. *Membranosphaera* spp. is often referred to as *Elytrocysta* spp. in Southern Ocean literature. Grey and black lines in the absolute quantitative dinocyst abundance panel reflect two different scales for illustration purposes.

4 Discussion

4.1. The carbon isotope excursion

The magnitude ($\sim$3‰) of the CIE recorded at ODP Site 1172 is similar to or slightly smaller than the 4–5‰ often recorded in $\delta^{13}C_{TOC}$ at other marine sites (Bolle et al., 2000; Crouch et al., 2001; Kaiho et al., 1996; Sluijs et al., 2006; Steurbaut et al., 2003). Since the lowest $\delta^{13}C$ values for most carbon isotope records across the PETM are located close to the onset of the event (e.g., Bowen et al., 2001; Sluijs et al., 2007a; Thomas et al., 2002), this could imply that the earliest part of the PETM is not represented in our record. Alternatively, the seemingly damped magnitude of the excursion may be caused by changing sources of organic matter as hypothesized for the small CIE at Tawanui, New Zealand (Crouch et al., 2003), but our palynological analyses do not support major changes in organic matter composition (Figs. 2, 3). Globally, most $\delta^{13}C$ curves from the PETM show a rapid drop at the onset, which probably took less than 10,000 years. Subsequently, some bulk carbonate and bulk organic carbon records suggest a slower, sometimes stepwise, continued decline that may span several tens of thousands of years (e.g., Bains et al., 1999; Nicolo et al., 2011). In other records, notably those of single specimen foraminifera, the first negative step is directly followed by $\sim$80 kyr of stable carbon isotope values and subsequent recovery that is also recorded in the bulk records, often referred to as the "body" of the CIE (e.g., McCarren et al., 2009; Thomas et al., 2002). Our record clearly shows an interval of stable values of around −29‰ between $\sim$611.9–611.7 rmbsf, implying that at least part of the stable peak phase is represented in the record. In fact, the magnitude of the CIE in our record is very close to the −3 to −3.5‰ that is generally assumed to have been the excursion in the global exogenic carbon (McCarren et al., 2009; Zachos et al., 2007). Hence, the record at ODP Site 1172 appears to contain at least a large part of the "body" of the CIE as well as the recovery, allowing comparison to other PETM sites.

4.2. Sea surface temperature evolution
4.2.1. Magnitude of PETM warming

The range of SST estimates based on TEX_{86} is slightly different for the two applied calibrations; LIU2009 gives relatively low temperature estimates and implies a very low sensitivity for TEX_{86} values at temperatures >30°C. In contrast, KIM2010 implies a higher sensitivity and absolute temperature estimates (Fig. 2). In the New Jersey PETM records for example, the KIM2010 is most consistent with mixed layer planktonic foraminifer stable oxygen isotope ($\delta^{18}O$) paleothermometry (Kim et al., 2010), also regarding the magnitude of warming. Because the awkwardly low sensitivity of the LIU2009 calibration and decent multi-proxy intercomparison we prefer the magnitude of warming implied by the KIM2010 calibration in this upper range of TEX_{86} values. A warming of $\sim$7°C is similar to or slightly less than the only other Southern Ocean estimates from the Weddell Sea (Sites 689 and 690 at Maud Rise), based on the $\delta^{18}O$ excursion in the surface dwelling foraminifer *Acarinina* (Thomas et al., 2002; Zachos et al., 2007). The magnitude of the SST rise is also similar to that recorded at marginal marine sites on the New Jersey Shelf based on foraminiferal $\delta^{18}O$ and TEX_{86} (John et al., 2008; Sluijs et al., 2007b; Zachos et al., 2006). However, the magnitude of warming was smaller in open-ocean and continental

124

settings, and in the Arctic (e.g., Sluijs et al., 2006; Thomas and Shackleton 1996; Tripati and Elderfield 2005; Weijers et al., 2007a; Wing et al., 2005; Zachos et al., 2003). This suggests that, while the Arctic warmed with a magnitude comparable to the global average (Sluijs et al., 2006), some marginal marine regions warmed slightly more and some polar amplification may have occurred in the Southern Hemisphere. If so, this amplification may have been caused by three mechanisms. First, the melting of small ice sheets on high mountains in Antarctica may have reduced albedo and thus amplified Antarctic warming. This would be consistent with the reconstructed PETM eustatic rise (Sluijs et al., 2008b). Secondly, an increase in atmospheric heat transport may have occurred. Indeed, increased precipitation in Southern Hemisphere PETM records would suggest more latent heat transport from tropical regions to the Antarctic (Crouch et al., 2003; Robert and Kennett 1994). However, Arctic sections also exhibit evidence for intensified regional hydrology (Pagani et al., 2006), but no amplification of warming is recorded there (Sluijs et al., 2006). Third, a change in ocean circulation during the PETM may have resulted in regionally enhanced warming in the southwest Pacific Ocean and Weddell Sea.

4.2.2. Absolute temperatures

Late Paleocene SSTs average ~26℃ for both the LIU2009 and KIM2010 calibrations (Fig. 2). Average PETM SSTs are ~31℃ following KIM2010 and ~29℃ for LIU2009, with peak values of almost 33℃ and 31℃ for the two calibrations, respectively. Because of this calibration difference and as these temperatures are outside the range of modern SSTs, care should be taken in interpreting absolute PETM SST values. Considering the uncertainties of the respective calibrations, TEX_{86} indicates that maximum PETM SSTs at ODP Site 1172 were in the range between 29–34℃. Although TEX_{86} is calibrated to mean annual SST in the modern ocean, like other proxies seasonal and depth biases can occur with the TEX_{86} paleothermometer (Castañeda et al., 2010; Huguet et al., 2007). The marine Crenarchaeota (recently renamed Thaumarchaeota by Spang et al. (2010)) currently mainly proliferate during winter in both the Arctic and Antarctic oceans (Alonso-Saez et al., 2008; Kalanetra et al., 2009). Likely, this is because most of them are chemoautotrophs (Wuchter et al., 2006a), living on ammonia and being not directly depending on light. Also in the present day North Sea they preferentially grow during winter, low light and no competition with algae for ammonia (Wuchter et al., 2006a). Because Crenarchaeota/ Thaumarchaeota have low kinetics for ammonia (Martens-Habbena et al., 2009), they outcompete bacteria and algae at low ammonia concentrations. These microbiological studies would thus imply that the TEX_{86} signal may be even skewed towards winter temperatures. However, the export of membrane lipids to the sea floor is not necessarily a function of Thaumarchaeotal cell abundance but rather of export through faecal pelleting (Wakeham et al., 2003; Wuchter et al., 2006b). Because the dominant season of export in Paleogene polar oceans was likely summer, we previously suggested that TEX_{86} values in such regions might be skewed towards summer temperatures (Sluijs et al., 2006, 2008a). Multi-proxy comparison has indicated good correspondence between TEX_{86} and the molecular MBT/CBT proxy for continental air temperature across the PETM in the Arctic (Weijers et al., 2007a). Although direct comparison to terrestrial reconstructions in the southwest Pacific is complex, our temperatures are qualitatively consistent with dominant angiosperm

pollen in the Site 1172 record. Moreover, records from New Zealand generally exhibit subtropical to tropical floral incursions, including occurrences of the mangrove palm *Nypa* in the Taranaki, East Coast, and Canterbury Basins (e.g., Crouch et al., 2005; Crouch and Visscher 2003). In marine sections from New Zealand, TEX$_{86}$ agrees with foraminiferal stable oxygen isotope and Mg/Ca ratios in the Eocene (Creech et al., 2010; Hollis et al., 2009). Moreover, TEX$_{86}$ and U$_{37}^{K'}$ yield similar SST estimates across the Middle Eocene Climatic Optimum ($\sim$40 Ma) at ODP Site 1172 (Chapter 6). However, it should be noted that seasonal biases cannot be excluded in any of these proxies. Oxygen isotope ratios of reputedly Eocene molluscs from Seymour Island in coastal Antarctica indicate significantly cooler conditions on the Antarctic shelf (Ivany et al., 2008). However, it should be noted that these molluscs are relatively poorly dated and may also be of middle Eocene age. Nevertheless, if the molluscs are early Eocene and their δ^{18}O values yield reliable SST estimates, this might imply that TEX$_{86}$ is skewed towards summer temperatures. Moreover, temperature estimates for deep-sea waters, which at time most likely derived from Antarctic surface waters, are 10–15°C cooler than our SST estimates (McCarren et al., 2009; Thomas and Shackleton 1996; Tripati and Elderfield 2005). Again, part of this discrepancy can result from seasonal biases. The temperature of deep waters most likely represents winter SST along the Antarctic margin. Collectively, although with the available data it remains impossible to exclude that TEX$_{86}$ represents MAT, we consider it likely that the proxy is biased towards summer temperatures in the polar oceans of the Paleogene. Even if TEX$_{86}$ temperatures are skewed towards summer SST estimates, the values are surprisingly high for this latitude, even with the conservative LIU2009 calibration.

Pre-PETM SSTs of $\sim$26–27°C and maximum PETM SSTs of $\sim$30–34°C are only 3–6°C cooler than on the New Jersey Shelf at a paleolatitude of $\sim$35–40°N (Sluijs et al., 2007b; Zachos et al., 2006), and 8–12°C warmer than those in the Arctic (Sluijs et al., 2006) depending on the applied calibration. Even these reflect maximum summer temperature estimates, the difference with late Paleocene and early Eocene SSTs 30–35°C from Tanzania at $\sim$17°S (Pearson et al., 2007) and northern South America at $\sim$10°N (Jaramillo et al., 2010) is extremely small. Originally, the high temperatures in the southern ocean were explained by the supply of warm water through poleward ocean currents (Kennett 1977; Murphy and Kennett 1986), but more recent work has indicated that the East Tasman Plateau was more likely influenced by an Antarctic-derived Tasman Current (Chapter 3; Hollis et al., 2009; Huber et al., 2004). Although the warmest regions have not been sampled yet (Huber 2008), this supports previous observations (Chapter 2; Hollis et al., 2009) of a significantly reduced temperature gradient between the southwest Pacific and low latitudes. As indicated for the Northern Hemisphere data (Sluijs et al., 2007b; Sluijs et al., 2006; Zachos et al., 2006), and then particularly the high winter temperatures (e.g., Sluijs et al., 2009a), the small meridional gradients remain problematic to reconcile with current generation climate models, although recent modeling work has reduced the discrepancy (Abbot et al., 2009; Heinemann et al., 2009). Interestingly, regardless of the calibration, peak PETM SSTs are similar to those recorded for the EECO at ODP Site 1172 (Chapter 2). Unless Earth's surface temperatures were not sensitive to changing greenhouse forcing at this high temperature end, this suggests that atmospheric greenhouse gas levels were comparable during the peak of the PETM and long-term warmth of the EECO. In fact,

although regional differences exist, peak PETM temperatures were similar to those during ETM2 (Sluijs et al., 2009a; Stap et al., 2010). If so, one may speculate that the long-term late Paleocene – early Eocene warming and associated carbon isotope trend that culminated in the EECO as well as the superimposed hyperthermals, were caused by carbon injection (Hancock et al., 2007) from the same reservoir. Such a scenario requires a source that slowly added carbon to the global exogenic pool during the long-term trend resulting in the EECO. During the hyperthermals, it must have released carbon catastrophically, perhaps when an orbital (Lourens et al., 2005) threshold was surpassed, followed by partial recharge. One reservoir that may behave like this is the methane hydrate reservoir (Dickens 2003). Several potential problems exist with methane hydrates as the only source of ^{13}C-depleted carbon, such as the volume and residence time of this reservoir during the Paleogene. However, a long-term net leakage from hydrates during late Paleocene – early Eocene warming is qualitatively consistent with a concomitant decrease in deep ocean δ^{13}C as observed in benthic foraminiferal calcite (Zachos et al., 2001) and a long-term deepening of the calcite compensation depth (Hancock et al., 2007).

4.3. Leads and lags

The genus *Apectodinium* originated close to the Danian- Selandian boundary (Brinkhuis 1994; Guasti et al., 2006) but abundant occurrences were restricted to low latitudes until the PETM (Bujak and Brinkhuis 1998; Iakovleva et al., 2001). On a global scale, *Apectodinium* is an important (often dominant) constituent of the dinoflagellate cyst assemblages described from the PETM (Heilmann-Clausen 1985; Steurbaut et al., 2003; Sluijs et al., 2007a; Sluijs et al., 2006). At ODP Site 1172, however, the lowermost acme starts approximately 70 cm below the CIE. An influx of abundant *Apectodinium* has also been shown to lead the CIE on the New Jersey Shelf, the Central North Sea and, perhaps, New Zealand (Sluijs et al., 2007b), by approximately 5 kyrs (perhaps slightly longer if sedimentation rates decreased in response to sea level rise; see Sluijs et al. (2008b)). The early *Apectodinium* acme recorded at ODP Site 1172 can be interpreted in two ways. First, if the uppermost Paleocene record at ODP Site 1172 is relatively expanded, the early acme may actually correlate to the early onset recorded at other sites. Latest Paleocene sedimentation rates of ~10 cm/kyr are required to support this hypothesis, which is significantly higher than the average across this part of the section at ODP Site 1172, but quite common for marginal marine settings. If so, the rise in TEX$_{86}$ around 612 rmbsf might comprise the early warming recorded in New Jersey between the onset of the *Apectodinium* acme and the CIE (Sluijs et al., 2007b), although the latter records do not show a subsequent cooling immediately prior to the onset of the CIE. We cannot exclude this hypothesis because of the poor constraints on sedimentation rates in the uppermost Paleocene part of the section. Secondly, average sedimentation rates of 5.7 m/Myr suggest that this acme leads the CIE by some 100 kyr. If so, the early *Apectodinium* acme may imply that conditions at ODP Site 1172 locally became similar to low-latitude equatorial environments ~100 kyr prior to the CIE, but unrelated to the PETM. This would imply extremely anomalous environmental change on the East Tasman Plateau, associated with the first and mass occurrence of a typical low-latitude dinoflagellate in the Southern Ocean, which is not accompanied by significant change in other proxies and, critically, not recorded in nearby sections at lower

latitudes in New Zealand (Crouch and Brinkhuis 2005; Crouch et al., 2003; Crouch et al., 2001). Although, this hypothesis seems inconsistent with the strictly low-latitude biogeography of abundant Paleocene *Apectodinium*, we cannot exclude it with the present data.

4.4. Sea level, hydrology and productivity

Low sedimentary Ca values are in line with a shallow marine depositional environment, dominated by siliciclastic input. The Ca record shows a peak during the lower part of the PETM. The persistence of carbonate accumulation in this interval indicates that the calcite compensation depth resided below the shelf at the East Tasman Plateau during the entire event, consistent with other shelf locations (e.g., Bolle et al., 2000; John et al., 2008). Representatives of the genus *Senegalinium* dominate dinocyst assemblages for most of the studied interval (Fig. 3). Dinocysts assignable to this genus have been shown to tolerate very low salinities (Brinkhuis et al., 2006). High *Senegalinium* abundances have been associated with salinity stratification on the New Jersey Shelf during the PETM (Sluijs and Brinkhuis 2009). Moreover, *Senegalinium* likely represents heterotrophic dinoflagellates, thereby thriving in relatively nutrient-rich waters (Sluijs et al., 2005). Accordingly, consistent with lithological information, we interpret high abundances of *Senegalinium* spp. prior to the PETM as to indicate near-shore, relatively high-productive shelf settings, with sustained fresh-water runoff from nearby rivers. This interpretation is corroborated by relatively abundant river-transported terrestrial palynomorphs (Fig. 3). Although the low BIT values suggest a relatively low contribution of riverine transported soil derived organic matter, and relative terrestrial palynomorph abundances are not as high as seen in other shelf settings across the PETM (e.g., Crouch et al., 2003; Sluijs et al., 2006) these aspects most likely rather reflect high burial fluxes of marine organic matter and isoprenoid GDGTs compared to terrestrial organic matter. A decrease in *Senegalinium* abundance at 613.3 rmbsf is accompanied by a peak in *Glaphyrocysta* spp., a taxon of which abundant occurrences are often associated with transgressive system tracts and sea level rise (Iakovleva et al., 2001; Pross and Brinkhuis 2005), suggesting uppermost Paleocene transgression at ODP Site 1172. Generally increasing abundances of other normal marine taxa, such as *Pyxidinopsis* spp., *C. fibrospinosum* cpx., *Spiniferites* spp., and *Operculodinium* spp. support this interpretation. A second peak in *Glaphyrocysta* spp. at the onset of the CIE also suggests renewed sea level rise during the early stages of the PETM, supported by dropping BIT index values and a second drop in *Senegalinium* abundances. This transgression is seen along continental margins on a global scale, including in aspects of dinocyst assemblages in the New Zealand sections (Crouch and Brinkhuis 2005), and has, hence, shown to represent eustatic rise (Sluijs et al., 2008b). A record of sea level rise near the Antarctic margin is particularly interesting regarding the potential contributing role of melting Antarctic ice sheets. If an Antarctic ice sheet would have been present in the Paleocene, its self-gravitation should have increased sea level around the continental margin. If the ice sheet melted during the PETM, the loss of gravity should have lead to a decrease in sea level during the PETM close to Antarctica, similar to that projected for the Greenland margin if the Greenland Ice Sheet would melt (Bamber et al., 2009). Hence, the record of sea level rise during the PETM at ODP Site 1172 might imply that no significant ice sheet was present on Antarctica during the Paleocene.

4.5. Hydrology and sediment supply

Most studied marginal marine sites exhibit a vast increase in sediment supply from the continent during the PETM (Sluijs et al., 2008b), including New Jersey (John et al., 2008), Lomonosov Ridge in the Arctic Ocean (Sluijs et al., 2008a), the North Sea (Sluijs et al., 2008b; Steurbaut et al., 2003), the Bay of Biscay (Pujalte and Schmitz 2006; Schmitz et al., 2001), the Central Northern Tethys (Giusberti et al., 2007), California (John et al., 2008), and several sections in the southwest Pacific region, notably in New Zealand, such as Tawanui (Crouch et al., 2003) and the Clarence River Valley (Hollis et al., 2005; Nicolo et al., 2011). This is generally interpreted as an increase in terrestrial weathering and resulting supply of siliciclastic material to the shelf by rivers. At ODP Site 1172, no obvious change in average sedimentation rates was recorded during the PETM. This could mean that the strong hydrological response as recorded elsewhere did not take place in the region of the East Tasman Plateau, which is supported by the decrease in fresh-water tolerant dinocysts (Fig. 2). Alternatively, transgression caused sediment condensation at this proximal site. In addition, during the sampling of the core, we did note that the PETM interval is slightly coarser-grained, although, as yet, no data were generated to quantify this. Hence, potentially the sediment-water interface experienced more intense winnowing, hampering the deposition of large amounts of clay.

4.6. Speculations on salinity and storminess

A peak of *Eocladopyxis* spp., a member of the extant family Goniodomidae that mainly inhabits polysaline, lagoonal environments (Wall et al., 1977), occurs within the PETM. Potentially, SSTs were only warm enough for this species to thrive in the Southern Ocean during peak PETM warmth, which may also seasonally have caused regional hypersaline conditions. The ecology of extant Goniodomids provides room for speculations regarding the conditions for the local environment. Abundant representatives of related species in the modern ocean (e.g., the harmful species *Pyrodinium bahamense*) have been related to hypersaline conditions (Reichart et al., 2004). Critically, however, in many regions, seasonal storm activity appears important to resuspend dormant cysts into the water column to hatch and fulfil their life cycle (Villanoy et al., 1996; Villanoy et al., 2006). In such systems, the subsequent bloom initiates in the wake of the storm when salinities drop due to increased river run off, and when turbulence is minimal (e.g., Dale 2001; Siringan et al., 2008). Increased river runoff and surface ocean stratification might be induced by tropical storms. Hence, storm activity and seasonal river input might have increased in the southwest Pacific region during the PETM, consistent with increased abundances of terrestrial palynomorphs. In any case, a peak in Goniodomids has at ODP Site 1172 only been recorded during the PETM (Fig. 3) and the EECO (Bijl et al., 2007; Brinkhuis et al., 2003a; Sluijs et al., 2003), indicating a very particular environment for this region, likely associated with a change in seasonality of regional hydrology, maximum temperatures and perhaps storm activity.

4.7. Highly variable assemblages within the PETM

The dinocyst record suggests relatively stable conditions through the latest Paleocene and some more variation close to the onset of the CIE and within the PETM, with short-lived abundances of *Glaphyrocysta*, *Eocladopyxis*, *Pyxidinopsis*,

129

Cordosphaeridium fibrospinosum complex, *Spiniferites*, *Operculodinium*, and *Membranosphaera*. Such intra-PETM variability has also been recorded in continental deposits from Wyoming (Bowen et al., 2004; Kraus and Riggins 2007; Wing et al., 2005) and on the New Jersey Shelf (Sluijs and Brinkhuis 2009). Although at the moment the cause of these variations are unknown, they do suggest that climate during the PETM may have been much more variable and dynamic on time scales of 10^3–10^4 years, perhaps on a global scale.

5 Conclusions

A relatively complete PETM record was identified in sediments recovered from the East Tasman Plateau during ODP Leg 189, deposited at a paleolatitude of ~65°S. Sediments are almost devoid of biogenic calcite but yield rich organic microfossil assemblages. TEX$_{86}$ paleothermometry indicates that SSTs warmed by ~7°C to maximum values of 33°C during the PETM, with a magnitude similar to or slightly larger than the global estimate of warming. Such surprisingly warm SSTs for this latitude indicate that meridional temperature gradients were very low across the Paleocene- Eocene transition, even though the reconstructed SSTs may be biased towards summer temperatures. Maximum temperatures were similar to those during the EECO, perhaps implying similar greenhouse gas concentrations. If so, one may speculate that long-term late Paleocene to early Eocene warming, carbon isotope trends and superimposed hyperthermals, were associated with carbon release from the same reservoir, perhaps methane hydrates. The globally recorded acme of the taxon *Apectodinium* leads the CIE at ODP Site 1172, which may represent the same early onset as recorded on the New Jersey Shelf and the North Sea. A decrease in the abundance of the fresh water-tolerant dinoflagellate cyst *Senegalinium* suggests a decrease in the influence of river run off at the core site during the PETM, possibly in concert with sea level rise. However, a unique abundance of the euryhaline taxon *Eocladopyxis* may indicate a change in the seasonality of the regional hydrological system and an increase in storm activity. Finally, significant variations in dinocyst assemblages within the PETM indicate that southwest Pacific climates varied much more significantly over time scales of 10^3–10^4 years during the event, than during background late Paleocene – early Eocene times, consistent with records from the New Jersey Shelf and continental North America.

