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Early Eocene to middle Miocene cooling and aridification of East Antarctica

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[1] Few high-latitude terrestrial records document the timing and nature of the Cenozoic “Greenhouse” to “Icehouse” transition. Here we exploit the bulk geochemistry of marine siliciclastic sediments from drill cores on Antarctica’s continental margin to extract a unique semiquantitative temperature and precipitation record for Eocene to mid-Miocene (~54–13 Ma). Alkaline elements are strongly enriched in the detrital mineral fraction in fine-grained siliciclastic marine sediments and only occur as trace metals in the biogenic fraction. Hence, terrestrial climofunctions similar to the chemical index of alteration (CIA) can be applied to the alkaline major element geochemistry of marine sediments on continental margins in order to reconstruct changes in precipitation and temperature. We validate this approach by comparison with published paleotemperature and precipitation records derived from fossil wood, leaves, and pollen and find remarkable agreement, despite uncertainties in the calibrations of the different proxies. A long-term cooling on the order of ≥8°C is observed between the Early Eocene Climatic Optimum (~54–52 Ma) and the middle Miocene (~15–13 Ma) with the onset of
transient cooling episodes in the middle Eocene at ~46–45 Ma. High-latitude stratigraphic records currently exhibit insufficient temporal resolution to reconstruct continental aridity and inferred ice-sheet development during the middle to late Eocene (~45–37 Ma). However, we find an abrupt aridification of East Antarctica near the Eocene-Oligocene transition (~34 Ma), which suggests that ice coverage influenced high-latitude atmospheric circulation patterns through albedo effects from the earliest Oligocene onward.

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1. Introduction

[2] Reconstruction of the Cenozoic continental environmental history of Antarctica is important to understand the role of high-latitude physical and biogeochemical processes in the global ocean and climate system. Due to extensive ice cover and erosion of sedimentary archives, obtaining terrestrial paleoenvironmental records across major climate transitions has been challenging. We present new bulk geochemical data from Integrated Ocean Drilling Program (IODP) Site U1356 [Escutia et al., 2011], Ocean Drilling Program (ODP) Site 1166 [O’Brien et al., 2001], and McMurdo Erratic D1 [Harwood and Levy, 2000], along with existing data from the Cape Roberts Project (CRP) drill cores [Barrett, 2007; Pascshier and Krissek, 2008] and the CIROS-1 drill core (Figure 1) [Rosser and Pyne, 1989]. We apply terrestrial climofunctions derived from chemical weathering indices for semiquantitative paleoclimatological interpretations, an approach pioneered by Nesbitt and Young [1982].

[3] The average exposed continental crust contains more than 60% feldspars by volume, with subordinate contributions from volcanic glass and other labile minerals [Nesbitt and Young, 1982]. Chemical weathering by meteoric and soil waters containing dissolved carbonic and organic acids transforms alkaline aluminosilicates to clay minerals, thereby depleting minerals in cations, such as Na⁺, K⁺, and Ca²⁺, while conserving Al³⁺. Numerous recent studies of efflux from watersheds suggest a strong imprint of climate on the chemical weathering in soils [e.g., White et al., 1999; Turner et al., 2010; Rasmussen et al., 2011; Eiriksdottir et al., 2013]. In addition, studies of geochemical mass balance and cosmogenic nuclides show that the degree to which bulk soil profiles become depleted in labile elements is strongly dependent on the prevailing climate [Sheldon et al., 2002; Riebe et al., 2004; Nordt and Driese, 2010]. Sheldon et al. [2002] demonstrated by analysis of a database of 126 North American soils [Marbut, 1935] that different molar oxide ratios of K, Na, and Ca versus Al in soil weathering horizons anticorrelate with mean annual temperature (MAT) and mean annual precipitation (MAP). Sheldon et al. [2002] proposed two climofunctions, which provided paleotemperature and precipitation estimates for paleosols comparable to those acquired via other independent proxies.