CHAPTER 5

Linkage between early opening of the Tasmanian Gateway and Antarctic cooling

The transition from the warm early Paleogene "Greenhouse" climates towards the "Icehouse" initiated at the end of the early Eocene, ~50 million years ago (Ma), when both deep ocean and southern ocean temperatures started cooling (Zachos et al., 2008; Chapter 2). Although recent work argues that cooling was mainly forced by declining atmospheric greenhouse gas concentrations (pCO$_2$; DeConto and Pollard 2003a, b; DeConto et al., 2008), the role of the opening of critical Southern Ocean gateways in Antarctic climate change (Kennett et al., 1974; Kennett 1977; Lawver and Gahagan 2003; Lagabrielle et al., 2009) remains uncertain. Here we present proxy-records from recently recovered sediments from the Wilkes Land Margin, Antarctica (Escutia et al., 2011), deposited west of the Tasmanian Gateway in the Australo-Antarctic Gulf. Integrated marine and terrestrial microfossil and organic geochemical analyses show a strong cooling of east Antarctic surface waters (3-4°C) and adjacent landmasses (~6°C), at ~49 Ma. Critically, this cooling coincides with the first incursion of typical southwest Pacific phytoplankton that migrated across an apparently shallow southern opening of the Tasmanian Gateway at that time. Collectively, this implies that the onset of the relatively cool westward Antarctic Counter Current through the Tasmanian Gateway diminished the influence of the warmer Australo-Antarctic waters. This suggests a significant role of the shallow opening of the Tasmanian Gateway in East Antarctic cooling, presetting this region for massive cryosphere development at the Eocene-Oligocene boundary at ~34 Ma.

Early work suggested that the opening of critical Southern Ocean oceanic conduits, the Drake and Tasmanian gateways, played a major role in Cenozoic Antarctic cooling and cryosphere development (Kennett et al., 1974; Kennett 1977). Recent new field and model evidence from the Tasmanian region essentially led to the rejection of this hypothesis: both the timing of the critical deepening (Stickley et al., 2004b), and reconstructed oceanic circulation patterns (Huber et al., 2004), do not support a direct link between gateway opening and Antarctic glaciation. Rather, ever declining Eocene atmospheric CO_2 levels, although poorly documented (Beerling and Royer 2011), are now regarded as the principal driver (DeConto and Pollard 2003a, b). The role of oceanic gateways may still have been important for regional climate change, particularly along eastern Antarctica. Prior to opening, this region was bounded by a southern derivative of the Proto-Leeuwin Current (PLC) in the Australo-Antarctic Gulf (AAG), which presumably yielded low-latitude Indian Ocean waters (Huber et al., 2004; Fig. 1). Early through flow of relatively cool surface waters from the Pacific should in this region significantly affect bioprovincialism and regional climate. However, very little is known about the timing, dynamics and climatic significance of shallow oceanic connection along the Tasmanian Gateway (Veevers et al., 1991; Woodburne and Case 1996), hampering quantification of the environmental consequences of opening.

Recently, Integrated Ocean Drilling Program (IODP) recovered a relatively complete early - middle Eocene shelf sequence at Site U1356 that yields ample organic matter, suitable to reconstruct the environmental conditions along the east Antarctic margin and hinterland (Fig 1). Combined organic geochemical and marine and terrestrial palynological information portray a near-tropical setting for the early Eocene (~53 Ma; Contreras et al., in prep.; Bendle et al., in prep.), and a sea-ice dominated earliest Oligocene (Escutia et al., 2011), evidencing major middle – late Eocene cooling. To capture the timing and duration of this cooling and assess the potential role of early Tasmanian Gateway opening, we studied assemblages of dinoflagellate cysts for ocean circulation reconstructions (Chapter 3), and biomarkers for continental air temperature (MBT/CBT; Weijers et al., 2007b) and sea surface temperature (TEX_{86}; Schouten et al., 2002; Kim et al., 2010; see Supplements for detailed discussions on these proxies).

The MBT/CBT and TEX_{86} from the Wilkes Land Margin indicate near-tropical conditions in the early Eocene, with air temperatures of 25-27°C and SSTs of 23°C or 32°C, depending on the applied calibration. In the latest early Eocene, MAAT cooled by ~6°C and SSTs cooled by 3-5°C, regardless of the calibration (see Supplements; Fig. 2). The TEX_{86} data from the Pacific side of the Tasmanian Gateway (ODP Site 1172; re-calibrated from Chapter 2; see Supplements), show a distinctly different pattern: southwest Pacific Ocean SSTs remain tropically warm or warmed slightly in the same interval (Fig. 2; Chapter 2). Critically, the SST trends in the southwest Pacific Ocean closely track deep-ocean temperatures, as indicated by the global stack of benthic foraminiferal oxygen isotopes (Fig. 2; Chapter 2). This implies that the recorded cooling starting at ~52-49 Ma in the Wilkes Land sector, was regional in nature.
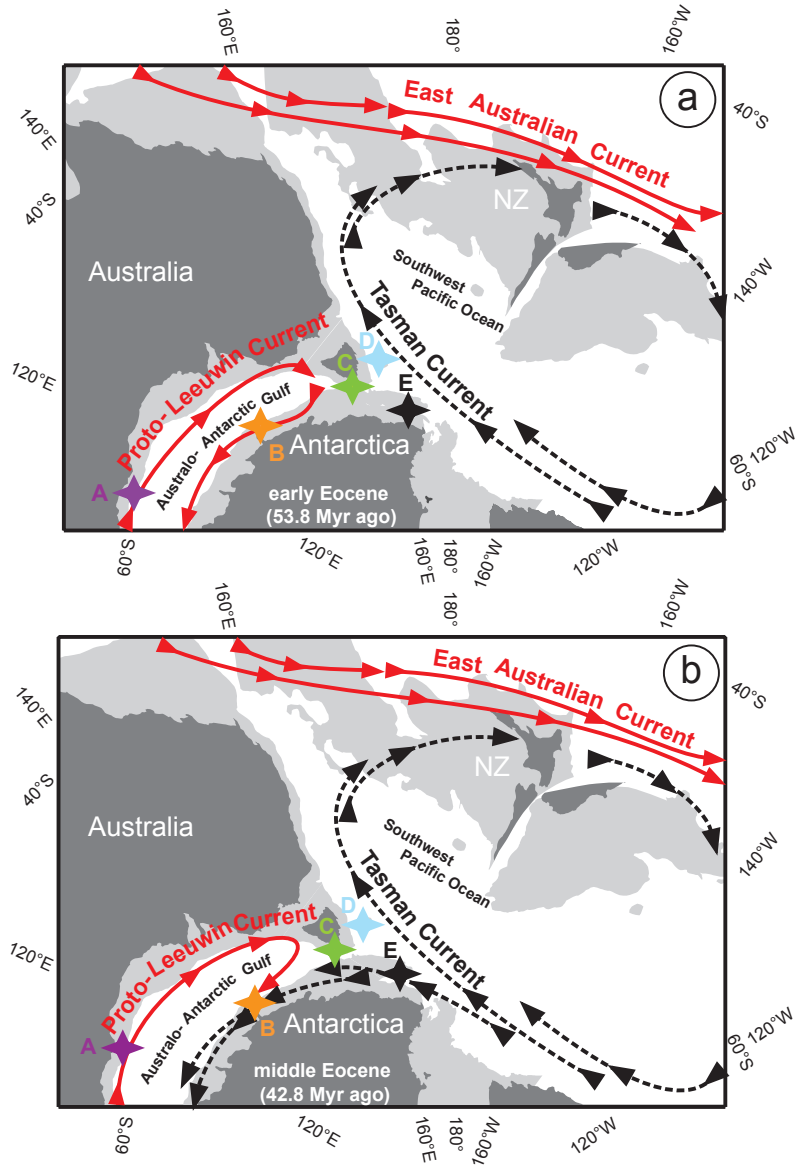
Figure 1. Eocene continental configurations of the Australian sector of the Southern Ocean for the early Eocene (a; Subchron C24n; 53.8 Ma) and middle Eocene (b; Subchron C20n; 42.8 Ma). Maps (modified from Cande and Stock (2004)) show the positions of the sites used in this study: ODP Site 1128 (A); IODP Site U1356 (B); *RV Sonne* and *RV Rig Seismic* gravity core sites (C); ODP Site 1172 (D) and ODP Sites 1171 (E). Dark-grey areas reflect present-day shorelines, light-grey areas are submerged continental blocks above 3000 meters water depth. NZ refers to New Zealand. The maps are overlain by surface currents as interpreted from modeling experiments of Sloan and Rea (1995), later reproduced by Huber et al. (2004). We here present evidence that the Tasmanian Gateway was open from 52-49 Ma onwards, which allowed for the through flow of the Antarctic Counter Current.
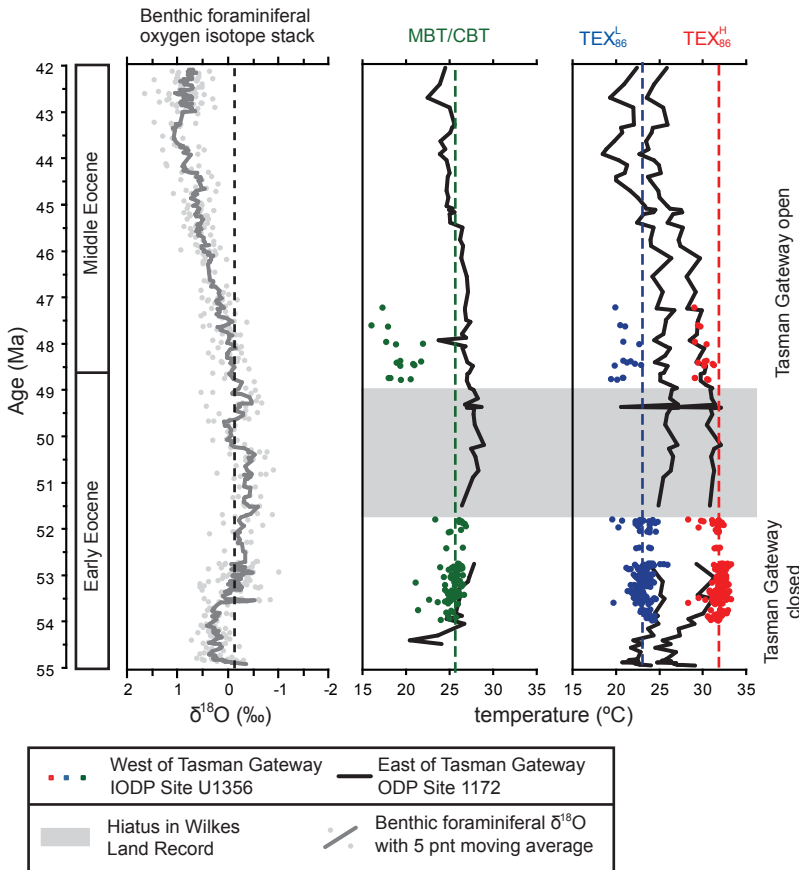
Figure 2. Benthic foraminiferal oxygen isotope stack (from Zachos et al., (2008) with amendments; see Chapter 2) MBT/CBT- (green) and TEX_{86}-based reconstructions (in two calibrations: TEX_{86}^{H} (red) and TEX_{86}^{L} (blue); see Supplements) for mean annual air and sea-surface temperature, respectively, from the west side of the Tasmanian Gateway (IODP Site U1356). These data are plotted with the temperature evolution from the east side of the Tasmanian Gateway (black lines; ODP Site 1172). Dashed lines represent average early Eocene temperatures for Wilkes Land as a reference for the lower middle Eocene temperatures.

We reconstruct oceanic surface circulation by tracking assemblages of organic cysts of dinoflagellates (dinocysts), which are highly sensitive recorders of surface ocean properties (Sluijs et al., 2005). Early Paleogene Southern Ocean dinoflagellate cyst (dinocyst) assemblages closely track different water masses (Chapter 3). Typically, assemblages from the southwest Pacific Ocean are dominated by endemic Antarctic dinocysts, while those from the AAG are mainly characterized by cosmopolitan and low- and mid-latitude-derived dinocyst taxa (Chapter 3). These bioprovinces mirror surface currents as projected by modeling experiments: the AAG is characterized by the low-latitude-derived PLC, while the southwest Pacific Ocean receives Antarctic-derived surface waters via the Tasman Current (Sloan and Rea 1995; Huber et al., 2004; Fig. 1a). In addition, middle Eocene dinocyst assemblages

from Ocean Drilling Program (ODP) Site 1128 on the Australian Bight almost entirely consist of cosmopolitan low- and mid-latitude taxa, consistent with the modeled circulation (Fig. 1; 3; see Supplements).

Abundant occurrences of the genus *Apectodinium* are restricted to (sub-) tropical settings in the early Paleogene (Iakovleva et al., 2001; van Roij 2009), apart from during Paleocene-Eocene Thermal Maximum, when it becomes globally abundant (Crouch et al., 2001; Sluijs et al., 2007a). Southern Ocean sediments, including the Southwest Pacific (Fig. 3), are typically devoid of *Apectodinium* even for the warm early Eocene. Uniquely, *Apectodinium* is dominant along the Wilkes Land margin in the early Eocene (50-80% of the dinocyst assemblage throughout; Fig. 3). We consider this dinocyst signal to document the influence of the PLC in the AAG (Fig. 1a). Endemic dinocysts exhibit much higher abundances in the early Eocene on the east side of the Tasmanian Gateway (~30%) as compared to the AAG (<10%; Fig 3), reflecting the influence of the Antarctica-derived Tasman Current in the Southwest Pacific Ocean (Fig. 1a).

The first incursion of Antarctic endemic dinocysts into the AAG is recorded in Wilkes Land in the uppermost early Eocene, at ~49 Ma, while these taxa already dominated South Pacific assemblages from ~50 Ma onwards (Fig. 3). Also in samples obtained from the southern tip of Tasmania (RV *Rig Seismic* and RV *Sonne* cores (Truswell 1997); Fig. 1, 2) endemic dinocysts dominate. Other sites on the Australian side of the AAG (ODP Site 1128; Feary et al., 2000) are consistently dominated by low-mid-latitude dinocysts during the middle Eocene (Fig. 3). Younger sites in that region are typically dominated by low-latitude and/or cosmopolitan taxa as well, with clear affinities to the PLC (Chapter 3). The dominance of endemic dinocysts with Pacific affinities at the Wilkes Land margin is hence unique for the western side of the Tasmanian Gateway. We consider this distinct pattern to reflect the onset of through flow of an Antarctic Counter Current from the southwest Pacific Ocean into the Southern part of the AAG (Fig. 2b). The timing of the onset of dinocyst endemism on the Wilkes Land shelf lies between 52 and 49 Ma, the uncertainty caused by a 2-3 million year long hiatus in Hole U1356A (see age model section in Supplements). The middle Eocene section from the Wilkes Land margin contains dinocyst taxa that are unknown from the southwest Pacific Ocean, such as the Northern Hemisphere *Phthanoperidinium alectrolophum*. This testifies to a persistent influence of the PLC on the AAG. It also implies that the Tasmanian Gateway apparently effectively blocked any eastward flow of the PLC waters. Through flow of this current to the Southwest Pacific Ocean did not occur until 35.5 Ma, when the influx of cosmopolitan and low-latitude dinocysts species to the southwest Pacific sites is documented (Stickley et al., 2004b).

The onset of Antarctic dinoflagellate endemism at the Wilkes Land margin at ~50 Ma coincides with an abrupt increase in spreading rates of oceanic crust in the AAG (Cande and Stock 2004), at least with the present resolution of dinocyst data and tectonic reconstructions. Sedimentary records from the continental blocks surrounding Tasmania suggest a gradual deepening during the middle Eocene (Röhl et al., 2004a). We hypothesize that the increased spreading rates in the AAG initiated a

gradual drowning of the continental blocks of the Tasmanian Gateway. Early stages of opening of the gateway are likely to have occurred between the South Tasman Rise and the Antarctic Margin, via the active Balleny Fracture Zone, south of 60° Slat. (Cande and Stock 2004). Because of the prevailing easterlies, this would be particularly favorable to westward through flow of an Antarctic Counter Current.
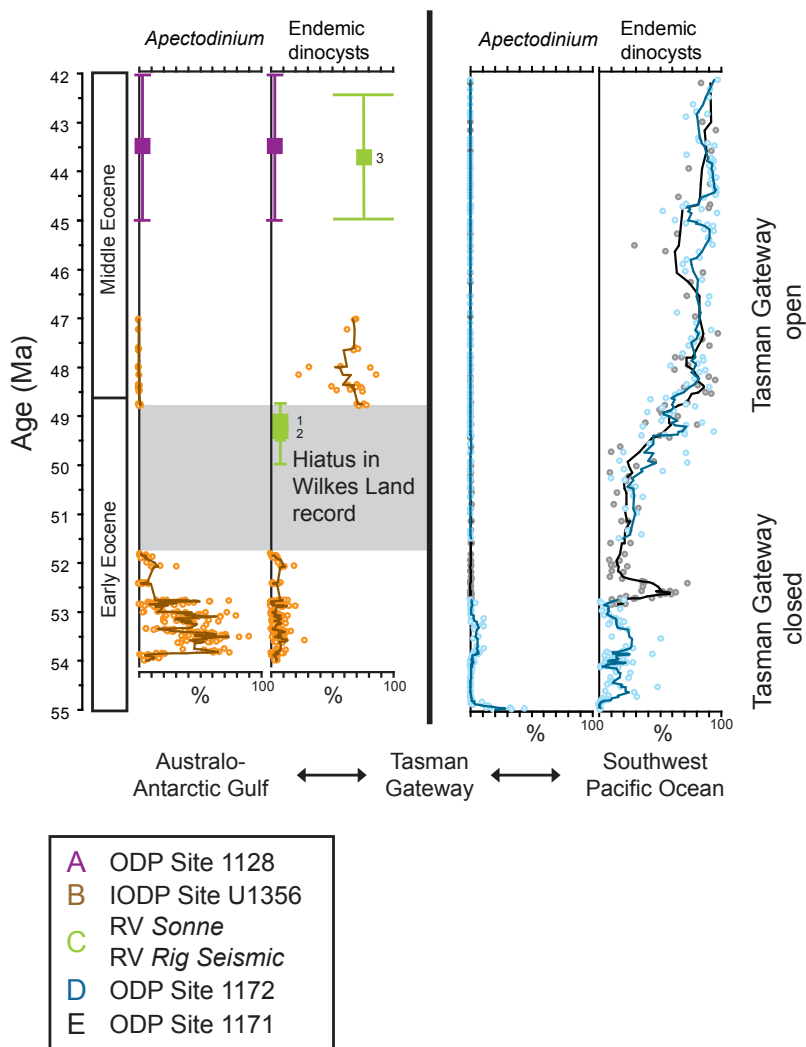

Figure 3. Selected dinocyst data from four investigated localities (letters refer to Fig. 1), combined for both sides of the Tasmanian Gateway: ODP Site 1128 (purple); IODP Site U1356 (orange); *RV Sonne* and *RV Rig Seismic* Gravity Core Sites (green; SO36-22SL (1), SO36-30KL (2) and RV *Rig Seismic* Core 147/GC01 (3) (Truswell 1997); ODP Site 1172 (black) and ODP Site 1171 (blue). Plotted is the relative abundance of *Apectodinium* spp. and endemic dinocysts.

General circulation model (GCM) experiments suggest that the climatic consequences of the late Eocene deepening of the Tasmanian Gateway were limited to a regional scale only and that there was little effect on inter-regional, let alone global climates (Huber et al., 2004; Sijp et al., in press). Deep and wide northern opening of the Tasmanian Gateway (which is the scenario investigated in these models) leads to the initiation of eastward through flow of the subtropical PLC. Eventually, this led to distinct warming of the SE offshore Australian sector by incoming subtropical waters, but only minor cooling along the Antarctic margin according to (Huber et al., 2004). However, a southern opening as reconstructed here, with through flow of just the Antarctic Counter Current, has not yet been investigated with GCM models, because that requires small grid cells and extensive computational time. Other GCM experiments point to declining atmospheric greenhouse gas concentrations as a major forcing for Paleogene global cooling (DeConto et al., 2008), although field data supporting this is little (Zachos et al., 2008). Our data support inferences that the onset of westward through flow of an Antarctic Counter Current invoked regional cooling along the eastern Antarctic margin, and that this process occurred much earlier than previously thought.

The tectonic phase separating the southern Tasmanian continental margin from Antarctica closely coincides with the first steps of global cooling following the warmest time interval of the past 100 million years. Proxy data and climate modeling suggested that greenhouse gas concentration changes were an important global climate forcing element in the greenhouse-icehouse transition. However, ice sheet models suggest that Southern Ocean gateways do modulate the timing and magnitude of continental ice sheet formation with declining atmospheric $p\mathrm{CO}_2$ (DeConto and Pollard 2003a, b). Hence, the opening of critical ocean gateways may still have provided critical climatic boundary conditions that increased Antarctic sensitivity to greenhouse gas forcing.

Supplement to Chapter 5

1. Material

1.1 Site and lithology descriptions
1.1.1 IODP Site U1356

Integrated Ocean Drilling Program Site U1356 is located at ~64° southern latitudes (Slat.) off Wilkes Land, Antarctica, at the transition between the continental rise and the abyssal plain at 4003 meters below sea level (Escutia et al., 2011). The lowermost 110 meters of core (890-1000 mbsf) are dated as early-middle Eocene based on dinocyst- and magnetostratigraphy. Sediments from 890 to 945 mbsf consist of interbedded stratified and massive sandstones, diamictites, silty claystones, and siltstones with graded bedding and parallel lamination. Abundant intraformational clasts are present in the fining-upward sandstones.

1.1.2 ODP Site 1128

Ocean Drilling Program Leg 182 Site 1128 is located on the upper continental rise at 3875 meters water depth (Feary et al., 2000). The Eocene section (290-452 meters below sea floor (mbsf)) consists of bioturbated clayey siltstone and sandy siltstone with glauconite. Initial results from the expedition suggest that the low concentration of carbonate in the lowermost sediments recovered at the site argues for deposition below the CCD or lysocline (Feary et al., 2000). Dinocyst stratigraphy confirms the age provided by the benthic foraminifer assemblages as late early Eocene (Feary et al., 2000). However, based on dinocyst-derived data we argue for syn-depositional dissolution of carbonate in shelfal environments, because dinocyst assemblages suggest upper shelfal settings in the Eocene. Also, in light of the drowning of the Wilkes Land and Tasmanian continental blocks in the Eocene (Stickley et al., 2004b; Close et al., 2009), the sections in the Australian Bight may have experienced deepening from shelfal settings to continental rise settings in the latter part of the Eocene.

1.1.3 ODP Site 1171

Ocean Drilling Program Leg 189 Site 1171 is located in lower bathyal water depths of about 2150 meters on the southern part of the South Tasman Rise, at 48° Slat. (Fig. S1; Exon et al., 2001). Drilling at ODP Site 1171 penetrated a small north-south oriented rift basin bounded to the east by the Balleny Fracture Zone (Exon et al., 2001). The Eocene part of the succession is cored within Hole 1171D, which penetrated to ~959 mbsf. The age model for this site suggests strata from early Eocene (mid Subchron C24n) to early Oligocene times (Stickley et al., 2004b; Chapter 1) in Hole 1171D. Sediments are composed of dark greenish-grey or olive-grey nannofossil-bearing claystones and silty claystones. Carbonate content decreases downhole, evidencing deepening trend going upsection. The bottom ~40 m of the sedimentary succession is pervasively laminated. Carbonate content is <1 wt% and organic carbon content (0.9 wt%) is higher than upsection. Sedimentation rates average 4-12 cm/kyr for the entire Eocene section (Exon et al., 2001).

1.1.4 ODP Site 1172

Ocean Drilling Program Leg 189 Site 1172 is located on thinned continental crust on the submerged flat western side of the East Tasman Plateau (ETP) at 45° Slat. and at a water depth of 2620 meters (Fig. S1; Exon et al., 2004). The plateau is roughly circular (200 km across), lies in water depths of 2200–2800 m, has late Cretaceous oceanic crust to the east and probably to the southwest and south, and is attached to Tasmania to the northwest (Exon et al., 2001). During the middle Eocene, the ETP was at ~65° Slat. and at much shallower water depths, before its rapid collapse and subsequent northward movement (55 km/m.y.) commenced in the late Eocene (Stickley et al., 2004b). Four holes were drilled in total, of which Hole 1172A extends to the mid-Lutetian (Exon et al., 2001). Hole 1172D contains sediments from mid-Lutetian downwards, with good overlap between the two holes.
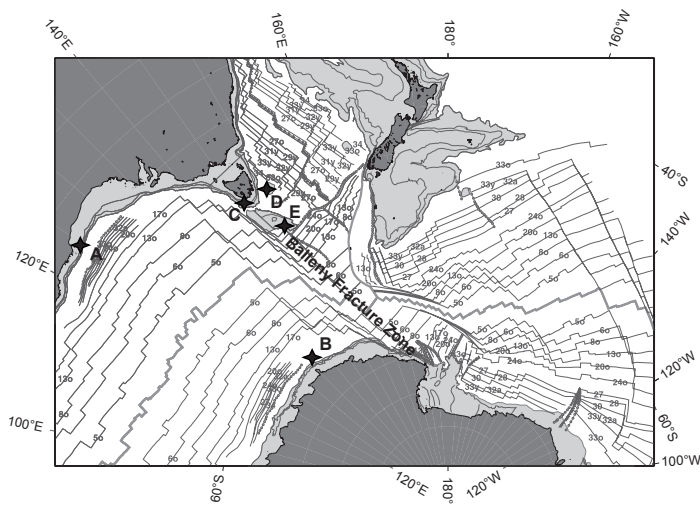

Figure S1. Present-day continental configuration of the Australian Sector of the Southern Ocean (map modified after Cande and Stock (2004)), showing the position of the sites used in this study: ODP Site 1128 (A); IODP Site U1356 (B); *RV Sonne* and *RV Rig Seismic* Gravity Core Sites (C) and ODP Sites 1172 (D) and 1171 (E). Also plotted are magnetic anomalies in the oceanic crust. Dark grey areas are subaerial, light grey areas are submerged continental blocks above 3000 meters water depth, white is below 3000 meters water depth.

Sediments used in this study are from about 500 to 605 mbsf and consist of olive-grey claystone with minor amounts of silty claystone, nannofossil-bearing claystone, and clayey siltstone (Exon et al., 2001). Age control in the section relies on dinocyst biostratigraphy and magnetostratigraphy (Stickley et al., 2004a; Chapter 1). A first inventory of dinocyst species found during ODP Leg 189 is published in the Scientific Results section of the Ocean Drilling Program (Brinkhuis et al., 2003a). The magnetostratigraphy allowed for comparison of the Southern Ocean bio-events found at ODP Site 1172 to global dinocyst events (Williams et al., 2004), and a dinocyst zonation scheme for the early Paleogene was proposed (Chapter 1). The sandstones also contain mica and pyritized shell fragments. Within Core -98R distinctive reddish brown silty claystones are repetitively interbedded at the decimeter- to meter-scale

with greenish grey and brown sandy mudstones with parallel laminae. The lowermost 50 meters of sediment consists of dark green bioturbated claystone that is faintly stratified (millimeter- to meter-scale) as defined by colour variations (light to dark bands; Escutia et al., 2011).

2. Methods

2.1 Palynology

2.1.1 Sample processing (ODP Site 1128 and IODP Site U1356)

Palynological data from ODP Sites 1171 and 1172 were published in Chapter 1 and 3; we refer to that chapter for processing methodology for samples of those sites. Standard procedures of the Laboratory of Palaeobotany and Palynology of Utrecht University were used to process samples for palynology, with some amendments. Samples were crushed to chips of several millimeters, and oven-dried (60˚C) or freeze-dried for several days. Subsequently, samples were weighted and put in 200 ml plastic bottles with screw-on caps. A tablet containing a known amount of *Lycopodium* spores (18,583 ±4.1% spores per tablet) was added to each sample for quantitative analyses. Samples were wetted with a 10% Agepon wetting detergent, and subsequently 10% HCl was added to carbonate-free samples. In case carbonate was present, 30% HCl was added, and samples were settled overnight, decanted, water was added and centrifuged (2000 rpm, 5 minutes). Samples were decanted again. For all samples, silicates were removed by adding 38% HF. Samples were shaken on a shaker table for 2 hours (300 rpm), bottles were filled up with water and samples settled overnight. Subsequently, bottles were decanted, and an overload of 30% HCl was added to remove silica gels. Samples were centrifuged (2000 rpm, 5 minutes), decanted and bottles were filled to halfway with HF. Samples were shaken for 2 hrs on a shaker table (300 rpm), filled up with water and settled overnight. The next day samples were decanted, and removal of silica gels was done again by adding an overload of 30% HCl, samples were centrifuged (2000 rpm, 5 minutes), decanted and subsequently rinsed with water, centrifuged and decanted. The rinsing was done 3 times.

From the organic residue, a small subsample was taken for pyrite analyses. The remainder of the residue was sieved over a 250-micrometer nylon sieve to remove large particles. The filtrate was subsequently sieved over a 10-micrometer sieve. For sieving, an ultrasonic bath was used to disintegrate palynodebris. The organic residue was separated from heavier particles (pyrite and other particles) by placing it into a ceramic bowl in the ultrasonic bath for 5 minutes: the heavy particles settled, and the palynofacies remained suspended. The ceramic bowl was subsequently decanted and the palynofacies were transferred into a glass vial. Heavy particles left over in the ceramic bowl after decanting, were transferred into the vial containing the unsieved residue for pyrite analyses. Palynofacies residues were centrifuged (2000 rpm, 5 minutes), decanted and glycerin water was added. The residue was transferred to microscope slides and slides were sealed using nail polish. Two slides were made for each sample.

2.1.2 Dinocyst taxonomy and endemism

Dinocyst taxonomy follows that as cited in Fensome and Williams (2004) and with emended dinocyst species and formally described new species such as Sluijs et al. (2009b) and Clowes and Wilson (2006). We follow the inventory of endemic dinocysts as listed in Chapter 3.

2.2 Organic geochemistry IODP Site U1356

Organic compounds were extracted from powdered and freeze- dried sediments of Site U1356 with dichloromethane (DCM)/ methanol (MeOH) (9:1, v/v) using the accelerated solvent extraction technique (Dionex). Excess solvent was removed using rotary evaporation under vacuum. The total lipid extracts (TLEs) were dried under nitrogen flow, weighed, and split where possible. An aliquot of the TLE was further separated in an apolar, ketone, ethyl acetate (EtOAc) and polar fractions over an activated Al_2O_3 column using hexane:DCM (9:1, v/v), hexane:DCM (1:1, v/v), EtOAc:DCM (1:1 v/v) and DCM:MeOH (1:1, v/v), respectively. Aluminum oxide (Al_2O_3) was activated by oven-drying for at least 4 hrs at 150°C. All fractions were dried over nitrogen flow and subsequently weighed.

The polar fractions were dissolved in a 99:1 hexane/propanol solvent, and filtered using a 0.45 µm, 4 mm diameter polytetrafluoroethylene (PTFE) filter, before being analyzed using a high performance liquid chromatography/ atmospheric pressure positive ion chemical ionization mass spectrometry (HPLC/ APCI-MS). HPLC/APCI-MS analyses were done according to Schouten et al. (2007a) using an Agilent 1100 series LC/MSD SL and separation and a Prevail Cyano column (2.1 x 150 mm, 3 mm; Alltech), maintained at 30°C. The GDGTs were eluted using a changing mixture of hexane and propanol as follows: 99% hexane: 1% propanol for 5 minutes, then a linear gradient to 1.8% propanol in 45 minutes. Flow rate was 0.2 ml per minute. Single ion monitoring was set to scan the 5 [M+H]+ ions of the GDGTs with a dwell time of 237 ms for each ion.

The TEX_{86}^L and TEX_{86}^H ratios are based on glycerol dialkyl glycerol tetraethers (GDGTs; Fig. S2) and were calculated according to Kim et al. (2010). For conversion to temperature we used the following equations based on core top calibrations of Kim et al. (2010):

$$SST = 68.4* TEX_{86}^H + 38.6 \ (R^2 = 0.87; n = 255; p < 0.0001) \qquad (5.1)$$

$$SST = 67.5* TEX_{86}^L + 46.9 \ (R^2 = 0.86; n = 396; p < 0.0001) \qquad (5.2)$$

MBT/CBT ratios were calculated according to Weijers et al. (2007b). They were converted to air temperature using the calibration:

$$MBT = 0.122 + 0.187 * CBT + 0.020 * MAT \ (R^2 = 0.77) \qquad (5.3)$$

Figure S2. Chemical structure of the glycerol dialkyl glycerol tetraethers that were used to establish the TEX_{86} proxy for mean annual sea surface temperature.

3. Application and uncertainties of TEX_{86} paleothermometry

Soil-derived input of membrane lipids can significantly influence TEX_{86} SSTs (Weijers et al., 2006). The input of soil-derived membrane lipids relative to the marine is estimated using the Branched versus Isoprenoid Tetraether (BIT) index (Hopmans et al., 2004). We discarded TEX_{86} SST estimates in those samples with BIT>0.4. Consequently, we excluded 25% of TEX_{86} data points (Fig. S3), but that did not significantly affect average early Eocene or middle Eocene SSTs. The latest core top data set for TEX_{86} and the statistical evaluation of the relationships between

Crenarchaeal lipids (Fig. S2) abundances and temperature resulted in two separate calibrations for TEX_{86}. It was found that the abundance of regional isomer of crenarchaeol (cren'), used in the original TEX_{86} calibration (Schouten et al., 2002) was not varying with temperatures in polar oceans (Kim et al., 2010). Therefore the TEX_{86} index has a much higher coefficient when the polar core top data (i.e., those with SSTs <5°C) are removed and thus the TEX_{86}^H (which is the log of TEX_{86}) was proposed to estimate SSTs >15°C:

$$SST = 68.4 * TEX_{86}^H + 38.6 \quad (r^2 = 0.87; n = 255; p < 0.0001) \tag{5.4}$$

Where

$$TEX_{86}^H = \log((II+III+cren')/(I+II+III+cren')) \tag{5.5}$$

For SSTs below 15°C, in particular the polar oceans and alternative index was proposed without the crenisomer, TEX_{86}^L. These two calibrations provide statistically the best-fit solutions to the core-top dataset (Kim et al., 2010). It is as yet unknown why the relationship between TEX_{86} and SST is atypical in colder environments, compared to the rest of the world. One could hypothesize that the archaeal community consists of different species at high latitudes, with different membrane adaptation to SST. Another possibility is that the extreme seasonality at high latitudes, with half a year of darkness, biases the TEX_{86} signal. In any case, this difference causes a conundrum when TEX_{86} is applied to high latitude settings in the early Paleogene greenhouse world. The Southern high latitudes experienced extreme warmth in the Eocene. Pollen analyses indicate coldest month minimum temperatures approaching 13°C. This would argue for the use of TEX_{86}^H. Yet, like present day, also the Paleogene Southern Ocean experienced extreme seasonal cycles, with several months of darkness each year. Until the mechanism explaining the difference in the two TEX_{86}-calibrations is solved, it is uncertain which calibration can be applied. For instance, if the seasonal extremes cause the present-day atypical TEX_{86}-SST relationship, then the geographical position of the Southern Ocean suggest the use of TEX_{86}^L irrespective of the SSTs being above or below 15°C. However, if the crenisomer is important in temperature adaptation at higher temperatures, as implied by Kim et al. (2010), then TEX_{86}^H should be applied in the Paleogene Southern Ocean. An argument for TEX_{86}^L is that it predicts SST >15°C just as good as TEX_{86}^L: the latter becomes important when extrapolations outside present-day temperature ranges are encountered. On the other hand, TEX_{86}^H provides the best fit with $U_{37}^{K'}$ SSTs in the middle Eocene of the southwest Pacific Ocean (Chapter 6). The calibration error is not plotted in the main text, but is ± 3°C for TEX_{86}^H and ± 4°C for TEX_{86}^L.
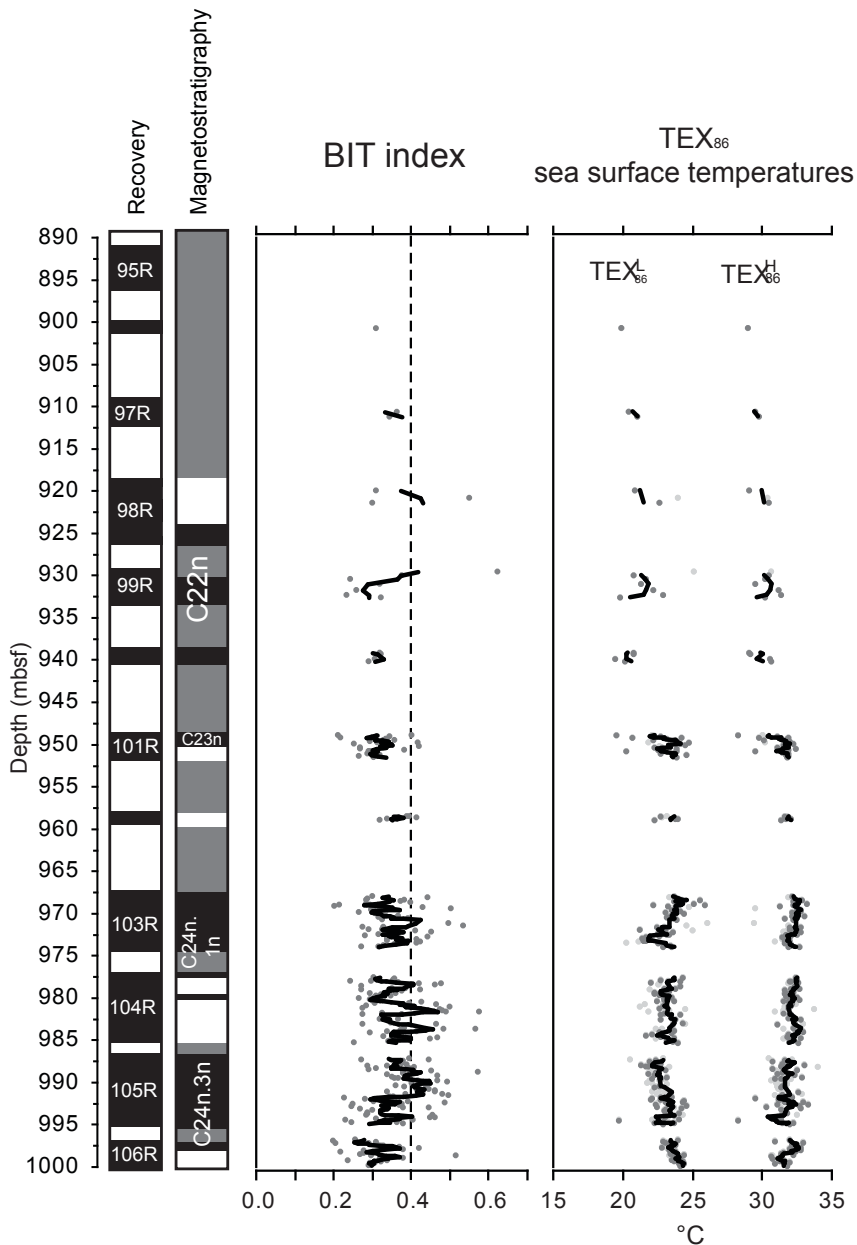
Figure S3. Branched versus isoprenoid tetraether (BIT) index values (left), and TEX_{86} SST inferences (right; in two calibrations, see text) of IODP Site U1356A. We use a cutoff value for BIT values of 0.4 (black dashed line), above which we discard the TEX_{86} estimates (grey TEX_{86} data points). This resulted in the loss of ~25% of the TEX_{86} data.

4. Constraints on MBT/CBT air temperature proxy

The MBT/CBT proxy for air temperature is a proxy derived from branched glycerol dialkyl glycerol tetraether (GDGT) membrane lipids (Fig. S4) and its relationship to pH. These membrane lipids are produced by a yet unknown group of anaerobic soil bacteria (Weijers et al., 2007b). In marine sedimentary records, notably those in shelf settings, these compounds were recognized as well, and were interpreted to originate from continental soils: soil erosion processes and runoff further transported the molecules to the marine realm (Hopmans et al., 2004; Walsh et al., 2008). The process involved in this transport causes -by definition- that the MBT/CBT air temperature signal is averaged over large areas and long time scales. However, MBT-CBT records of coastal marine sites shows that temperature changes are in phase with known climate events and do not show a considerable time lag (Weijers et al., 2007a; Weijers et al., 2007c).
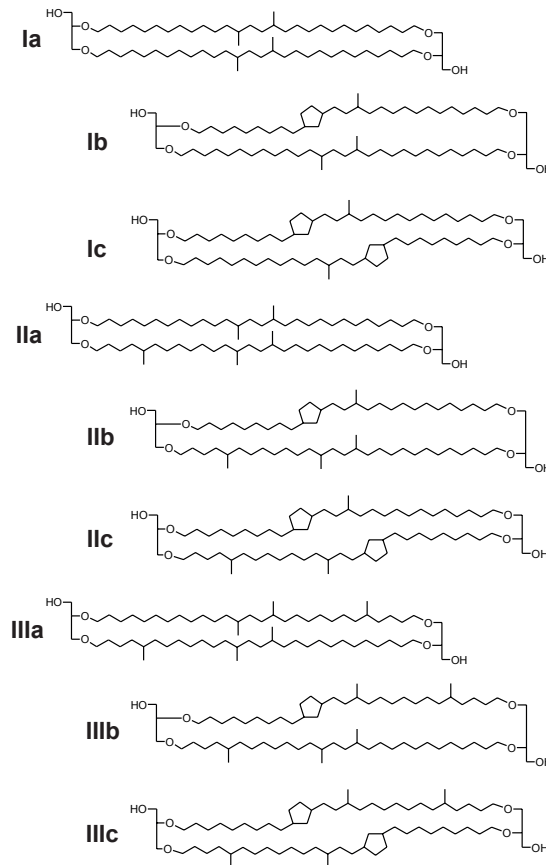
Figure S4. Chemical structure of the branched glycerol dialkyl glycerol tetraethers that were used to establish the MBT/CBT proxy for mean annual air temperature and soil pH.

The MBT/CBT proxy is calibrated against mean annual air temperature and a recent study could not detect a seasonal change in this proxy, although this does not exclude a seasonal bias (Weijers et al., 2011). Indeed, application of this proxy in the Cenozoic suggests that at high latitudes there may be a seasonal bias and that the proxy mainly reflects summer temperatures (Eberle et al., 2010). In contrast, other high latitude applications of this proxy show good matches with MAT records from other proxies (Schouten et al., 2008; Ballantyne et al., 2010). Therefore, it is uncertain to which extend the MBT/CBT proxy will be seasonally biased in Antarctica. The MBT-CBT temperature data fall in the middle of the temperature ranges of the pollen and spores, which suggests that the MBT-CBT proxy indeed reflects mean annual air temperature.