[4] We assume that the detrital geochemistry of the mudrocks in the continental margin successions semiquantitatively preserves the average paleoenvironmental signature of the soil mantle in the source drainage basins (Figure 1). Physical erosion in continental uplands provides a fine-grained sediment source from a broad spectrum of source rock lithologies. In glacially fed sediment transport systems, specifically, silt-sized glacial rock flour is transported to the coastal zone as suspended load via wind and proglacial meltwater streams. The detrital geochemistry of continental margin sediments adjacent to both nonglaciated and glaciated catchments primarily records the climatic conditions in the vegetated coastal zone, where silicate weathering rates are high. Here the sediment resides...
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weathering where changes in the
temporally, by temperature, precipitation (runoff),
ically limited weathering is regulated, spatially and
complicate the climate-weathering relationship, kinet-
soils database
terrestrial climofunctions based on the North Ameri-
can soils database. Published vegetation data (Table S1 in the supporting


stabilized in floodplains, immediately prior to the final phase of erosion, transport, and offshore de-

[5] Silicate chemical weathering in uplands and periglacial environments is kinetically limited due to the high erosion rates and abundant supply of silt-sized feldspars with a large specific surface area [Anderson, 2005]. The rate of leaching of cations is initially rapid but diminishes exponentially within a few thousand years as mineral surfaces age [Egli et al., 2001; Anderson, 2005]. In contrast to transport-limited weathering where changes in the fluid saturation state complicate the climate-weathering relationship, kinetically limited weathering is regulated, spatially and temporally, by temperature, precipitation (runoff), and vegetation [West et al., 2005; Anderson, 2005, 2007; Eiriksdottir et al., 2013].

[6] Building upon the above considerations, we apply terrestrial climofunctions based on the North American soils database [Marbut, 1935] to siliciclastic sediments on the Antarctic continental margin. We acknowledge that uncertainties may arise from regional differences between North America and Antarctica but argue that the application of terrestrial climofunctions is justified by the fact that both East Antarctica and North America represent petrologically diverse high-latitude cratonic regions periodically covered by ice sheets. Moreover, in combination with published vegetation data (Table S1 in the supporting information) from several Antarctic sites [Pross et al., 2012; Francis et al., 2007, 2009 and references therein], this approach provides a unique opportunity to reconstruct Antarctic continental paleotemperature and precipitation dynamics from the peak of the early Eocene “Greenhouse” to the middle Miocene “Icehouse” (~54–13 Ma).

2. Methods and Materials

[7] The CRP [Barrett, 2007] and ODP Site 1166 [O’Brien et al., 2001] cores were recovered from the Antarctic continental shelf in the Ross Sea and Prydz Bay, respectively, whereas IODP Site U1356 was drilled on the Wilkes Land continental rise (Figure 1) [Escutia et al., 2011]. As documenting Antarctic continental conditions was a primary objective for these expeditions, the drill sites were positioned into seaward prograding sediment wedges that received material eroded from East Antarctic sources [Cooper et al., 1991]. The sample of McMurdo Erratic D1 is part of a collection of erratics documenting the otherwise poorly sampled middle to late Eocene Antarctic glacial history [Harwood and Levy, 2000].

[8] The major element chemistry of bulk mudrock samples from the Antarctic drill cores was determined using XRF and inductively coupled plasma atomic emission spectrometry (ICP-AES) instrumentation. We also incorporated data from strata recovered in the CIROS-1 drill hole in McMurdo Sound [Roser and Pyne, 1989] into our compilation, but a lack of carbonate data precluded application of the precipitation climofunction. Age models for the drill cores were based on Tauxe et al. [2012] for IODP Hole U1356A, Lavelle [1998], McIntosh [1998], and Florindo et al. [2005] for the CRP drill cores, Florindo et al. [2003] for ODP Hole 1166A, and on Harwood et al. [1989], Rieck [1989], and Wilson et al. [1998] for the CIROS-1 drill core. The middle Eocene age assessment for McMurdo Erratic D1 is derived from Levy and Harwood [2000]. All ages were converted to Gradstein et al. [2004] (Tables S2 and S4, supporting information). It should be noted however that the nanofossil datums used to calibrate the magnetostratigraphy for CIROS-1 [Wilson et al., 1998] are controversial [Watkins, 2007]. Published bulk chemical data are also available for the middle to upper Eocene of the La Meseta Formation on the northern Antarctic Peninsula [Dingle et al., 1998]. These sediments,

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1 All supporting information may be found in the online version of this article.
however, contain more than 40% sand [Dingle et al., 1998]; hence, interpretations of CIA-based weathering ratios derived from whole-rock geochemistry are not possible.

[9] Most of the new ICP-AES data for IODP Site U1356 were collected on board the JOIDES Resolution. Additional Miocene samples from IODP Site U1356 and all samples from ODP Site 1166 were analyzed in the ICP laboratory at Montclair State University. The sample preparation on board the JOIDES Resolution followed identical protocols to those applied in the ICP laboratory at Montclair State University [Murray et al., 2000]. In brief, the procedures involve fusion of 0.1 g of sample mixed with 0.4 g LiBO4 flux in a furnace and dissolution of the pellets in 7% HNO3, followed by filtration and a dilution prior to analysis. Elemental concentrations were calculated using 10 or more U.S. Geological Survey standards. The analytical uncertainties for oxide weight percentages in repeat runs and monitor standards are typically 1–5%.

[10] The bulk major element geochemistry of marine mudrocks reflects provenance, weathering history, and post-depositional alteration or diagenesis and deposition of salt [McLennan et al., 1993]. In sediments with limited diagenetic alteration and salt deposition, the relative abundances of labile mineral-bound alkali and alkaline Earth elements are controlled by paleoclimatic conditions [Nesbitt et al., 1996] which affect the intensity of chemical weathering in the source areas [Riebe et al., 2004]. The margin architecture [Cooper et al., 1991] and high sedimentation rates on the order of 10–100 m/Myr [Florindo et al., 2003, 2005; Tauxe et al., 2012] indicate that the dominant sediment flux is derived from nearby Antarctic continental sources. In addition to directly routed terrigenous sediments, the marine sediments investigated here contain subordinate proportions of wind-blown dust; marine biogenic components, such as calcareous foraminifera and nannofossils; siliceous microfossils; and salt. In sediments with a large terrigenous fraction, corrections can be made for the presence of carbonate [Nesbitt and Young, 1982]. Furthermore, Al, K, and Na occur only in trace metal concentrations in the biogenic fraction [Brewer et al., 2008]. Hence, in our analysis and interpretation of the bulk chemical data, we considered the following:

1. The elements of interest to the climofunctions are major elements in most rock-forming minerals and occur in trace metal concentration in other (biogenic) components of marine sediments [Brewer et al., 2008]. As a result, the weight percentages of these elements in average upper crust [Wedepohl, 1995] are between 11 and 34 times higher than the weight percentages in a purified diatomite. For the samples in this study, we estimated from smear slides that the terrigenous fraction contributes between 30% and 100% to the overall sediment composition. Following the above cited proportions of alkaline elements between biogenic and detrital phases, this means that even in the most diatom-rich samples, 79–94% of the alkaline earth elements in the bulk sediment will be contributed from the detrital fraction.

2. Kryc et al. [2003] demonstrated that scavenging of Al is negligible for sediment with >3–5 wt % terrigenous matter, which applies to the cores studied here, where the terrigenous fraction is estimated to be between 30% and 100% and the remaining fraction primarily consists of biogenic silica [Escutia et al., 2011]. Given the low density of biogenic silica, the lower value of 30% terrigenous fraction is a conservative estimate and it likely corresponds to >30% weight percent terrigenous fraction.