3 Age models

3.1 Age model ODP Site 1171 and 1172

An integrated dinocyst-magneto-isotope stratigraphic framework is developed from ODP Sites 1171 and 1172 (Chapter 1). Key dinocyst events were recognized at both sites that could be correlated to New Zealand sections (Crouch 2001; Crouch and Brinkhuis 2005) as well. Both sections were calibrated to magnetostratigraphy, also with additional age constraints from carbon isotope stratigraphy for the Middle Eocene Climatic Optimum (Chapter 6) and the Paleocene-Eocene Thermal Maximum (Chapter 4). The magnetostratigraphic age model from Chapter 1 is applied.

3.2 Age model IODP Site U1356

The coeval presence of *Impagidinium cassiculum* and *Samlandia delicata* in the lowermost samples of the record from IODP Site U1356A correlates to dinocyst zone NZE2b of Crouch and Brinkhuis (2005) and SP4 as described in Chapter 1 (Fig. S5). At ODP Site 1172, this dinozone is correlated to just below the C24r-C24n.3n magnetic reversal. That makes the oldest magnetic reversal in Core U1356A-106R (at 997.80 mbsf) the C24r-C24n.3n reversal (Fig. S5). This dinocyst zone also sees the FO of *Schematophora obscura*. The First Occurence (FO) of *Dracodinium waipawaense* is tentatively correlated to nannoplankton zone NP11 in New Zealand. At ODP Site 1172, the FO of *D. waipawaense* is within a hiatus spanning the latter half of Subchron C24n and C23r. Therefore, the FO of *D. waipawaense* correlates to Subchron C24n. At ODP Site 1171, the FO of *D. waipawaense* is poorly correlated to magnetostratigraphy, but seems to be correlated to mid-Subchron C24n. Consistently, at Site U1356A we find the FO of *D. varielongitudum*, which is the Northern Hemisphere ecophenopypic morphotype of *D. waipawaense*, correlated to a magnetic reversal between Cores U1356A-105R and -104R (~986 mbsf) that we here assign the C24n.3n-C24n.2r chronboundary to (Fig. S5). The subsequent cores (U1356A-104R and -103R) bear little biostratigraphic datums to allow for a high-resolution correlation to other sites; all dinocyst species present correlate to Subchron C24n. The top of Core U1356A-104R and Core -103R has a normal polarity, which we, based on the presence of *D. varielongitudum*, assign to Subchron C24n.1n. Core U1356A-102R has a reversed polarity and still contains *D. varielongitudum*. This core is correlated to Subchron C23r. The first occurrence of *Wetzeliella samlandica* and *Charlesdowniea columna*

within Core U1356A-101R (~951.04 and 948.9 mbsf) correlates to the onset of C23n. In turn, the reversal within Core U1356A-101R (~950 mbsf) is correlated to the C23r-C23n.2n reversal. Dinocyst assemblages within these cores are similar to those found in age-equivalent strata of ODP Sites 1172 and 1171, with high abundances of the Goniodomid and Wetzeliellloid complexes. Dinocyst assemblages in Core U1356A-100R are distinctly different from those below. Between Cores U1356A-101R and -100R, the last occurrence of *Wetzeliella samlandica*, *Homotryblium tasmaniense* and *Palaeocystodinium golzowense* is observed. Within Core -100R, *Arachnodinium antarcticum* has a FO, together with *Phthanoperidinium alectrolophum* and *Membranophoridium perforatum*. The FO of *A. antarcticum* correlates to Subchron C22n. Few specimens of *Damassadinium crassimuratum* were identified, with a LO within Core -98R, that correlates to C22n. The absence of *Enneadocysta multicornuta* and *Enneadocysta dictyostila* in Cores U1356A-100R through -96R correlates this entire section to Subchron C21r or older. The LO of in situ *Membranophoridium perforatum* makes the recovered section in Cores 95R-4 and 100R older than 46 Ma.
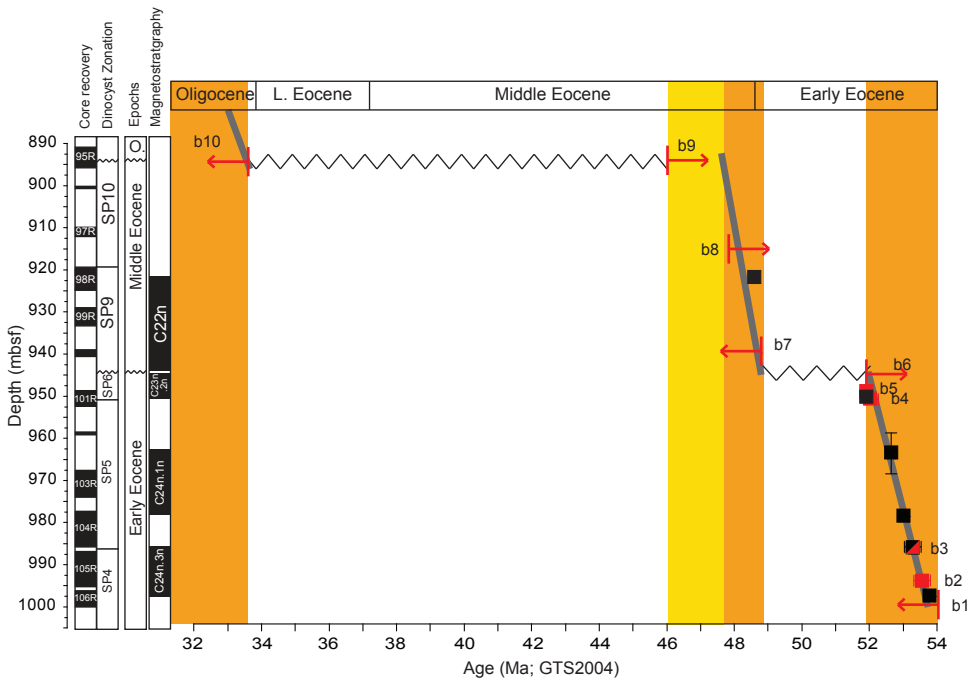
Figure S5. Age-depth plot of the lowermost 110 meters of core of IODP Site U1356. Plotted are dinocyst events (red) and (shipboard) magnetic reversals (black). Orange shading indicates which time intervals are recovered, yellow shaded area is uncertain. Grey line represents best fit through the bio-magnetostratigraphic data points. Biostratigraphic labels are as follows: b1 FO *Impagidinium cassiculum*; b2 FO *Schematophora obscura*; b3 FO *Dracodinium varielongitudum/ waipawaense*; b4 FO *Wetzeliella samlandica*; b5 FO *Charlesdowniea columna*; b6 LO *Palaeocystodinium golzowense*; b7 FO *Arachnodinium antarcticum*; b8 LO *Damassadinium crassimuratum*; b9 LO in situ *Membranophoridium perforatum*; b10 FO *Malvinia escutiana*.

Within Core U1356A-95R Section 3 interval 82-85 cm. we recognized the FOs of the Eocene species *Enneadocysta multicornuta, E. dictyostila, Deflandrea* sp. A sensu Brinkhuis et al., 2003a, *Schematophora speciosa, Reticulatosphaera acticoronata, Turbiosphaera sagena*, and *Stoveracysta kakanuiensis*. All these species have magnetostratigraphically calibrated FOs in the latter part of the Eocene (Subchrons C21n, C20r, C16n.1r, C16n.1r, C15r, C15r, and C13r, respectively). Concomitantly, this sample contains the FO of *Malvinia escutiana* (Houben et al., 2011). The first occurrence of this species has been calibrated to the Oligocene isotope event 1 (Oi-1) in the South Atlantic Ocean (Houben et al., 2011). The LO is not defined at the type locality, but ranges at least to 31.5 Ma (Houben et al., 2011). The coeval FOs of all species mentioned above suggests a hiatus between U1356A-95R section 3 and -95R section 4 that covers at least the time interval correlated to the FO of *Enneadocysta multicornuta* (onset of Subchron C21n) and the Oi-1 isotope event. According to the GTS2004 time scale, this hiatus spans at least from 47.24 Ma to 33.6 Ma. Subsequent strata contain numerous dinocyst species that originate from the time period covered by the hiatus.

3.3 Age model ODP Site 1128

We processed 22 samples from Cores 1128C-11H to -26X in Hole C (94-237 mbsf) and from 1128D-2R to -18R of Hole D (232-397 mbsf). Nine samples were barren for dinocysts. Dinocyst-bearing samples contain many taxa that are stratigraphically significant. These suggest a late Eocene age for the interval between 94 and 366 mbsf, with the presence of *Stoveracysta kakanuiensis*, *S. ornata, Corrudinium incompositum, Reticulatosphaera acticoronata* and *Schematophora speciosa* (Sluijs et al., 2003). The two lowermost samples, however, do not yield these species and hence are considered older. The presence of (rare) *Enneadocysta diktyostila* and *Cleistosphaeridium diversispinosum*, and the absence of *Corrudinium incompositum* suggest an age between 45 Ma and 40 Ma for the lowermost sample at Hole 1128D.

3.4 Age model RV Sonne and RV Rig Seismic gravity cores

With the Southern Ocean dinocyst stratigraphy being well-calibrated to magnetostratigraphy, we can confidently evaluate the ages for the sediments cored at Site SO36-22SL, SO36-30KL and 147/GC01. The presence of *Membranophoridium perforatum* (SPDZ7-SPDZ11; Chapter 1) in samples from SO36-22SL corresponds to the lower Lower *Nothofagidites asperus* Zone as correlated to by Truswell (1997). The absence of *Arachnodinium antarcticum* places the record time equivalent to Subchron C22n. SO36-30KL contains *Charlesdowniea edwardsii*, which has a narrow range in the latest Ypresian (dinocyst zone SPDZ9/SPDZ10 boundary; Subchron C22n; Chapter 1). Core 147/GC01 has a considerable younger age as is suggested by the presence of *Enneadocysta dictyostila* (as *E. partridgei*; Truswell 1997). This species has a first common occurrence in dinocyst zone SPDZ12, at ~45 Ma.

CHAPTER 6

Transient middle Eocene atmospheric CO₂ and temperature variations

Enneadocysta multicornuta (Eaton, 1971) emend. Stover and Williams, 1995
The central body has a sub-circular to sub-quadrate outline and a finely granular surface.
The processes are intratabular and have solid fibrous stems of variable length and breadth.
Each process is distally expanded and bifurcate; the attitude of the bifurcations varies from
patulate to recurved. Fenestration is sometimes present in a few of the process stems and
produces stem branches. The antapex is typically offset to the right of the mid-ventral line
and bears one process (1""). The archaeopyle is apical, tetratabular.

The long-term warmth of the Eocene (~56-34 Ma) is commonly associated with elevated partial pressure of atmospheric carbon dioxide (pCO$_2$). However, a relationship between the two has not been established for short-term climate perturbations. Here we reconstruct changes in both pCO$_2$ and temperature over an episode of transient global warming called the Middle Eocene Climatic Optimum (MECO; ~40 Ma). Organic molecular paleothermometry indicates a 3-6°C warming of southwest Pacific sea surface temperatures (SSTs). Reconstructions of pCO$_2$ indicate a concomitant two- to three-fold increase. The marked consistency between SST and pCO$_2$ trends during the MECO suggests that elevated pCO$_2$ played a major role in global warming during the MECO.

The Middle Eocene Climatic Optimum (MECO; ∼40 Ma; Bohaty and Zachos 2003) interrupts a long-term middle Eocene cooling trend (Zachos et al., 2001), with a globally uniform 4-6°C warming of both surface and deep oceans within ∼400 kyrs, as derived from foraminiferal stable oxygen isotope records (Bohaty et al., 2009). A decrease in carbonate mass accumulation rates during the MECO argues for ocean acidification induced by a rise in pCO$_2$ (Bohaty et al., 2009). Application of paleo-pCO$_2$ proxies across the MECO has yet to confirm whether pCO$_2$ changes are indeed associated with MECO warming.
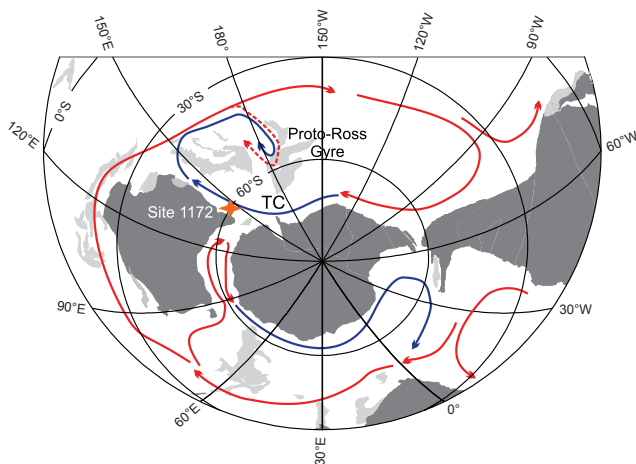

Figure 1. Paleogeographic configuration of the southern high-latitudes during the middle Eocene (∼49-37 Ma; map obtained from www.odsn.de) and ocean surface current configurations inferred from general circulation model experiments (Huber et al., 2004). The orange star indicates the paleogeographic location of ODP Site 1172 at 65°S in the southwest Pacific Ocean (Exon et al., 2004), under influence of the Antarctic-derived Tasman Current (TC).

We investigated a sedimentary succession spanning the MECO recovered from the East Tasman Plateau at Ocean Drilling Program (ODP) Site 1172, which at that time was situated on the shelf (∼65° South paleolatitude; Fig. 1; Supplements; Exon et al., 2004). To fully capture the magnitude of the SST change associated with the MECO at this site we applied two independent temperature proxies: the alkenone unsaturation index (U$^{K'}_{37}$; Brassell et al., 1986) and the index of tetraethers consisting of 86 carbon atoms (TEX$_{86}$; Schouten et al., 2002; Supplements). At the onset of the MECO, U$^{K'}_{37}$ and TEX$_{86}$ indicate a rise in SST of 3°C or 6°C, respectively, which, also at this location, stands out as an interruption of long-term middle Eocene cooling (Fig. 2). Bulk carbonate oxygen isotope values (δ^{18}O) decrease by 1.0-1.2‰, which, if forced by SST only, also indicates a ∼4–5°C SST rise. Additional evidence of warming is derived from assemblages of hypnozygotic organic cysts of surface dwelling dinoflagellates (dinocysts). While the middle Eocene dinocyst record at ODP Site 1172 is dominated by taxa that are endemic to the Southern Ocean (Brinkhuis et al., 2003a), an incursion of low-latitude dinocyst taxa characterizes the MECO (Fig. 2; Fig. S4). A 3-6°C SST increase is consistent with inferences from benthic foraminiferal and fine

154

fraction carbonate oxygen isotope records at other sites (Bohaty and Zachos 2003; Bohaty et al., 2009). The $U^{K\prime}_{37}$ and TEX_{86} proxies are independent of seawater $\delta^{18}O$. Hence, the consistent magnitude of warming between the proxies suggests that the carbonate $\delta^{18}O$ records were not affected by a change in $\delta^{18}O$ of seawater, and that global ice-volume did not change significantly during MECO.
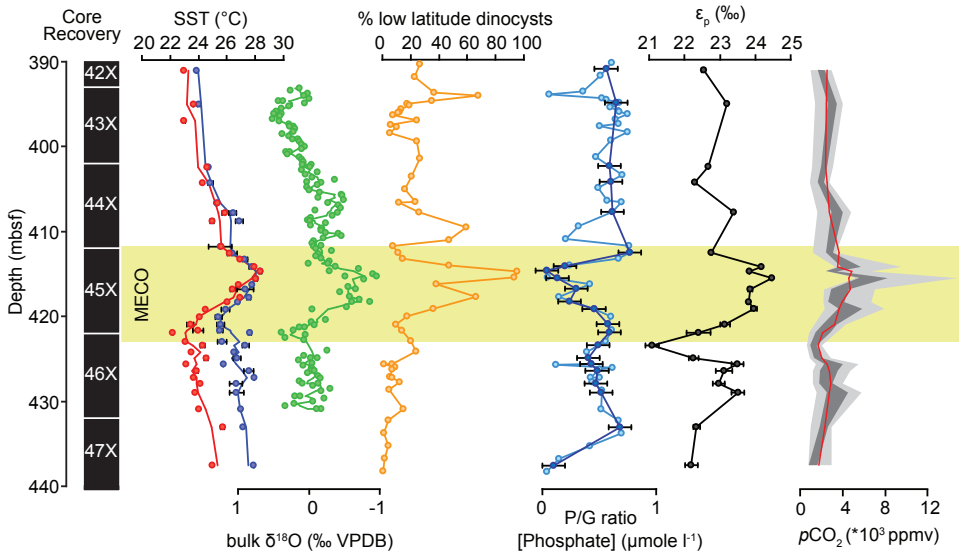

Figure 2. Geochemical and palynological results across the MECO at ODP Site 1172, Hole A, Cores 42X-47X. The MECO is identified by integrating magneto- (Stickley et al., 2004a), bio- (Stickley et al., 2004a) and isotope stratigraphy (Bohaty et al., 2009; see Fig. S2). The $U^{K\prime}_{37}$ (purple), TEX_{86} (red) and bulk $\delta^{18}O$ (green) SST reconstructions show a 3°C to 6°C warming. The yellow shaded area delimits the MECO interval. The increase in % low latitude dinocysts (in orange) at the expense of endemic dinocysts during MECO illustrates biotic response to warming. We estimated pCO_2 from carbon isotopic fractionation during carbon fixation (ε_p; black) by haptophyte algae with a phosphate concentration bandwidth of 0 and 1 μmol l^{-1} (light grey band). We further constrained phosphate estimates (dark blue line) by allowing phosphate concentrations to vary between 0.1±0.1 μm l^{-1} and 0.9±0.1 μm l^{-1} as a function of the ratio of peridinioid over gonyaulacoid dinocysts (P/G ratio; light blue line). This results in further constrained pCO_2 estimates (dark grey band). TEX_{86}, $U^{K\prime}_{37}$, oxygen isotope and pCO_2 data are plotted with a 3-point running mean (solid orange, purple, green and red lines, respectively). The error bars on TEX_{86} and $U^{K\prime}_{37}$ represent analytical error, and on ε_p the difference between TEX_{86} and $U^{K\prime}_{37}$ to calculate ε_p.

Absolute SSTs as indicated by $U^{K\prime}_{37}$ and TEX_{86} are consistent, with 26°C or 24°C just below the onset of MECO for the two proxies, respectively, and peak MECO SSTs exceeding 28°C. These values are much (~10°C) higher than those derived from fine fraction carbonate oxygen isotope measurements from elsewhere in the Southern Ocean (Bohaty and Zachos 2003; Bohaty et al., 2009). At least part of this large discrepancy is the result of diagenetic alteration of calcite (Pearson et al., 2001). We assessed pCO_2 changes by determining the stable carbon isotopic composition ($\delta^{13}C$) of alkenones – long-chained ketones exclusively synthesized by specific haptophyte

algae. Carbon isotopic fractionation during carbon fixation (ε_p) by haptophyte algae varies as a function of dissolved CO_2 $\left[CO_{2(aq)}\right]$ (Pagani 2002) specific cell-physiological parameters (which show good correspondence to the surface-water concentrations of soluble phosphate), and other environmental parameters, primarily light intensity. Surface ocean $\left[CO_{2(aq)}\right]$ originates from atmospheric pCO_2 and deep waters, of which the latter is of major importance in marginal marine upwelling areas. Changes in upwelling through time may substantially change $\left[CO_{2(aq)}\right]$ and hence would skew the reconstructed pCO_2 record. At Site 1172 we argue that a change in upwelling is not responsible for the signal we recorded in the alkenones, because that would have led to prominent shifts in the bulk carbonate carbon isotope profile (Supplements).

The carbon isotopic composition of diunsaturated alkenones ($\delta^{13}C_{37:2}$) ranges between -32.5 and -35.5‰ (Supplements). We used bulk carbonate $\delta^{13}C$ to estimate the $\delta^{13}C$ value of the dissolved inorganic carbon pool (DIC) in seawater to determine ε_p. The data show background ε_p values of 21-22‰ rising up to 24.5‰ during MECO (Figs. 2, 3, Supplements). The relationship between ε_p and pCO_2 is exponential, which results in a relatively large uncertainty in reconstructed pCO_2 levels with high ε_p values (Fig. 3). Temperature variations, however, play a minor role in the range of temperatures indicated by TEX_{86} and $U_{37}^{K'}$ (Fig. 2) and cannot explain the high ε_p values (Fig. 3). It seems unlikely that changes in light intensity influenced ε_p substantially at ODP Site 1172 (Supplements). The soluble phosphate concentration exerts a strong influence on the relation between ε_p and pCO_2, particularly if ε_p values are high.

To evaluate all possible absolute pCO_2 estimates from our record, we applied the full range of present-day surface water phosphate concentrations. These vary between 0 μmol l^{-1} in the oligotrophic gyres to >2 μmol l^{-1} in the Southern Ocean (Supplements). Yet, even when phosphate concentrations of 0 μmol l^{-1} are assumed, ε_p values between 21.2‰ and 24.5‰ yield pCO_2 estimates between 600 ppmv prior to MECO and 6,400 ppmv for the MECO (Fig. 2, 3). Hence, elevated levels of pCO_2 must in part be responsible for the high ε_p values, with middle Eocene pCO_2 being more than two times the pre-industrial value. When we assume maximal phosphate concentrations of 2 μmol l^{-1}, pCO_2 ranges between ~2,500 and ~24,000 ppmv (Fig. 2, 3). The marginal marine Eocene East Tasman Plateau (Exon et al., 2004) likely experienced phosphate concentrations that were higher than 0 μmol l^{-1} but smaller than 2 μmol l^{-1}, because closed oceanic gateways during the Eocene (Huber et al., 2004) prevented mixing associated with the Antarctic Circumpolar Current (ACC) that causes high phosphate levels in the present Southern Ocean. Eocene southwest Pacific surface water phosphate concentrations unlikely exceeded 1 μmole l^{-1}, which implies maximum pCO_2 estimates of 1,600 ppmv just prior to MECO and 15,000 ppmv for the MECO (Fig. 2, 3). With a realistic range of phosphate concentrations, pCO_2 values were between 600 and 1,600 ppmv just prior to MECO, which is in line with previous estimates of middle Eocene pCO_2 values using the same proxy (Pagani et al., 2005), and rose to between 6,400 and 15,000 ppmv during MECO (Fig. 2; light grey band). Critically, MECO values exceed any previous Eocene alkenone-based estimate even when we assume phosphate concentrations of 0 μmol l^{-1}. Despite uncertainties

regarding absolute pCO_2 values, we note that the trends in pCO_2 follow those in the SST records remarkably well (Fig. 2).

Surface water phosphate concentrations, however, may have varied in a marginal marine setting at ODP Site 1172. In fact, remains of dinoflagellates, which are known as extremely sensitive to surface-water nutrient availability changes, suggest a decrease in nutrient concentrations. The ratio between peridinioid and gonyaulacoid dinocyst groups (the P/G ratio) is often employed to reconstruct changes in relative nutrient abundance (Sluijs et al., 2005). The MECO at ODP Site 1172 shows a significant decrease in the P/G ratio (Röhl et al., 2004a; Fig. 2), suggesting a decrease in nutrient concentrations (Röhl et al., 2004a). As an experiment, we allowed phosphate to linearly vary as a function of the P/G ratio (Fig. 2). The most prominent shift towards low P/G ratio values occurs at MECO warming, resulting in lower phosphate concentration estimates. Also when this drop in phosphate is taken into account, pCO_2 rises during MECO and follows the SST trends (Fig. 2; dark grey band). Hence, regardless of the constraints on phosphate concentrations and other environmental parameters, pCO_2 levels must have been substantially higher during the MECO compared to the middle Eocene background.
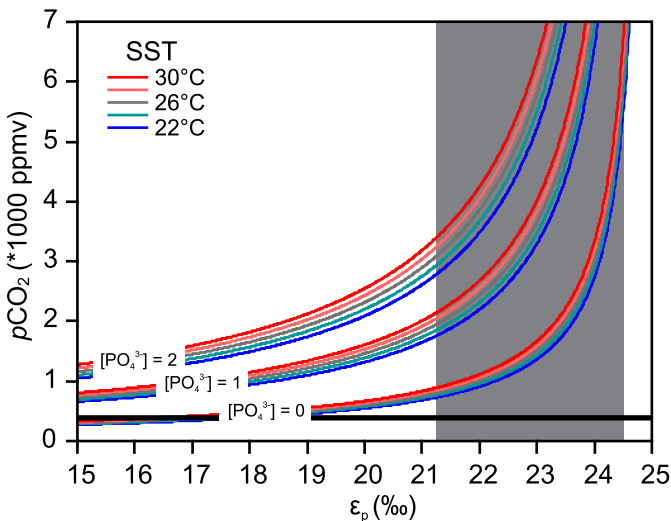
Figure 3. The relationship between ε_p (in ‰; versus VPDB) and pCO_2 (in ppmv). The phosphate concentration ranges plotted are those from the present day surface ocean, and the SST ranges (22°C to 30°C) are those inferred from TEX_{86} and $U^{K'}_{37}$ data presented in Fig. 2 and a suite of Southern Ocean sites (Chapter 2; Burgess et al., 2008; Hollis et al., 2009; Liu et al., 2009). The grey vertical bar indicates the values for ε_p as reconstructed for the MECO interval at ODP Site 1172, with use of TEX_{86}- and $U^{K'}_{37}$-based SST reconstructions for these same levels, and bulk carbonate $\delta^{13}C$ measurements on the same section. Black horizontal line represents present day pCO_2. Figure is modified from Pagani et al. (2002).

One outstanding issue is the source of carbon responsible for the increase in middle Eocene atmospheric CO_2. The rise in pCO_2 by 2,000 to 3,000 ppmv emerging from our data requires a carbon source capable of injecting vast amounts of carbon

into the atmosphere. Moreover, the absence of a significant negative carbon isotope excursion excludes reservoirs with δ^{13}C signatures significantly below that of marine DIC (Bohaty et al., 2009). One mechanism capable of emanating carbon with such a geochemical signature is the metamorphic alteration of carbonates (decarbonation; Bohaty and Zachos 2003). Massive decarbonation occurred until the late Eocene, with the subduction of vast amounts of Tethyan Ocean pelagic carbonates under Asia as India drifted northwards (Aitchison et al., 2007; Dupont-Nivet et al., 2008; Kent and Muttoni 2008). However, the flux of carbon required to increase pCO$_2$ by 2,000 to 3,000 ppmv within ~400 kyrs, appears high to invoke metamorphic (volcanic) outgassing as the sole mechanism.

Our pCO$_2$ and SST reconstructions allow for a tentative assessment of high-latitude climate sensitivity to CO$_2$ forcing on ~100 kyr time scales, assuming that all MECO warming was caused by pCO$_2$ and associated feedbacks. With an average 5°C SST increase and a two- to three-fold increase in pCO$_2$ we arrive at a climate sensitivity of ~2-5°C per pCO$_2$ doubling. When we consider the pCO$_2$ decline from the middle Eocene (~2,000 ppmv; this study) to the latest Eocene (~1,000 ppmv; Pearson et al., 2009) and the coeval high-latitude temperature decline (~3.5°C; Liu et al., 2009; Zachos et al., 2008) we derive similar values. Thus, long-term climate sensitivity to pCO$_2$ forcing in a world without the amplifying effects of ice-albedo feedbacks (Pagani et al., 2010) may have been larger than previously anticipated.

Supplements to Chapter 6

1. Material

The middle Eocene section at ODP Site 1172 is represented by Lithological Unit IIIA as described in the Initial Reports volume of Ocean Drilling Program Leg 189 (Exon et al., 2001). This unit consists of greenish grey and dark brownish grey diatom- and nannofossil-bearing claystones. Increasing calcium carbonate content, as well as the gradual up-section trend towards more offshore, oligotrophic organic walled dinoflagellate cyst (dinocyst) and diatom assemblages indicate a gradual deepening, starting ~42 Ma, and is probably related to the earliest stages of subsidence of the East Tasman Plateau (ETP; Röhl et al., 2004a). Low carbonate concentrations, however, still suggests marginal marine depositional settings; dinocyst assemblages confirm that the site was still located on the ancient shelf (Brinkhuis et al., 2003a). Subsidence rates of the ETP vastly increased due to Tasmanian Gateway deepening during the latest Eocene (starting around 35.5 Ma), leading to the deposition of carbonate oozes (Stickley et al., 2004b). The carbonate content is generally low (~0 to 15%) throughout the study interval, but increases somewhat up-section (Röhl et al., 2004a). The carbonate is mainly derived calcareous nannofossils (Exon et al., 2001).

2. Magneto-chemostratigraphic age model

The inclination data at ODP Site 1172 are characterized by a strong normal overprint (Exon et al., 2001). Fuller and Touchard (2004) used the z-intensity to interpret normal polarities, and with use of the available biostratigraphic datums, a bio-magnetostratigraphic age model was developed (Stickley et al., 2004a; Fig. S1; Column A). However, Bijl et al. (Chapter 1, 2) reinterpreted the magnetostratigraphy after the identification of the Paleocene-Eocene Thermal Maximum (PETM) and a hiatus spanning Magnetosubchron C23r (Bijl et al., 2007; Chapter 1). Following this interpretation, all normal polarity intervals up to C18n had to represent a younger Magnetochron, which was consistent with the available biostratigraphic information (Fig. S1; Column B).

Carbon isotope stratigraphy provides another calibration tool to correlate middle Eocene paleomagnetic signal to the Geomagnetic Polarity Time Scale. We assigned 9 tie points in the bulk carbonate carbon isotopes of ODP Site 1172 and correlated those to the calibrated carbon isotope records from ODP Site 748, 738 and 702 (Bohaty et al., 2009; Fig. S2). This shows that the original magnetostratigraphic interpretation by Stickley et al., (2004a) for Subchrons C18n to C20n was correct, which we incorporate in our age model (Fig. S1C). Because of the large uncertainties in the age model at high sample resolution we have decided to plot the data in the depth domain in this paper.
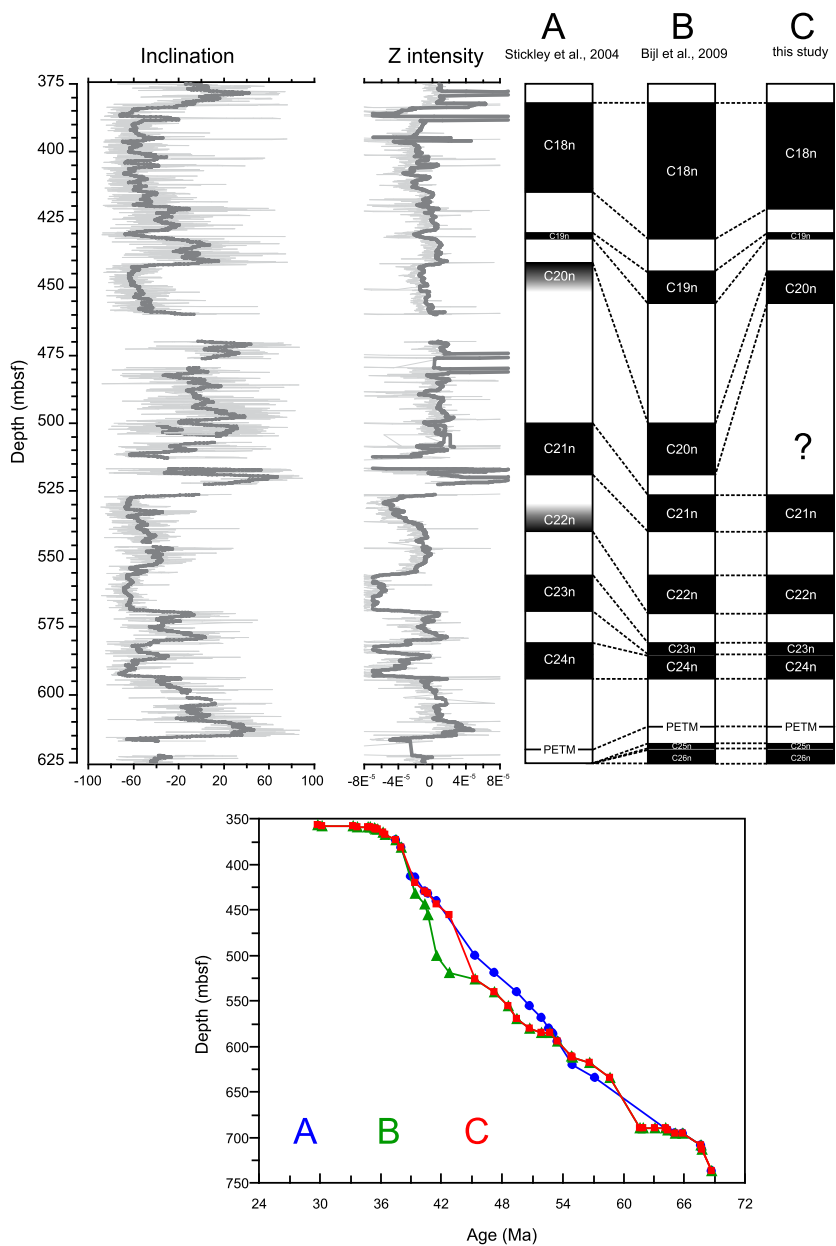
Figure S1. Magnetostratigraphy option A is published by Stickley et al. (2004a) in the Scientific Results section of the Ocean Drilling Program. Subsequently, the identification of the PETM and a hiatus spanning magnetosubchron C23n results in a revised magnetostratigraphic interpretation (Chapter 2) (B). Now, with carbon isotope stratigraphy for the MECO the magnetostratigraphic interpretation was revised again (C). All stratigraphies are (re-) calibrated to the time scale of Gradstein et al. (2004).

3. Methods

3.1. Palynology

Palynological processing followed the standardized quantitative methods used at the Laboratory of Palaeobotany and Palynology, Utrecht University, the Netherlands (Sluijs et al., 2003). Briefly, this involves reaction with ~30% HCl and ~38% HF, and ultrasonic separation. A 15μm nylon mesh was used for sieving. Residues were mounted on slides for microscope analysis. At least 200 dinocysts were counted per sample and identified to the species level at 500x magnification. Taxonomy is according to that cited in the Lentin and Williams Index of Fossil Dinoflagellates, 2004 (Fensome and Williams 2004), unless stated otherwise. We follow Chapter 3 (see Table S1) for groupings of endemic, cosmopolitan and low-latitude taxa. All samples and slides are stored in the collection of the Laboratory of Palaeobotany and Palynology, Utrecht University, the Netherlands.

3.2. Stable isotope analysis on bulk carbonate and inferred SSTs

Bulk-sediment stable isotope analyses of samples from the MECO interval of ODP Hole 1172A were performed using a VG Optima dual-inlet isotope-ratio mass spectrometer at the University of California, Santa Cruz, USA. The samples were first freeze-dried, and subsamples (~1.0 g) were then ground with an agate mortar and pestle in order to fully homogenize the sediment.

3.3. Organic geochemical analysis

Organic compounds were extracted from powdered and freeze- dried sediments with dichloromethane (DCM)/ methanol (MeOH; 9:1, v/v) using the accelerated solvent extraction technique (Dionex). Excess solvent was removed using rotary evaporation under vacuum. The total extracts were separated in apolar and polar fractions over an activated Al_2O_3 column using hexane:dichloromethane (DCM; 1:1, v/v) and DCM:MeOH (1:1, v/v), respectively. Approximately 200-500 μg of sample was reacted at 90°C in a common acid bath using an automated carbonate preparation system with a carousel device. NBS-19, an in-house Carrara marble standard, and a bulk-sediment consistency standard were included in all sample runs. All values are reported in standard delta notation (δ) in parts per mill (‰) relative to VPDB (Vienna Pee Dee Belemnite), and analytical precision is estimated at 0.04‰ (1σ) for $\delta^{13}C$ and 0.06‰ (1σ) for $\delta^{18}O$. Several samples with extremely low carbonate concentrations (<0.5%) have been excluded from the Hole 1172A dataset. We used the equation of Bemis et al. (1998) to interpret SST changes from bulk carbonate $\delta^{18}O$, where we assume no change in $\delta^{18}O_{sw}$.

3.3.1 TEX$_{86}$ paleothermometry

The polar fraction was dissolved in a 99:1 hexane/propanol solvent, and filtered using a 0.45 μm, 4 mm diameter polytetrafluoroethylene (PTFE) filter, before being analyzed using a high performance liquid chromatography/ atmospheric pressure positive ion chemical ionization mass spectrometry (HPLC/ APCI-MS). HPLC/APCI-MS analyses were done according to Schouten et al. (2007a) using an Agilent 1100 series LC/MSD SL and separation and a Prevail Cyano column (2.1 x 150 mm, 3 mm; Alltech), maintained at 30°C.
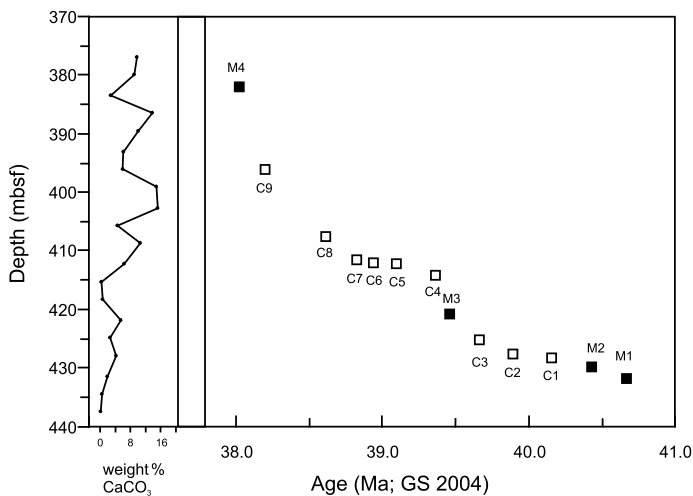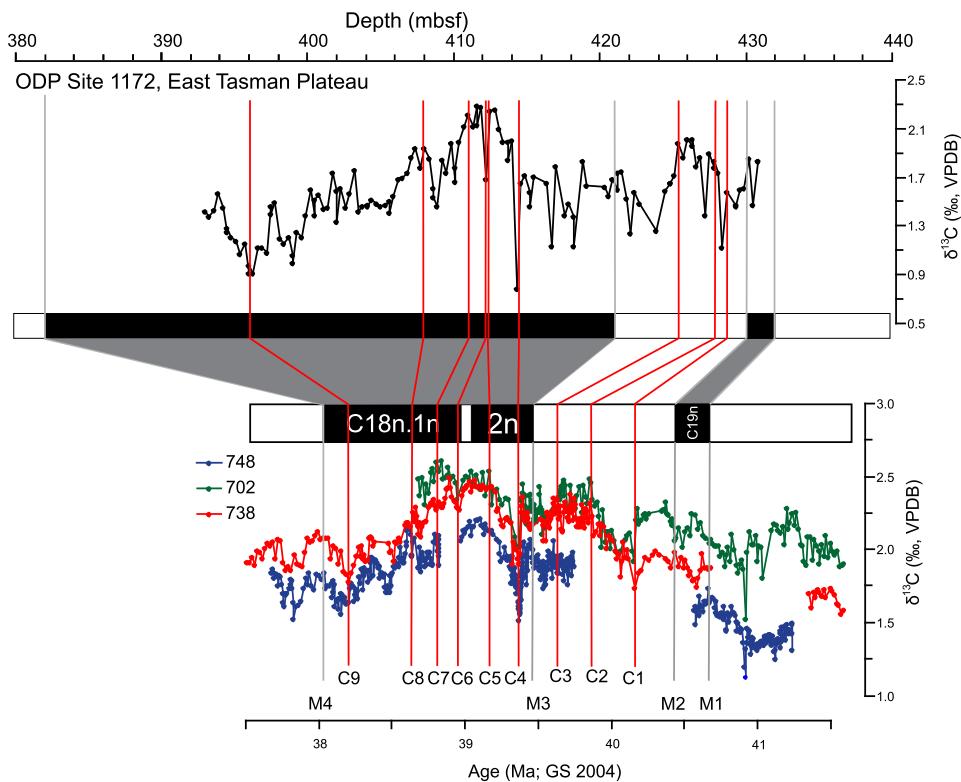
Figure S2. The bulk carbonate carbon isotope record from ODP Site 1172 was correlated to the magnetostratigraphically calibrated MECO records from ODP Site 702 (green) and the two Southern Ocean records ODP Sites 748 (blue) and 738 (red). C1-9: carbon isotope tie points, M1-4: magnetostratigraphic tie points.

Endemic and bipolar cysts	Low-latitude cysts	Protoperidinioid cysts	Cosmopolitan cysts	Unknown/other cysts
Alterbidinium distinctum	*Phthanoperidinium comatum*	*Brigantedinium* spp.	*Achomosphaera alcicornu*	*Batiacasphaera* spp. (pars)
Deflandrea antarctica	*Dracodinium rhomboideum*	*Lejeunecysta* spp.	*Achomosphaera ramulifera*	*Cerebrocysta* spp.
Phthanoperidinium sp. A*	*Achilleodinium biformoides*	*Octodinium askiniae*	*Lingulodinium machaerophorum*	*Membranosphaera* spp.
Phthanoperidinium sp. C*	*Batiacasphaera compta*	*Selenopemphix armata*	*Spiniferites* spp.	*Eocladopyxis* spp.
Phthanoperidinium stockmansii	*Cleistosphaeridium* spp.	*Selenopemphix nephroides*	*Thalassiphora pelagica*	*Paucisphaeridium* spp.
Spinidinium macmurdoense	*Corrudinium incompositum*		*Senegalinium* spp.	
*Spinidinium schellenbergii***	*Cribroperidinium* spp.			
*Vozzhennikovia roehliae***	*Cordosphaeridium fibrospinosum*			
*Corrudinium regulare****	*Dapsilidinium* spp.			
*Vozzhennikovia stickleyae***	*Enneadocysta multicornutum* group			
Vozzhennikovia apertura	*Enneadocysta pectiniformis*			
Arachnodinium antarcticum	*Glaphyrocysta* spp.			
Enneadocysta dictyostila group****	*Heteraulacacysta porosa*			
	Histiocysta spp.			
	Hystrichokolpoma rigaudiae			
	Hystrichosphaeridium tubiferum			
	Impagidinium spp.			
	Melitasphaeridium pseudorecurvatum			
	Hemiplacophora semilunifera			
	Hystrichosphaeridium truswelliae			
	Operculodinium spp.			
	Pentadinium spp.			
	Tectatodinium spp.			

Table S1. Dinocyst species encountered throughout the MECO record at ODP Site 1172 and their biogeographic affinity following Chapter 3. Taxonomy is based on that cited in the Lentin and Williams Index of Fossil Dinoflagellates, 2004 (Fensome and Williams 2004) apart from those indicated with an asterix: * undescribed species sensu Goodman and Ford (1983); ** species described by Sluijs et al. (2009b); *** species described by Clowes and Wilson (2006); **** species emended by Fensome et al. (2006).

The GDGTs were eluted using a changing mixture of hexane and propanol as follows, 99% hexane: 1% propanol for 5 minutes, then a linear gradient to 1.8% propanol in 45 minutes. Flow rate was 0.2 ml per minute. Single ion monitoring was set to scan the 5 [M+H]+ ions of the GDGTs with a dwell time of 237 ms for each ion. We applied the core top calibration equation from Kim et al. (2010) to translate TEX_{86} index values (Fig, S3; Table S2) to mean annual sea surface temperature:

$$SST = 68.4*TEX_{86}^{H} + 38.6 \tag{6.1}$$

Where

$$\text{TEX}_{86}^{\text{H}} = \log(\text{TEX}_{86}) \tag{6.2}$$

About half of the analyzed samples were run in duplicate, to estimate the reproducibility. In all of these samples, the reproducibility is better than 0.5°C and in most cases better than 0.25 °C.