3. Carbonate content was routinely measured for CRP, ODP, and IODP samples [Dietrich and Klosa, 1998; Dietrich et al., 2000, 2001; Shipboard Scientific Party, 2001; Escutia et al., 2011]. Most samples yielded <2 wt % carbonate, and corrections for CaO present in calcium carbonate were applied [Nesbitt and Young, 1982]. A few samples, including McMurdo Erratic sample D1, contained >10 wt % carbonate and were eliminated from the calculations of CIA-K to avoid the introduction of errors related to extensive carbonate corrections. Samples with abundant carbonate cement in the CIROS-1 drill hole [Bridle and Robinson, 1989] were also eliminated. No carbonate correction was applied to the samples from Site 1166, because the CaO concentrations were <1 wt % and carbonate corrections yielded negative values. Because no carbonate wt % data are available for the CIROS-1 drill hole, no CIA-K and MAP calculations are presented.

4. The contribution of Na from salts is negligible in sediments with a large siliciclastic fraction. Krissek and Kyle [2000] showed that Na values and major element ratios for washed and unwashed samples taken less than a meter apart in the CRP cores were indistinguishable.

5. Dissolution of feldspar and authigenic clay formation, due to the geothermal heat flux affecting porewater temperatures, is generally negligible for burial depths shallower than approximately 1000–1500 m. However, in the case of cores
in the Ross Sea, active magmatism during the past 25 Myr [Martin et al., 2010] may have increased porewater temperature conditions at shallower depth. Therefore, core sections were carefully chosen and sections with documented evidence for aluminosilicate diagenesis, such as the base of CRP-3 below 340 m below seafloor [Ehrmann et al., 2005], were eliminated from the analysis [cf. Passchier and Krissek, 2008].

6. Samples from the CIROS-1 drill hole were selected based on particle size (more than 60% mud), and samples with Al2O3/TiO2 ratios <18 were eliminated to avoid a volcanic overprint [cf. Passchier and Krissek, 2008].

[11] Following the approach of Sheldon et al. [2002], applying their climate data set and the soil chemistry database of Marbut [1935], we constructed slightly modified climofunctions. The soils of the Marbut [1935] database encompass a wide range of topographic settings, relief, and source rock lithologies and are covered by a variety of vegetation types [Sheldon et al., 2002]. Therefore, climofunctions developed from this database potentially find widespread application on continents with similar physiographic attributes as North America. Using the climate data from Sheldon et al. [2002] and the major element geochemistry data for soil B horizons in the Marbut [1935] database, we created linear least squares regression equations for MAP and MAT (Figure 2). The equation for MAT is identical to the climofunction used in Sheldon et al. [2002]. For MAP, we eliminated data points for pedons with calcrite horizons (with carbonate nodules) as our goal was to address the depletion of CaO in the silicate fraction.

[12] The climofunctions we applied to the bulk major element geochemistry are

\[
\text{MAT} = 18.5S + 17.3 \tag{1}
\]

in which the S index is represented by the molar ratio of Na2O and K2O to Al2O3 [Sheldon et al., 2002]. The applied climofunction for MAP was slightly modified using only the data points for pedons without calcrite horizons, to emphasize the silicate mineral-bound components:

\[
\text{MAP} = 143.75e^{0.0232(CIA-K)} \tag{2}
\]

in which CIA (chemical index of alteration) [Nesbitt and Young, 1982], represents 100 times the molar ratio of Al2O3 to Na2O, CaO*, K2O and Al2O3, and CIA-K, the same equation without K2O [Maynard, 1992]. CaO* designates CaO in detrital silicate minerals only. Although the regression coefficients are small for the MAT data, the correlations for both sets of variables are significant with \( p < 0.01 \). Using the regression equations to estimate MAT and MAP from S and CIA-K, respectively, we calculate the standard error of the prediction on the y axis (MAT and MAP) using the STEYX function in Excel [2007]. The standard errors were ±3.6°C and ±182 mm for MAT and MAP, respectively.