3.3.2. Alkenone $U_{37}^{K\prime}$ paleothermometry

To purify the alkenones, the apolar fractions were separated into three fractions by activated Al_2O_3 column chromatography using hexane:DCM (9:1, v/v); hexane:DCM (1:1, v/v; the ketone fraction) and DCM. The ketone fractions were analyzed for alkenones by gas chromatography (GC) using an Agilent 6890 equipped with an on column injector. Separation was achieved using a 50m CPSil 5CB column (0.32 mm internal diameter, 0.12 micrometer film thickness). The ratio of $C_{37:2}$ and $C_{37:3}$ alkenones (Fig. S3; Table S2) were converted into temperature values by applying the core-top calibration of Müller et al. (1998):

$$T = (U_{37}^{K\prime} - 0.033)/0.044 \tag{6.3}$$

All samples were analyzed in duplicate and the reproducibility of the analysis is better than 0.5°C. Selected fractions were analyzed by GC/mass spectrometry (GC/MS) using a TRACE DSQ to confirm the presence of the alkenones and the absence of co-eluting compounds.

3.3.3. Alkenone stable carbon isotope analyses

The stable carbon isotopic composition of the C_{37} di-unsaturated alkenone ($\delta^{13}C_{C37:2}$) was measured in duplicate using GC-isotope ratio monitoring (IRM) MS using a Agilent 6890 GC coupled to a Thermo Science Delta V isotope mass spectrometer. GC conditions were as described above. The isotopic compositions of the compounds were calibrated against external reference CO_2 gas (in per mill versus VPDB) and the performance of the GC-IRMMS was checked by daily injections of two perdeuterated n-alkanes and comparison to off-line measured values. Average reproducibility was ± 0.2 (see Fig. S3; Table S3).

3.4 Alkenone pCO$_2$ estimates

Estimation of pCO$_2$ from $\delta^{13}C_{37:2}$ requires three steps: calculation of ε_p, [$CO_{2(aq)}$] and, ultimately, pCO$_2$.

3.4.1 Calculation of ε_p

From the stable carbon isotope ratio in alkenones ($\delta^{13}C_{37:2}$), the fractionation factor during photosynthesis (ε_p) of haptophyte algae (Fig. S3; Table S3) is expressed in the following equation (Pagani 2002):

$$\varepsilon_p = [(\delta_d + 1000/\delta_p + 1000) - 1]*1000 \tag{6.4}$$

Site (Hole)	Core (type)	Section (type)	Interval (cm)	Depth (mbsf)	Average TEX_{86}^{H} index	Average TEX_{86}^{H} SST	TEX_{86}^{H} SST error	Average $U_{37}^{K'}$ index	Average $U_{37}^{K'}$ SST	$U_{37}^{K'}$ SST error
1172A	42X	6W	10-12	391.00	-0.223	22.97		0.832	23.86	0.000
1172A	43X	2W	42-44	394.92	-0.224	23.62	0.035	0.837	24.02	0.063
1172A	43X	3W	85-87	396.85	-0.217	22.99	0.041			
1172A	43X	5W	37-39	402.37	-0.205	24.59	0.006	0.859	24.68	0.041
1172A	44X	2W	10-12	404.20	-0.209	24.29		0.864	24.84	0.121
1172A	44X	3W	102-104	406.62	-0.191	25.32		0.878	25.28	0.017
1172A	44X	4W	62-64	407.72	-0.191	25.90	0.158	0.918	26.47	0.139
1172A	44X	5W	10-12	408.70	-0.201	24.96	0.095	0.931	26.87	0.187
1172A	44X	7W	10-12	411.70	-0.190	25.60		0.887	25.54	0.576
1172A	45X	1W	32-34	412.52	-0.181	26.21		0.915	26.40	0.242
1172A	45X	1W	102-104	413.22	-0.163	26.91		0.944	27.27	0.141
1172A	45X	2W	32-34	414.12	-0.152	27.93	0.022	0.960	27.76	0.145
1172A	45X	2W	98-100	414.68	-0.151	28.36	0.074	0.979	28.35	0.037
1172A	45X	3W	22-24	415.52	-0.155	28.02		0.969	28.03	0.088
1172A	45X	3W	102-104	416.22	-0.176	26.84		0.961	27.78	0.048
1172A	45X	4W	10-12	416.80	-0.176	26.43	0.129	0.945	27.30	0.395
1172A	45X	4W	102-104	417.72	-0.171	26.93		0.953	27.55	0.105
1172A	45X	5W	10-12	418.30	-0.183	26.06		0.928	26.80	0.033
1172A	45X	5W	92-94	419.12	-0.206	24.48		0.899	25.91	0.202
1172A	45X	6W	32-34	420.02	-0.216	24.07		0.882	25.40	0.171
1172A	45X	6W	122-124	420.92	-0.221	23.41	0.160	0.887	25.54	0.190
1172A	45X	7W	42-44	421.62	-0.225	23.96	0.268	0.886	25.52	0.132
1172A	46X	1W	10-12	421.90	-0.240	22.18		0.956	27.63	0.076
1172A	46X	1W	112-114	422.92	-0.219	23.07		0.891	25.66	0.231
1172A	46X	2W	10-12	423.40	-0.208	24.28	0.106	0.945	27.31	0.174
1172A	46X	2W	92-94	424.22	-0.212	23.50		0.919	26.51	0.075
1172A	46X	3W	10-12	424.90	-0.207	24.53	0.097	0.926	26.72	0.173
1172A	46X	3W	82-84	425.62	-0.221	23.13		0.893	25.74	0.001
1172A	46X	4W	10-12	426.40	-0.214	23.87	0.062	0.953	27.55	0.255
1172A	46X	4W	92-94	427.22	-0.219	23.62		0.967	27.96	0.004
1172A	46X	5W	10-12	427.90	-0.212	24.11		0.923	26.64	0.306
1172A	46X	5W	114-116	428.96	-0.217	23.78		0.925	26.69	0.351
1172A	46X	7W	10-12	430.90	-0.201	24.04		0.934	26.98	0.052
1172A	47X	2W	10-12	433.00	-0.187	25.70	0.088	0.939	27.13	0.000
1172A	47X	5W	10-12	437.50	-0.199	24.97		0.965	27.89	0.015

Table S2. TEX_{86}^{H} and $U_{37}^{K'}$ index values and inferred SSTs for ODP Site 1172.

For this calculation, two unknowns need to be constrained:

- δ_p is the carbon isotopic composition of haptophyte organic matter, which is enriched in ^{13}C by 4.2‰ relative to $\delta^{13}C_{37:2}$ (Laws et al., 2000).
- δ_d is $\delta^{13}C$ of dissolved CO_2 ($\delta^{13}C_{aq}$), which we can derive from the $\delta^{13}C$ of bulk carbonate.

We interpolated the bulk carbonate carbon isotope record (Fig. S3) using the Analyseries (Paillard et al., 1996) software to fit the resolution of the carbon isotope record to that of the alkenone record. The equation to estimate ε_p with use of carbonate $\delta^{13}C$ uses foraminiferal $\delta^{13}C$. Unfortunately, only bulk $\delta^{13}C$ was available at our site, which composes of calcareous nannoplankton. We added 0.5‰ to bulk carbonate $\delta^{13}C$ to correct for differential biotic fractionation between foraminiferal calcite and calcareous nannoplankton calcite (Pagani 2002). The carbon isotopic fractionation factors of from biogenic calcite ($\delta^{13}C_{calc}$) to gaseous CO_2 ($\delta^{13}C_{CO_{2(g)}}$) are required to calculate $\delta^{13}C_{aq}$.

$$\varepsilon_{calc-CO_{2(g)}} = 11.98-(0.12*SST) \tag{6.5}$$

And that from $CO_{2(g)}$ to $CO_{2(aq)}$:

$$\varepsilon_{CO_{2(g)}-CO_{2(aq)}} = -373/(273.15+SST)+0.19 \tag{6.6}$$

SST was obtained from the organic paleothermometer proxies TEX_{86} and $U^{K'}_{37}$. Differences in absolute SST reconstructions derived from these proxies are included as error bars (Fig. 2; Fig. S3; Table S3).

To calculate $\delta^{13}C_{CO_{2(g)}}$ from $\delta^{13}C_{calc}$ we use;

$$\delta^{13}C_{CO_{2(g)}} = ((\delta^{13}C_{calc}+1000)/(1+\varepsilon_{calc-CO_{2(g)}}*0.001))-1000 \tag{6.7}$$

This $\delta^{13}C_{CO_{2(g)}}$ is then used to calculate the $\delta^{13}C_{CO_{2(aq)}}$ following;

$$\delta^{13}C_{aq} = ((0.001*\varepsilon_{CO_{2(g)}-CO_{2(aq)}}+1)*(\delta^{13}C_{CO_{2(g)}}+1000))-1000 \tag{6.8}$$

3.4.2. Calculation of $[CO_{2(aq)}]$

The estimation of $[CO_{2(aq)}]$ from ε_p follows that of Pagani et al. (2005):

$$\varepsilon_p = \varepsilon_f - b/[CO_{2(aq)}] \tag{6.9}$$

where ε_f represents the carbon isotopic fractionation due to carboxylation (which is ~25‰ in the modern day ocean (Pagani et al., 2002; Popp et al., 1998)), and b represents the sum of all physiological factors, such as growth rate and cell geometry. These physiological factors are strongly correlated to surface water $[PO_4^{3-}]$ because this is one of the limiting nutrients, or it serves as a proxy for other limiting nutrients.

The simplified term b is related to phosphate via the following equation (Pagani et al., 2005):

$$b = [(118.52[PO_4^{3-}]) + 84.07] / (25-\varepsilon_p) \qquad (6.10)$$

In our experiment, to assess the influence of varying phosphate concentrations, we allowed $[PO_4^{3-}]$ to vary throughout the record as a function of the abundance of peridinioid organic walled dinoflagellate cyst taxa in the same samples. The middle Eocene record at ODP Site 1172 shows large fluctuations in the abundance of peridinioid cysts (Röhl et al., 2004a; Fig. 2), and we therefore allow $[PO_4^{3-}]$ to vary between 0.1±0.1 and 0.9±0.1 as a function of the percentage peridinioid cysts (see discussion under Section 4).
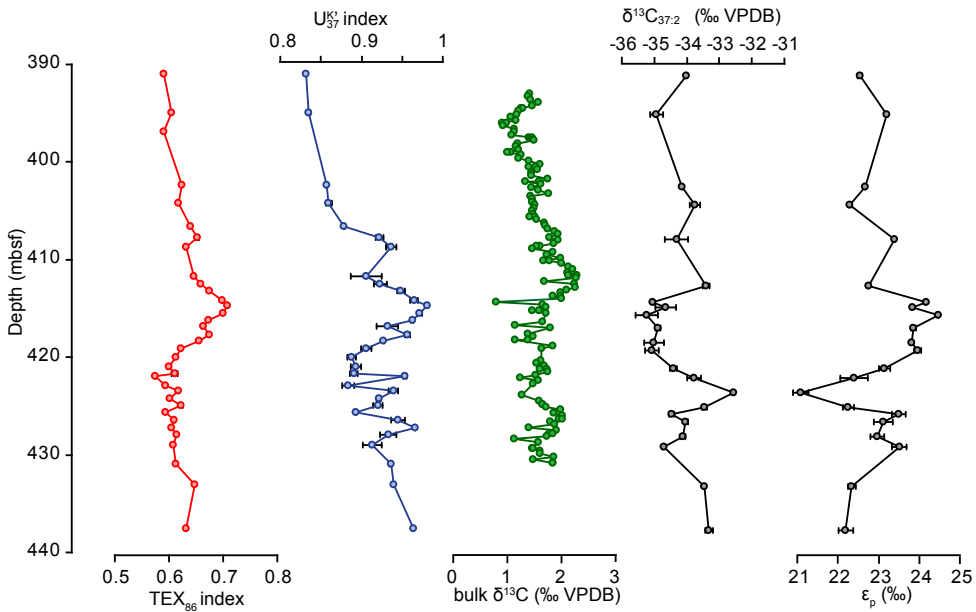
Figure S3. Index values for TEX$_{86}$ and U$_{37}^{K\prime}$, carbon isotope values for the bulk carbonate at ODP Site 1172 and for the 37:2 di-unsaturated alkenones, fractionation factor ε_p, all in per mill and versus Vienna Pee Dee Belemnite.

3.4.3. Calculation of pCO$_2$

The conversion of $[CO_{2(aq)}]$ to pCO$_2$ uses Henry's Law (Weiss 1974):

$$p\text{CO}_2 = [CO_{2(aq)}] / 10\text{^}[-60.2409+93.4517*(100/SST)+$$
$$23.3585*\ln(SST/100)+SSS*(0.023517-0.023656*(SST/100)+$$
$$0.0047036*((SST)/100)*(SST/100))] \qquad (6.11)$$

We assume a sea surface salinity (SSS) of 35 p.s.u. (Pagani 2002) and SSTs derived from TEX$_{86}$ and U$_{37}^{K'}$ (converted to the Kelvin scale). Because the temperature proxies disagree on absolute SSTs prior to the MECO, the error for reconstructed pCO$_2$ is higher in that interval (Fig. 2).

4. Discussions

4.1. Dinoflagellate response to MECO warming

Dinocyst assemblages in the middle- and late Eocene Southern Ocean are characterized by dominance of taxa that are biogeographically restricted (endemic) to the circum-Antarctic region (Chapter 3; Wrenn and Beckman 1982). In this time period, the relative abundance and distribution of endemic dinocysts is found to closely relate to SST changes as reconstructed by TEX$_{86}$ (Chapter 3). This is now particularly well recognized in the MECO record at ODP Site 1172, where we find that endemic dinocyst assemblages declined in abundance at peak MECO warming (Fig. S4). Biogeographically cosmopolitan taxa (with no apparent relation with temperature) show no large variation in abundance associated to MECO. However, a marked increase in low latitude taxa (Fig. 2; Fig. S4, Table S1) provides evidence for biotic response in consequence of MECO warming. Apparently, when the Southern Ocean surface waters warmed, conditions began to approach those of lower latitudes. The protoperidinioid dinocyst taxa dominate Southern Ocean assemblages from the Oligocene onwards (Kemp 1975; Mohr 1990), but are only present in low abundance during MECO times (see also discussions under section 4.3).

4.2. The effect of light-limited growth on pCO$_2$ estimates

An important factor that influences ε_p in alkenones from the main modern alkenone producer *Emiliania huxleyi* grown in incubation studies and chemostat experiments is light intensity (Rost et al., 2002). In these experiments, the carbon isotopic fractionation of alkenones varies up to 2 per mille when light intensity acts a growth-limiting factor during carbon fixation (Rost et al., 2002). In incubation studies, this effect is more pronounced under nutrient-repleted conditions (Rost et al., 2002). In theory, natural environments with eutrophic surface waters would particularly experience light limitation, as high surface-water productivity causes limitations for light-penetration. A modeling experiment suggested that the effect of coccolith blooms on light penetration is considerable, particularly because the white surfaces of the coccolith platelets act as a reflective surface (Tyrrell et al., 1999). Following this model, the light intensities used in the experiments of Rost et al. (2002); 15 to 150 μmole m^{-2} s^{-1}) assume that coccolith concentrations increased from 0 to 300 mg CaCO$_3$ m^{-3}. Critically, this is equivalent to a transition from no coccoliths to a full bloom, and may be considered an extreme situation. Along with the notion that the above experiment describes an extreme case, several other points compromise the hypothesis that changes in the magnitude of light-limited growth forced the recorded changes in ε_p. First of all, the magnitude of light-limited growth as recorded in the laboratory is not easily extrapolated to the natural environment. The relation between ε_p, phosphate/nutrients and pCO$_2$ has been shown in a number of environments (Benthien et al., 2002; 2005), and in more recent geological times (Palmer et al., 2010;

Seki et al., 2010) while no relation with light limitation and ε_p has, at this point, been reported for the natural environment.

pCO₂ lower	1709	2598	1976	1685	2947	2458	3857	2438	5560	2981	2674	3597	2217	1465	1010	1337	2282	1935	1854	2524	1860	873
pCO₂ upper	2357	3423	2549	2241	3985	3115	5680	3451	8251	4611	4142	5845	3473	2806	1568	2080	4017	3553	3032	4514	2602	1608
pCO₂ average	2022	3000	2259	1956	3449	2779	4762	2950	6918	3763	3384	4652	2808	2082	1273	1690	3087	2682	2406	3445	2214	1219
$[PO_4^{3-}]$ (±0.1)	0.56	0.65	0.59	0.60	0.62	0.77	0.20	0.04	0.13	0.30	0.24	0.45	0.58	0.59	0.49	0.41	0.43	0.48	0.47	0.52	0.68	0.10
ε_p error	0.06	0.03	0.01	0.03	0.04	0.01	-0.01	0.00	0.00	0.06	0.05	0.09	0.13	0.34	0.19	0.14	0.16	0.23	0.16	0.18	0.09	0.18
ε_p average	22.54	23.21	22.68	22.30	23.38	22.76	24.17	23.84	24.47	23.86	23.82	23.97	23.15	22.39	21.08	22.26	23.49	23.12	22.97	23.51	22.34	22.19
$\varepsilon_p\ U_{37}^{K'}$	22.60	23.23	22.68	22.33	23.42	22.77	24.16	23.84	24.47	23.91	23.86	24.06	23.28	22.74	21.27	22.40	23.66	23.35	23.13	23.70	22.43	22.38
$\varepsilon_p\ TEX_{86}^{H}$	22.48	23.18	22.67	22.26	23.35	22.75	24.18	23.84	24.47	23.80	23.77	23.88	23.01	22.05	20.88	22.12	23.33	22.89	22.81	23.33	22.25	22.01
$U_{37}^{K'}$ SST	23.86	24.02	24.68	24.84	26.47	26.40	27.76	28.35	28.03	27.30	26.80	25.91	25.54	27.63	27.31	26.72	25.74	27.55	26.64	26.69	27.13	27.89
TEX_{86}^{H} SST	22.97	23.62	24.59	24.29	25.90	26.21	27.93	28.36	28.02	26.43	26.06	24.48	23.41	22.18	24.28	24.53	23.13	23.87	24.11	23.78	25.70	24.97
$\delta^{13}C_{calc}$	2.090	1.739	1.947	1.983	2.260	2.578	2.071	2.089	2.150	2.040	1.932	2.168	2.136	2.012	1.867	2.124	2.427	2.353	2.153	2.103	2.100	2.100
$\delta^{13}C_{37:2}$ error	0.015	0.202	0.018	0.156	0.349	0.117	0.016	0.315	0.353	0.075	0.298	0.211	0.110	0.208	0.020	0.083	0.071	0.069	0.083	0.061	0.012	0.113
Average $\delta^{13}C_{37:2}$	-34.016	-34.934	-34.136	-33.755	-34.320	-33.407	-35.044	-34.654	-35.231	-34.895	-35.015	-35.076	-34.416	-33.775	-32.562	-33.454	-34.468	-34.039	-34.126	-34.704	-33.464	-33.319
$\delta^{13}C_{37:2}$ measurement 2	-34.026	-34.791	-34.149	-33.865	-34.566	-33.324	-35.033	-34.431	-34.981	-34.948	-34.804	-35.225	-34.338	-33.628	-32.576	-33.512	-34.518	-34.087	-34.067	-34.661	-33.472	-33.399
$\delta^{13}C_{37:2}$ measurement 1	-34.005	-35.076	-34.123	-33.644	-34.073	-33.489	-35.055	-34.876	-35.480	-34.842	-35.225	-34.926	-34.494	-33.922	-32.548	-33.395	-34.417	-33.990	-34.185	-34.747	-33.455	-33.239
Depth (mbsf)	391.00	394.92	402.37	404.20	407.72	412.52	414.12	414.68	415.52	416.80	418.30	419.12	420.92	421.90	423.40	424.90	425.62	426.40	427.90	428.96	433.00	437.50
Interval (cm)	10-12 cm	42-44 cm	37-39 cm	10-12 cm	62-64 cm	32-34 cm	32-34 cm	98-100 cm	22-24 cm	10-12 cm	10-12 cm	92-94 cm	122-124 cm	10-12 cm	10-12 cm	10-12 cm	82-84 cm	10-12 cm	10-12 cm	114-116 cm	10-12 cm	10-12 cm
Section (type)	6W	2W	5W	2W	4W	1W	2W	2W	3W	4W	5W	5W	6W	1W	2W	3W	3W	4W	5W	5W	2W	5W
Core (type)	42X	43X	43X	44X	44X	45X	45X	45X	45X	45X	45X	45X	45X	46X	46X	46X	46X	46X	46X	46X	47X	47X
Site (Hole)	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A	1172A

Table S3. $\delta^{13}C_{37:2}$, $\delta^{13}C_{calc}$, ε_p, phosphate assumptions and inferred pCO₂ values. $\delta^{13}C_{calc}$ was interpolated from the bulk organic carbon isotope data presented in Fig. S3.

Second, coccolithophores generally occur in blooms, and hence always produce their 'own' light-limitation. The change in light penetration throughout our record thus should result from the abundance of other particles in the surface ocean.
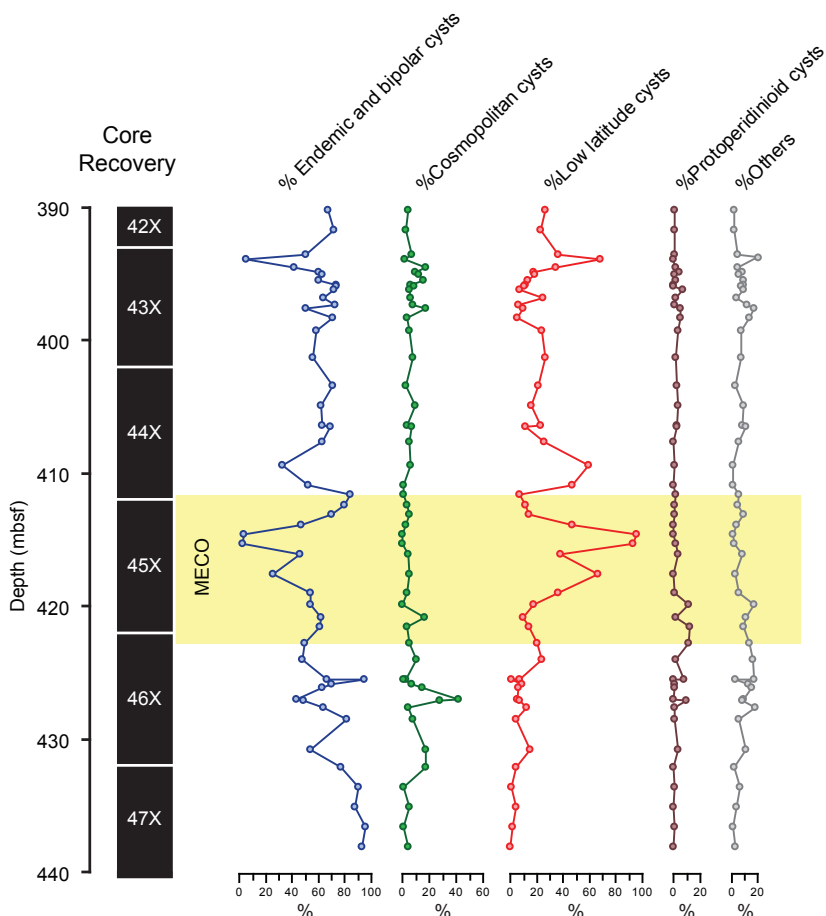
Figure S4. Dinoflagellate cyst assemblages grouped as follows (see Table S1): endemic and bipolar cysts; cosmopolitan (in terms of latitude) cysts, low latitude cysts; protoperidinioid cysts and others.

The only indications we have for changes affecting surface water turbidity are relative nutrient changes, based on e.g., dinocyst assemblages (Fig. 2; Röhl et al., 2004a); the more nutrients are available in the surface waters, the more particles reside in the water, and the higher the light limitation. A shift towards more oligotrophic settings is interpreted at the MECO. That would prescribe a decrease of the effect of light limitation on the ε_p values (Rost et al., 2002). Although this effect might explain part of the increased ε_p values at the MECO, we argue that it cannot explain the total change we observe. ODP Site 1172 is not comparable to an oligotrophic gyral system but was situated on an extended flat shelf, with a proximal-

distal nutrient gradient. Diatoms, abundant dinocysts and calcareous nannofossils are present in MECO sediments (Exon et al., 2001; Röhl et al., 2004a; Exon et al., 2004), and indicate more eutrophic conditions than central gyres. Compared to pre-MECO environments, however, nutrient concentrations decreased as interpreted from the dinocysts (Fig. 2). Light penetration in these settings strongly depends on the amount of particulate matter dwelling in the surface waters. It is unlikely that this factor changed dramatically throughout MECO times at the ETP, certainly not to the extent simulated in the incubation experiments. Hence the effect can only explain a minor part of the 3‰ change we observe associated with the MECO. Third, ODP Site 1172 was situated at a paleolatitude of 65°S. The growing season (i.e. austral summer) of this site experiences conditions with near-continuous light. Rost et al. (2002) showed that under these conditions, the effect of light limited growth was smaller compared to experiments with a light-dark cycle.

In summary, the effect of light-limited growth has not been shown to have a major effect on ε_p values in environments similar to the paleo-setting at ODP Site 1172 as much as in incubation experiments. Although we cannot exclude an effect of light limited growth in our $p\mathrm{CO_2}$ record, it is likely that the effect is relatively minor, and cannot be held responsible for the large magnitude of the ε_p change.
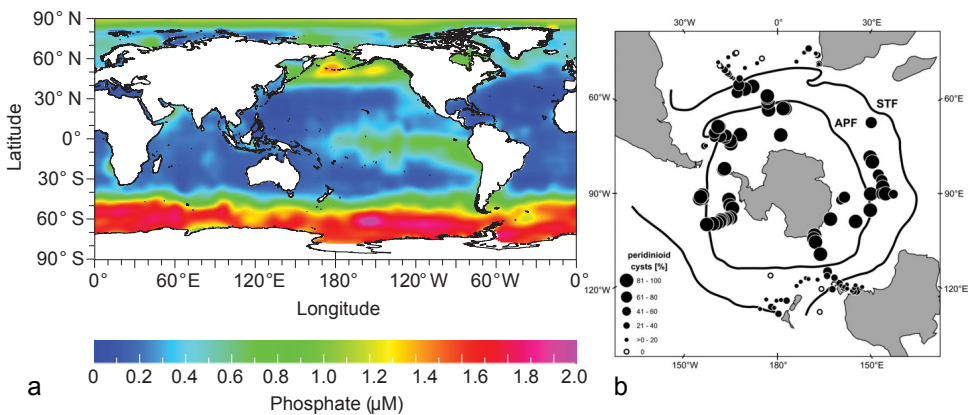
Figure S5. a: the mean annual concentration of surface water soluble phosphate from the World Ocean Database (Garcia et al., 2006) modified from Pagani et al., (2010). b: The relative abundance of peridinioid dinocysts in surface sediments. Figure obtained from Esper and Zonneveld (2007).

4.3. Phosphate concentrations and dinocysts
In our study, we derive concentration changes of sea surface soluble phosphate by using dinocysts preserved in sediments. The present day Southern Ocean has a steep lateral gradient of phosphate in the surface waters (Fig. S5A). Strong Antarctic circumpolar currents (the ACC) promote mixing of surface waters with deeper water, thereby constantly replenishing the photic zone with phosphate. Dinocyst assemblages within the ACC almost exclusively consist of protoperidinioid

taxa, which are produced by heterotrophic dinoflagellates. Outside the ACC, phosphate concentrations are significantly lower (Fig. S5a), and dinocyst assemblages are dominated by gonyaulacoid taxa. Esper and Zonneveld (2007) made a survey of the distribution of peridinioid dinocysts versus gonyaulacoid dinocysts (P/G ratio) within surface samples throughout the Southern Ocean, and identified the coherence between the ACC and the percentage of peridinioid dinocysts (Esper and Zonneveld 2007; Fig. S5b). The relationship between the abundance of protoperidinioid dinocysts and nutrients was also recognized in Holocene sediment cores from the Arabian Sea (Reichart and Brinkhuis 2003). Present day Southern Ocean protoperidinioid dinocysts share the same lineage with the Eocene peridinioid dinocysts (Fensome et al., 1993). Furthermore, a wealth of multi-proxy investigations suggests a relation between the abundance of peridinioid dinocysts and nutrient availability throughout the Cenozoic (Sluijs et al., 2005) With use of the Analyseries (Paillard et al., 1996) software, we interpolated the P/G ratio record to fit the resolution with the alkenone record.

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Cyst bilaterally asymmetrical with a thick-walled endocyst. The pericyst is
conspicuously folded, especially on the ventral surface, the folds extending
from the apex to antapical horns. The periphragm is granulate. The
archeopyle is more or less trapezoidal with the operculum remaining
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NEDERLANDSE SAMENVATTING

Hystrichosphaeridium truswelliae Wrenn and Hart, 1988
Subspherical to spherical skolochorate dinocysts that bear intratabular processes of various lengths. The apical processes are always the shortest; consequently, the endocyst is asymmetrically placed with respect to the imaginary shell formed by joining the process tips with a line. The hollow processes are distally open, flared, and fenestrate at process terminations. Apical processes shorter than all other processes. Periphragm is smooth, whereas endophragm is granular.

Met het begin van de industriële revolutie, ongeveer 150 jaar geleden, is een ongeëvenaard snelle stijging van de atmosferische druk van koolstofdioxide (pCO_2) begonnen. Blijven de CO_2 emissies onverminderd stijgen, dan wordt verwacht dat pCO_2 waarden rond het begin van de 21^e eeuw 1.000 deeltjes per miljoen in volume (ppmv) bereikt: veel hoger dan de ~390 van vandaag de dag, laat staan de ~280 van vlak voor de industrialisatie (IPCC, 2007). Het broeikasgas CO_2 absorbeert de uitgaande langegolfstraling (warmte) en straalt het naar alle richtingen uit, waaronder naar de aarde. Hogere CO_2 waarden zullen daarom naar verwachting resulteren in hogere oppervlaktetemperaturen. Bovendien is de verblijftijd van de uitgestoten CO_2 lang genoeg om de Aarde de opwarmingseffecten te doen ervaren: de CO_2 blijft zeker 100.000 jaar in het oceaan-atmosfeersysteem zitten (Walker en Kasting, 1992; Dickens, 2003; Archer, 2003). De klimaatrespons is reeds meetbaar bij veldobservaties (IPCC, 2007). Computermodelexperimenten worden uitgevoerd om de toekomstige klimaat- en milieu effecten gerelateerd aan de aangenomen verhoogde pCO_2 te voorspellen. Een basale aanname in deze computermodellen is het actualiteitsprincipe: de klimaats-forceringen en terugkoppelingen die heden ten dage aan de orde zijn, zullen ook gelden voor de toekomst. De vraag is alleen of deze forceringen en terugkoppelingen wel hetzelfde zijn onder verschillende klimaatscondities. Een manier om de betrouwbaarheid van de computer-modelvoorspellingen te testen is om te kijken in het geologische verleden, toen er condities heersten vergelijkbaar met die voor de toekomst worden verwacht. Reconstructies van de gevoeligheid en snelheid van aardse veranderingen bij perturbaties in het klimaatsysteem kunnen inzicht verschaffen voor toekomstige klimaatveranderingen.

Om een wereld te bestuderen met pCO_2 die onze toekomstige benadert, kijken paleoklimatologen ver terug in de tijd. De beschikbare informatie geeft aan dat over de laatste 35 miljoen jaar pCO_2 nooit boven de 1.000 ppmv is geweest. Verreweg de meeste tijd was pCO_2 zelfs onder de 500 ppmv (Beerling en Royer 2011). De tijdsperiode voor 35 miljoen jaar geleden wordt gekenmerkt door pCO_2 waarden die waarschijnlijk veel hoger dan 1.000 ppmv zijn geweest. Het bewijs voor veel warmere klimaatscondities is vooral overtuigend in poolgebieden. Vegetatiereconstructies van noordelijke hoge breedtegraden laten zien dat continenten rondom de Noordelijke IJszee tijdens het tijdvak Eoceen (~56 tot 34 miljoen jaar geleden) waren begroeid met (onder andere) *Metasequoia*, een boomsoort die nu alleen in gematigd regenwoud groeit (Francis, 1988). Krokodilschedels zijn gevonden in Alaska (Markwick, 1998), en pootafdrukken bewijzen dat voorvaderen van het Nijlpaard moerassen bevolkten op wat nu Spitsbergen is. Het is zelfs aangetoond dat sommige van deze zoogdieren het hele jaar door vertoefden in het noordelijke poolgebied (e.g., Eberle et al., 2009). Tijdens het toppunt van de Eocene warme periode, ongeveer 54 miljoen jaar geleden, groeiden er palmen rond de Noordelijke IJszee (Sluijs et al., 2009a), en rond 50 miljoen jaar geleden was de Noordelijke IJszee bedekt onder een dikke laag van de drijvende zoetwatervaren *Azolla* (Brinkhuis et al., 2006; Collinson et al., 2010). Het laatstgenoemde fenomeen, gecombineerd met de gereconstrueerde vegetatietypes op het omringende land doet vermoeden dat neerslag in ieder geval seizoenaal nat moet zijn geweest op noordelijke hoge breedtegraden (Pagani et al., 2006; Speelman et al., 2010, Barke et al., 2011).

Op zuidelijke hoge breedtegraden zijn Paleogene macrofossielen gevonden in ontsluitingen op het Antarctische continent, die de aanwezigheid van loofbos bewijst (Francis and Poole, 2002). Biomen op Nieuw Zeeland (toentertijd 50° Zuid; Crouch, 2001)) bevatten een keur aan palmsoorten in het laat Eoceen (Hartwich et al., 2010). Studies aan kleimineraalcomposities in Eocene kustsedimenten wijzen op chemische verwering onder invloed van regen in het Antarctische achterland (Ehrmann 1998). De ontdekking dat Antarctica nat en warm was, is opvallend omdat polaire gebieden ook in het Eoceen donker zijn voor de helft van het jaar. Om de Eocene warmte op polaire breedtegraden te verklaren werd een hypothese geopperd met de nadruk op het effect van de toenmalige continentenconfiguratie op het Zuidelijk Halfrond. Omdat Australië tijdens het grootste deel van het Eoceen nog min of meer aan Antarctica vastzat (Lawver en Gahagan, 2003), en ook de doorgang tussen Antarctica en Zuid Amerika nog niet open was (Lagabrielle et al., 2009), kon de –vandaag de dag isolerende- Antarctische circumpolaire stroming niet rond Antarctica stromen. In plaats daarvan zouden afgeleiden van warme, lage-breedtegraden-stromingen de Antarctische kustlijn hebben kunnen bereiken en opwarmen (Kennett 1977). Deze hypothese werd later in twijfel getrokken door resultaten van computer-modelexperimenten die oceaancirculaties simuleren (Sloan en Rea, 1995). Toepassingen van deze computer modellen op de Eocene situatie laten zien dat zelfs in afwezigheid van een isolerende circumpolaire stroming, de warme oceaanstromingen van lage breedtegraden niet de Antarctische kustlijn konden bereiken. In plaats daarvan circuleerde het oppervlakte water in de vorm van grote kloksgewijs circulerende stromingen, die het bereiken van de Antarctische kustlijn door warme stromingen effectief voorkwam (Sloan en Rea, 1995; Huber et al., 2004). De nu leidende hypothese die de warme condities in het vroeg Paleogeen verklaart dicht een belangrijke rol toe aan de hoge concentraties broeikasgassen, en de daarmee geassocieerde terugkoppelingen (Zachos et al., 2008) alhoewel voldoende direct bewijs, die deze hypothese ondersteunt, nog ontbreekt.

Vanaf 46 miljoen jaar geleden worden in de Arctische Oceaan *ice-rafted debris* en zee-ijs diatomeeën afgezet (Eldrett et al., 2007; St. John 2008; Stickley et al., 2009), wanneer Paleogene klimaatscondities hun topwarmte zijn gepasseerd. Desondanks wordt over het algemeen aangenomen dat de temperaturen en pCO$_2$ concentraties in die tijd nog steeds te hoog waren om de aanwezigheid van grote hoeveelheden ijs op de polen toe te staan (DeConto et al., 2008). Op het Zuidelijk Halfrond laten resten van planktonische dinoflagellaten een opvallende verandering zien naar dominantie van vormen die endemisch zijn voor de zuidelijke hoge breedtegraden (Wrenn en Beckman, 1982; Brinkhuis et al., 2003a; Warnaar, 2006). Intuïtief doet zo'n endemische dinoflagellatenpopulatie een afkoeling van het oppervlaktewater in de Zuidelijke Oceaan vermoeden. Dit is niet in overeenstemming met de gereduceerde meridionale temperatuur-gradiënten zoals is gereconstrueerd met behulp van de randen van fossiele bladeren (Greenwood en Wing, 1995), wat eerder de migratie poolwaards dan het ontstaan van endemisme zou faciliteren. Bovendien, door de afwezigheid van fysieke barrières die de Zuidelijke Oceaan afscheidt van aangrenzende wateren, staat letterlijk niets in de weg voor de migratie van warme dinoflagellatensoorten poolwaards.

Algemeen bezien, terwijl een keur aan bewijs suggereert dat het vroeg Eoceen warm was, en het jongere deel van het Eoceen een afkoeling onderging richting de 'IJskastwereld' van het Oligoceen en Neogeen, blijven nog steeds veel vragen bestaan, over de timing, gedrag en oorzaken van de Broeikas-IJskast overgang. Een betere kennis van de klimaatsevolutie in de Zuidelijke Oceaan begint bij continue, chrono-stratigrafisch goed opgeloste tijdseries van vroeg-Paleogene temperatuur, klimaat- en biotische veranderingen. De opgeboorde sequentie van de Arctic Coring Expedition (Moran et al., 2006) heeft een uniek eerste inzicht gegeven in paleo-ecologische veranderingen tijdens de Paleogene broeikaswereld op de Noordpool, maar de sequentie is gefragmenteerd, van lage resolutie en met grote hiaten.

Het sedimentaire archief van de Zuidelijke Oceaan daarentegen is veelvuldig aangeboord door het *Deep Sea Drilling Project* (DSDP) en later het *Ocean Drilling Program* (ODP) en het nog lopende *Integrated Ocean Drilling Program* (IODP). Een probleem die de bestudering van deze boorkernen bemoeilijkt is de afwezigheid van voldoende biostratigrafische middelen om de sedimenten te dateren: conventionele, uit calciet bestaande microfossielen zijn typisch afwezig in sedimenten van hoge breedtegraden. Deels door die afwezigheid weten we nog maar weinig van de klimaatsevolutie in de Zuidelijke Oceaan ten tijde van het vroeg Paleogeen.

Een indirecte benadering om de Paleogene klimaatsevolutie van de Zuidelijke Oceaan te evalueren is opgeslagen in de compilatie van zuurstofisotopenanalyses aan benthische foraminiferen ($\delta^{18}O$; Zachos et al., 2008). De $\delta^{18}O$ waarde in benthische foraminiferen wordt vooral bepaald door watertemperatuur-veranderingen en veranderingen in de $\delta^{18}O$ van het zeewater. Dit laatste wordt beïnvloedt door ijs-volume en saliniteit. In de afwezigheid van continentale ijskappen wordt zeewater $\delta^{18}O$ dus vooral bepaald door saliniteitsschommelingen. De saliniteit en temperatuur van diepere wateren is afhankelijk van de bron en mechanismes van diep water formatie. De Zuidelijke Oceaan wordt algemeen gezien als de belangrijkste bron voor diepwater formatie in het Eoceen (Thomas et al., 2003). Dus, het signaal uit benthische foraminiferen die leefden in de diepere wateren bevat in feite de signatuur van de klimaatsevolutie van oppervlaktewateren van de Zuidelijke Oceaan.

De globale benthische $\delta^{18}O$ compilatie (Zachos et al., 2008) laat een algemene afkoeling zien vanaf de Krijt-Tertiair grens, naar een minimum in het mid-Paleoceen (~58 miljoen jaar geleden; Zachos et al., 2008) Daaropvolgend stijgen de temperaturen gradueel tot ~52 miljoen jaar geleden, waar lange-termijn piektemperaturen voor het Cenozoicum worden bereikt tot ~50 miljoen jaar geleden: het *Early Eocene Climatic Optimum* (EECO). Hier bereiken diepere wateren temperaturen 13°C warmer dan vandaag de dag. De daaropvolgende 15 miljoen jaar, van 49 tot 34 miljoen jaar geleden worden gekenmerkt door graduele afkoeling van diepere wateren (Zachos et al., 2008), wat uiteindelijk uitmondt in het begin van het ontstaan van grootschalige glaciaties op Antarctica (Zachos et al., 1992), ongeveer 34 miljoen jaar geleden (Zachos et al., 2008). Omdat de aanwezigheid van continentale ijskappen voorafgaand aan het vroeg Oligoceen nog steeds ter discussie staat (Eldrett et al., 2007; St. John et al., 2008) is het vooralsnog onzeker hoeveel van de $\delta^{18}O$ stijging

in het midden en laat Eoceen het gevolg is van afkoeling, hoeveel van ijsopbouw en hoeveel van een verandering in de bron voor diep water formatie.

Tijdens deze trends hebben verscheidene periodes van klimaatverandering plaats gehad met een meer kortdurend, *transient* karakter. Het *Paleocene-Eocene Thermal Maximum* (PETM) is een kortdurend (~150 kyrs) interval met 5 tot 8°C opwarming, zoals gedocumenteerd in sedimentaire opeenvolgingen van de tropen en in midden- en hoge breedtegraden op beide halfronden (Kennett en Stott, 1991; Sluijs et al., 2006; zie Sluijs et al. (2007a) voor een samenvatting). Tijdelijke opwarming tijdens deze periode valt samen met een duidelijke negatieve excursie in de stabiele koolstofisotopen compositie in sedimentaire componenten (Kennett en Stott, 1991; Sluijs et al., 2007a). Dit bewijst een input van ^{13}C-verarmd koolstof in het wereldwijde exogene koolstofreservoir. Omdat wijdverbreid bewijs bestaat dat de hoeveelheid koolstof die toegevoegd is tijdens de PETM van dezelfde orde van grootte is als aangenomen wordt voor onze toekomst als alle fossiele brandstof wordt opgebruikt, wordt deze periode veel gezien als beste analoog voor toekomstige klimaatsveranderingen. Jongere, minder extreme periodes met veel gelijkenissen met de PETM zijn ontdekt in het tijdsinterval vanaf de PETM (56 miljoen jaar geleden) tot 51 miljoen jaar geleden (bijv. Cramer et al., 2003; Nicolo et al., 2007; Stap et al., 2010; Zachos et al., 2010). Deze zogenoemde '*hyperthermals*' hebben verschillende acroniemen toegewezen gekregen en worden, net als de PETM, gekenmerkt door een shift naar lagere $\delta^{13}C$ waarden van het exogene koolstofreservoir.

Een nog jongere, prominente verschijning van relatief kortdurende klimaatopwarming is het *Middle Eocene Climatic Optimum* (MECO; Bohaty en Zachos, 2003), wat rond 39 miljoen jaar geleden plaats vond (Bohaty et al., 2009). Net als de PETM en andere *hyperthermals*, valt de MECO in diepzeesequenties op door de carbonaatoplossing, welke intuïtief toe te schrijven is aan perturbaties van de koolstofcyclus. Verschillen met de *hyperthermals* zijn er ook, namelijk dat de MECO niet wordt gekarakteriseerd door een negatieve $\delta^{13}C$ excursie. Dit doet vermoeden dat, mocht CO_2 stijging een rol spelen bij de klimaatverandering tijdens MECO, de oorsprong van die koolstof die in de atmosfeer gekomen is aanmerkelijk van andere aard is dan bij de *hyperthermals*.

In dit proefschrift komen een aantal nog openstaande vragen aan de orde om beter begrip te krijgen van de klimaatsevolutie in de Zuidelijke Oceaan, zoals: hoe zijn de oppervlaktewateren geëvolueerd in het vroeg-Paleogeen, vooral in vergelijking met de klimaatsevolutie zoals gereconstrueerd in the benthische isotopen? Hoe gevoelig waren de oppervlaktewateren in de Zuidelijke Oceaan voor klimaats-oscillaties op lange en korte termijn in het vroeg-Paleogeen? Wat was de biotische respons op temperatuur- en circulatieveranderingen? Wat was precies de rol van CO_2, en wat van de opening en de daaropvolgende verdieping van de *Tasmanian Gateway*? Wat was de klimaatgevoeligheid van de Zuidelijke Oceaan in de Paleogene Broeikaswereld?

Om deze vragen te beantwoorden is een combinatie van micro-paleontologische en organisch geochemische technieken toegepast die uitermate bruikbaar bleek om klimaatverandering te reconstrueren op hoge breedtegraden. Anders dan de uit calciet bestaande microfossielen, zijn de fossiele resten van de in het oppervlaktewater levende dinoflagellaten, de dinoflagellaten 'cystjes', of dinocysten, wel zeer talrijk aanwezig in hoge breedtegradensedimenten. In de meeste gevallen, zeker bij zeer oude sedimenten, zijn de dinoflagellaten cysten zelfs het enig overgebleven biostratigrafische hulpmiddel, om de sedimenten met andere opeenvolgingen te kunnen correleren (Brinkhuis et al., 2003a; Eldrett et al., 2004; Williams et al., 2004). Bovendien bevatten assemblages van dinocysten bruikbare, kwantitatieve informatie over een keur aan fysische parameters van de oppervlaktewateren, zoals nutriënten, temperatuur, saliniteit en kustnabijheid (Dale 1996; Pross en Brinkhuis, 2005). Specifiek voor de Zuidelijke Oceaan ten tijde van het vroeg Paleogeen is de aanwezigheid van endemische vormen van dinocysten (Wrenn en Beckman 1982; Wrenn en Hart, 1988). Deze soorten werden aanvankelijk *transantarctic flora* genoemd, omdat aangenomen werd dat ze bewijs vormden voor de aanwezigheid van een transantarctische zeestraat (Wrenn en Beckman, 1982), en omdat blijkbaar in die tijd, dinoflagellaten onder de planten werden gerekend. Hoewel latere studies aantoonden dat dinoflagellaten zowel heterotrofe alsook autotrofe genera herbergt, en het vooralsnog niet evident is dat een transantarctische zeestraat daadwerkelijk bestond, wordt de term *transantarctic flora* desondanks toch nog gebruikt. Het werd duidelijk dat de *transantarctic flora* domineerde in die regio's die onder invloed waren van oceaanstromingen vanuit Antarctica (Huber et al., 2004). Deze observatie wijst intuïtief naar temperatuur als bepalende omgevingsfactor voor de biogeografische distributie van de endemische populatie, maar deze aanname is nog niet adequaat getest.

Bovenop de informatie verkregen door analyses van dinocysten kunnen organisch geochemische tools vaak worden ingezet op sedimenten die dinocysten bevatten, omdat de condities die nodig zijn om beide te preserveren in het sediment vrijwel gelijk zijn. De organisch geochemische *toolbox* om klimaat te reconstrueren is de afgelopen jaren erg vergroot, met de ontwikkeling van nieuwe methodes, zoals de TEX_{86} proxy voor zeeoppervlaktewatertemperatuur (Schouten et al., 2002; Kim et al., 2010) en pCO$_2$ gebaseerd op de stabiele koolstofsamenstelling van alkenonen (Pagani 2002). Door de palynologische en organisch geochemische tools te combineren op de grote hoeveelheid sedimentaire archieven uit de Zuidelijke Oceaan kan een betere reconstructie ontstaan van de dynamiek en hoge-resolutie klimaatevolutie van vroeg Paleogene zuidelijke hoge breedtegraden.

Dit proefschrift levert nieuwe inzichten in de paleomilieu- en paleoklimaatevolutie van de Zuidelijke Oceaan ten tijde van het vroeg Paleogeen. In **Hoofdstuk 1, 2** en **3** wordt een raamwerk gepresenteerd van de paleoklimaat en paleomilieuevolutie in de Zuidelijk Oceaan, in termen van dinocyst bio-magnetostratigrafie, organische *biomarker* paleothermometrie en dinocysten biogeografie in de Zuidelijke Oceaan. De **Hoofdstukken 4, 5** en **6** richten zich specifiek op kritieke intervallen met snellere klimaatverandering, in chronologische volgorde.

In **Hoofdstuk 1** wordt een aan magnetostratigrafie gekalibreerd biostratigrafisch zonatieschema van organische dinoflagellaten cysten voorgesteld voor boven Paleocene tot boven Eocene sedimenten van de Zuid-Pacifische Oceaan. We vergelijken chronologieën van Ocean Drilling Program (ODP) Holes 1171D en 1172A/D met dat van Integrated Ocean Drilling Program (IODP) Hole U1356A, en vervolgens met dinocysten zonaties van Nieuw Zeeland. Gebaseerd op deze correlaties presenteren wij een gereviseerd magnetostratigrafisch ouderdomsmodel voor ODP Holes 1171D en 1172A/D en stellen een Zuid-Pacifische dinocysten zonatie voor, bestaande uit dertien dinocysten zones. De *South Pacific Dinocyst Zonation* is gekalibreerd naar de Internationale Tijdschaal van Gradstein et al. (2004), met toegevoegde onafhankelijke ouderdomsbepalingen uit stabiele koolstofisotopen profielen, zoals die gecorreleerd naar de *Paleocene-Eocene Thermal Maximum* en de *Middle Eocene Climatic Optimum*, en magnetostratigrafie.

In **Hoofdstuk 2** wordt een, in zijn continuïteit unieke en chronostratigrafisch goed gekalibreerde TEX$_{86}$ curve van zeewatertemperatuur gepresenteerd van ODP Leg 189 Hole 1172D op het East Tasman Plateau (~65° Zuid; Fig. 1 *in Introduction and Synopsis*). We laten zien de oppervlaktewatertemperatuur in de zuidwest Pacifische Oceaan steeg tot boven hedendaagse temperaturen in de tropen (tot 34°C) tijdens het vroeg Eoceen (53 miljoen jaar geleden) en dat het geleidelijk afkoelde tot ongeveer 21°C in het laat Eoceen (36 miljoen jaar geleden). Onze resultaten doen vermoeden dat er nagenoeg geen latitudinale temperatuursgradiënt bestond tussen subequatoriale en subpolaire breedtegraden ten tijde van het vroeg Eoceen (~55-50 miljoen jaar geleden). Deze latitudinale temperatuurgradiënt werd beduidend sterker in het midden Eoceen. Als de Eocene afkoeling het resultaat is van dalende atmosferische CO$_2$ concentraties (Zachos et al., 2008), dan zijn er in theorie additionele mechanismes nodig om de relatieve stabiliteit van tropische oppervlaktetemperaturen (Pearson et al., 2007) te verklaren bij evidente afkoeling op hogere breedtegraden.

Verschillende vroeg Paleogene plankton assemblages worden, in meer of mindere mate, gedomineerd door endemische soorten, ondanks warme condities op de polen en zwakke meridionale temperatuurgradiënten. Om de evolutie van het Paleogene plankton endemisme in zuidelijke polaire gebieden te documenteren, wordt in **Hoofdstuk 3** de verspreiding in ruimte en tijd van dinocysten onderzocht. Bovendien bepalen we hun respons op veranderingen in de regionale oppervlaktewatertemperatuur. Paleocene en vroeg Eocene (~65-50 miljoen jaar geleden) dinocystenassociaties waren voor het overgrote deel kosmopoliet van aard in de Zuidelijke Oceaan, maar een prominente shift van kosmopoliet-gedomineerde naar endemisch-gedomineerde assemblages (de zogeheten *transantarctic flora*) vond plaats rond de vroeg-midden Eoceen overgang (~50 miljoen jaar geleden). De geografische verspreiding en patronen in relatieve abundantie van de *transantarctic flora* komen sterk overeen met oppervlakteoceaanstromingen, zoals ze gereconstrueerd zijn met behulp van computermodelexperimenten, in het Eoceen. Dinocysten zijn kwantitatief vergeleken met de TEX$_{86}$-gebaseerde zeewater-

temperatuurreconstructies zoals gepubliceerd in Hoofdstuk 2 voor het vroeg en midden Eoceen van ODP Site 1172. Een conclusie is dat het begin van de proliferatie van de *transantarctic flora* aan het begin van het midden Eoceen niet lineair is toe te schrijven aan regionale temperatuurreconstructies. Alleen wanneer de *transantarctic flora* goed en wel domineren in het midden Eoceen, mogelijk aangestuurd door productiviteitsveranderingen op oceaanschaal, zien we een goede relatie tussen de relatieve abundantie van *transantarctic flora* en temperatuur.

In **Hoofdstuk 4** wordt een PETM sectie behandeld van ODP Hole 1172D. De organische paleothermometer TEX_{86} laat zien dat de oppervlaktewatertemperatuur steeg van ∼26°C naar ∼33°C tijdens de PETM. Dergelijke temperaturen voor, tijdens en na de PETM zijn meer dan 10°C warmer dan die gesimuleerd worden door klimaatmodellen voor deze breedtegraad. Dus zou het zo kunnen zijn dat niet alleen Noordpool-, maar ook Zuidpooltemperaturen lager zijn ingeschat zijn in simulaties van voorbije broeikaswerelden door de modernste, meest geavanceerde en volledig gekoppelde klimaatmodellen. Er is een vroege verschijning van abundante *Apectodinium* spp. in Hole 1172D, in overeenstemming met andere secties met hoge sedimentaccumulatie, zoals in New Jersey. Dit doet eens te meer vermoeden dat milieuveranderingen aan de koolstofisotopenexcursie voorafgingen, op wereldwijde schaal. Organische dinoflagellaten cysten assemblages suggereren een versterkte aanvoer van riviermateriaal naar de boorplek ten tijde van de PETM, tegelijk met eustatische zeespiegelstijging. Bovendien zijn veranderingen in seizoenaliteit van het regionale hydrologische systeem, en wellicht stormactiviteit uit de assemblages te herleiden. Ten slotte laten de sterke variatie in dinoflagellaten cyst assemblages tijdens de PETM zien dat het zuidwest Pacifische klimaat significant varieerde over tijdschalen van $10^3 – 10^4$ jaren, een ontdekking die in lijn is met vergelijkbare studies van de PETM op het continentale plat van New Jersey.

In **Hoofdstuk 5** wordt de rol onderzocht van het open gaan van een cruciale zeestraat in de Zuidelijk Oceaan, de *Tasmanian Gateway*, in Antarctische klimaatverandering. Proxy gegevens worden gepresenteerd van recentelijk opgeboorde sedimenten van het Wilkes Land continentaal plat, voor de kust van Antarctica (Escutia et al., 2011), welke afgezet zijn aan de westkant van de *Tasmanian Gateway* in de *Australo-Antarctic Gulf*. Geïntegreerde analyses van zowel mariene als terrestrische microfossielen en organisch geochemische componenten laten een sterke afkoeling zien van oost-Antarctische wateren (3-4°C) en aangrenzende landmassa's (6°C), rond 50 miljoen jaar geleden. Opvallend is dat deze afkoeling samenvalt met het begin van het overheersen van typisch zuidwest Pacifisch fytoplankton welke gemigreerd zijn via een blijkbaar ondiepe zuidelijke opening van de *Tasmanian Gateway*. Tezamen impliceren deze observaties dat het begin van de *Antarctic Counter Current* door de *Tasmanian Gateway* de invloed van de warmere *Proto-Leeuwin Current* heeft geminimaliseerd. Dit suggereert een sterke rol voor de ondiepe opening van de *Tasmanian Gateway* in Antarctische afkoeling, wat deze regio voorbereidt op grootschalige cryosfeerontwikkeling ten tijde van de Eoceen-Oligoceengrens rond 34 miljoen jaar geleden.

De lange-termijn warmte in het Eoceen (~56-34 miljoen jaar geleden) wordt alom in verband gebracht met verhoogde atmosferische druk van koolstofdioxide (pCO$_2$). Echter, dit verband is nog niet vastgesteld voor korte-termijn perturbaties. In **Hoofdstuk 6** worden zowel pCO$_2$- als zeewatertemperatuurreconstructies getoond voor een periode van tijdelijke wereldwijde opwarming: de *Middle Eocene Climatic Optimum* (MECO; ~39 miljoen jaar geleden). Organische moleculaire paleo-thermometrie laat een opwarming van zuidwest Pacifische oppervlaktewateren van 3 tot 6°C zien, terwijl pCO$_2$ gelijktijdig twee- tot driemaal zo hoog wordt. De evidente overeenstemming tussen zeewatertemperatuur- en pCO$_2$ trends tijdens de MECO doen vermoeden dat verhoogde pCO$_2$ concentraties een grote rol speelden in wereldwijde opwarming ten tijde van de MECO.

Samenvattend levert dit proefschrift een gedetailleerde weergave van de vroeg Paleogene transitie van een Broeikas- naar een IJskastwereld in de Zuidelijk Oceaan. De dinocystenzonatie die een groot deel van het vroeg Paleogeen bestrijkt, kan worden gebruikt als leidraad voor toekomstige biostratigrafie en ouderdomsbepalingen, wat cruciaal is gezien de afwezigheid van alternatief biostratigrafisch gereedschap in de Paleogene Zuidelijke Oceaan. De zeewater-temperatuurevolutie in de zuidwest Pacifische Oceaan volgt die in zuurstof isotopen in benthische foraminiferen, wat doet vermoeden dat de zuid Pacifische Oceaan een sector was waar diep water werd gevormd. Echter, absolute temperaturen blijken veel hoger dan die bepaald met benthische foraminiferen. Deels kan dit het gevolg zijn van seizoenale afwijkingen, bijvoorbeeld dat het diepwater met name gevormd wordt in winter en dat de gereconstrueerde oppervlaktewatertemperaturen in de Zuidelijke Oceaan vooral naar zomertemperaturen neigen. De biotische reactie op de oppervlakte-watertemperatuurevolutie in de Zuidelijke Oceaan is controversieel: het domineren van endemische dinocysten assemblages lijkt niet gerelateerd aan temperatuur. De werkhypothese was intuïtief dat de verspreiding van een endemische fytoplanktongemeenschap op hoge breedtegraden het resultaat was van een succesvolle concurrentiestrijd met warmwatersoorten ten tijde van afkoeling. Nu blijkt dat de zeewatertemperatuur niet verantwoordelijk kan worden gehouden voor het begin van de wijde verspreiding van de endemische planktonsoorten, is er een alternatieve hypothese opgesteld, die een rol toedicht aan nutriëntentoename in de oppervlaktewateren in het midden Eoceen. Inderdaad, verscheidene onafhankelijke proxies laten gezamenlijk een verschuiving naar meer nutriëntentoevoer zien, beginnende in het late vroeg Eoceen (~50 miljoen jaar geleden), van export productiviteitstoename in de gehele Pacifische Oceaan tot een versterking van de gradiënt van stabiele koolstofisotopen tussen oceanen. De PETM is ook een opvallend interval in de Zuidelijke Oceaan gekenmerkt door zeer snelle opwarming van de oppervlaktewater-temperatuur: 5-8°C. Ook hier is een superdominantie van tropische representanten van dinocystengenus *Apectodinium* vastgesteld. Net als bij andere PETM secties is ook in de zuid Pacifische Oceaan een nadrukkelijke milieuverandering zichtbaar voordat de negatieve koolstof-isotopenexcursie zich voordoet. Dit is in samenspraak met verdenkingen dat initiële klimaatveranderingen vooraf gingen aan de uitstoot van koolstof verantwoordelijk voor de koolstof-isotopenexcursie. Vervolgens zou dit kunnen betekenen dat de koolstofisotopen-excursie een respons is op klimaatverandering, in plaats van een forcering. Een ander

voortdurend debat dat bediscussieerd wordt in dit proefschrift is de rol van het open gaan van bepaalde zeestraten en hun rol in klimaatverandering. Recentelijk (2010) opgeboorde sedimenten van het Wilkes Land continentaal plat, Antarctica, laten zien de doorstroming van de *Antarctic Counter Current* rond de vroeg-midden Eoceen overgang (52-50 miljoen jaar geleden) begon, gelijktijdig met een versterkte afkoeling van Antarctische landmassa's en kustwateren. Deze opening van de *Tasmanian Gateway* vond plaats op het moment dat Aarde langzaam maar zeker begon af te koelen naar 'ijskast'-condities zoals de afgelopen 35 miljoen jaar bestaan. Daarom zou het opengaan van de *Tasmanian Gateway* wel eens het pad vrij kunnen hebben gemaakt voor klimaatafkoeling, waar atmosferische CO_2 druk waarschijnlijk nog steeds een grote forceringsfactor was in Paleogene klimaatsevolutie. Inderdaad, we zien desondanks dat de relatief kortdurende klimaatverandering van de MECO samenvalt met een sterke toename in de atmosferische CO_2 concentraties, wat een grote rol toedicht aan CO_2 als forcering van klimaatveranderingen.

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This is where it has to happen. It was an intense feeling of shame that I felt putting just my own name on the front cover of this thesis. Rather, I would have liked to put all the names listed below on the front cover and my name here. Oh well, I guess I have to obey the rules...

In springtime 2006, after the Master's course 'Extreme Climate Transitions' I thought: let's ask that Henk Brinkhuis guy whether he has a Master's internship subject for me. From that day onwards, Henk has supervised me in the typical Brinkhuis way: letting go where it is possible and helping out wherever he could. After my graduation in September 2007, I was really hoping to continue this work in a PhD, which worked out really well thanks to the Faculty of Science and the LPP Foundation. I applied (after encouragements by Henk) for the IODP expedition to Wilkes Land, Antarctica. Subsequently, Henk was asked to become co-chief of the Leg. Coincidence dictated that Dan, his brother, became videographer, and I was invited to sail as palynologist. This is just one example where shear coincidence and serendipity has brought the most brilliant things together. For me it suffices to say that Henk is the 'totaalsupervisor'. With 'totaalvoetbal', the Dutch soccer coach Rinus Michels led Dutch teams to superior levels in the 70's and 80's. Soccer players were encouraged to show up at different positions on the pitch all the time, and therefore the ball circulated very quickly. Henk gave me the opportunity to show up everywhere in the 'scientific pitch' all the time during my Master's and PhD. Besides supervising, he has always been and still is in for good fun outside the University, with the endless poker nights, and nice 'Christmas' (November through March) dinners at his home. Likewise, Dutch soccer teams of the 70's were quite notorious for their entertainment outside the soccer pitch as well...

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Stefan, I visited the NIOZ way too little, and in my memory it was always during the harsh part of the year. Even though, while stormy and cold outside, within your group there was always the feeling of a warm welcome. It was really good fun traveling all the way up to the island of Texel, because I knew who were waiting there for me. Thank you so much for the pleasant cooperation, and I hope that the finishing of this thesis is not the end of that.

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The Urbino Summer School of Paleoclimatology was like a professional playground for me: I got to meet the most interesting people, team up in discussions with beers and good food, and experience the birth of so many new projects, views, opinions. That worked really inspiring and set me off in science right from the start. Many thanks therefore go out to Simone Galeotti and his Italian co-workers who made the summer schools logistically possible, but also people like Matthew Huber, Paul Pearson, Robert DeConto, James Zachos, Rich Pancost, Jerry Dickens and loads of people I forget now for so many interesting discussions by the pool.

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'For whatever we lose (like a you or a me)
it is always ourselves we find in the sea'

- **E.E. Cummings (1894-1962)**

Curriculum Vitae

Peter Bijl werd op 21 oktober geboren, in Hoogeveen, maar groeide op in Papendrecht. Na zijn VWO-examen aan De Lage Waard in Papendrecht in 2002 begon hij zijn studie Geologie aan Universiteit Utrecht. In het tweede jaar volgde hij het *Bachelor track Biogeology*, die hij afsloot met een praktisch onderzoek bij de leerstoelgroep Stratigrafie en Paleontologie van de Universiteit Utrecht, naar benthische foraminiferen assemblageveranderingen in de Waddenzee over de afgelopen 150 jaar. In 2005 startte Peter de *Master Biogeology* aan de Universiteit Utrecht, waarin hij in april 2006 zijn afstudeerscriptie begon bij Henk Brinkhuis en Appy Sluijs van de leerstoelgroep Palaeoecologie van de Universiteit Utrecht. Hierbij hij keek naar paleomilieuveranderingen in de Zuid Pacifische Oceaan ten tijde van het Paleoceen-Eoceen. Aan de hand van studies naar dinoflagellaten cysten assemblages werden oppervlaktewater veranderingen gereconstrueerd. Daarnaast heeft hij voor zijn afstudeerscriptie veranderingen in elementenverhoudingen geanalyseerd in boorkernen met behulp van een *X-ray fractionation scanner* bij Ursula Röhl aan de Bremen Universiteit. Ook heeft hij op het NIOZ Koninklijk Nederlands Instituut voor Zeeonderzoek op Texel organisch geochemische analyses uitgevoerd. Tijdens zijn master stage was Peter betrokken bij het winnende project van de Academische Jaarprijs. Het project had als doel wetenschappelijke bevindingen van de studie naar de Paleogene broeikaswereld en de lessen voor toekomstige klimaatverandering naar het middelbaar onderwijs te over te brengen. Middels informatieve clips, een website en prijsvraag voor bovenbouw HAVO/VWO werden leerlingen bekend met het aardwetenschappelijk onderzoek. Na het behalen van zijn *MSc.* in 2007 begon hij in januari 2008 zijn promotieonderzoek bij de leerstoelgroep Palaeoecologie, onder begeleiding van Henk Brinkhuis, Stefan Schouten (NIOZ) en Appy Sluijs.

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