[13] The relatively large standard errors for MAT and MAP originate from physiographic effects other than temperature and precipitation, such as topography, variable mineral-specific weathering rates due to source rock variability, vegetation, and soil pH. Recent field studies, however, confirm the strong climatic imprint on the chemical depletion of soils, despite uncertainties in the effects of these other variables [e.g., Turner et al., 2010; Ferrier et al., 2010; Nordt and Driese, 2010, Rasmussen et al., 2011]. We assume that there is negligible chemical alteration of detrital sediment components during suspended load transport of mud from coastal source to continental margin sink [e.g., Nesbitt et al., 1996].

3. Results

[14] On a CaO* + Na2O – Al2O3 – K2O (CN-A-K) ternary diagram (Figure 3), the samples from Sites

Figure 2. Least squares regression plots for (left) MAT and (right) MAP using data from soil B horizons [Sheldon et al., 2002].
1166 and U1356 and the CRP drill cores plot in approximate alignment with the CN-A axis, which is consistent with differentiation of detrital materials via chemical weathering from an intermediate crystalline source rock or a mixture of source rocks of average upper crustal composition [Nesbitt and Young, 1984, 1989]. The alignment of plotted data points with the CN-A axis (Figure 3) also illustrates that differences in source rock composition did not significantly impact the alkaline element geochemistry. If variability in the abundance of Ca- and Na-bearing plagioclase and K-feldspar was governed by source rock changes, the sample compositions would align parallel to the CN-K axis.

[15] Pronounced changes in alkaline element geochemistry are noted between the Eocene “Greenhouse” and Miocene “Icehouse” sediments (Figure 4). Values of CIA-K for the Antarctic drill holes decrease from an average of ~86 during the 

Figure 3. Ternary plots of CaO* + Na2O – Al2O3 – K2O (CN-A-K) system [Nesbitt and Young, 1984, 1989]. The distribution of the data points is consistent with loss of alkalinity through chemical weathering.

Figure 4. (a) Mean annual temperature (MAT) and (b) mean annual precipitation (MAP) calculated from the detrital geochemistry in Antarctic continental margin drillcores using climofunctions based on the Marbut [1935] soil chemistry database. Values for average (AVG) upper crust were calculated from Wedepohl [1995]. Gray shadings designate confidence intervals calculated from the standard errors. The smoothing trendline is derived using the LOWESS (Locally Weighted Scatterplot Smoothing) algorithm [Cleveland, 1979].
Early Eocene Climatic Optimum to ~63 during the middle Miocene “Icehouse,” whereas values for the S index increase from ~0.24 to ~0.60. The calculated terrestrial MAT ranges from ≤4 to 14°C and MAPs range from ~400 to 1200 mm (Figure 4). The maximum values are below the upper limits of MAT > 18°C or MAP > 1600 mm that have been suggested for complete leaching of labile minerals from soils [Sheldon et al., 2002; Rasmussen et al., 2011]. Corresponding to an absence of significant feldspar weathering, lower limits can be established from values for average upper crust (Figure 4) [Wedepohl, 1995], which yield MAT < 7.7°C and MAP < 460 mm in our calibrations. In portions of the Oligocene and Miocene record, CIA-K and S index values plot below average upper crustal compositions (Figure 4), and therefore, the calculated MAPs and MATs should be regarded as maximum values. Because of a lack of carbonate data for the CIROS-1 drill hole and very high carbonate values for D1 [Passchier, 2000], we were unable to calculate CaO*, and hence CIA-K, for these records.

An abrupt decrease in humidity from 1000–1250 mm to 500–800 mm is apparent at the Eocene-Oligocene transition (Figure 4b). In the Oligocene, regional differences emerge between sites with higher MAP values at Site U1356 off Wilkes Land (~750 mm) than at the CRP sites (~600 mm) in the Ross Sea (Figure 4b). The MAT average 14°C for the early and middle Eocene with some variability in the late Eocene and steadily decline through the Oligocene and Miocene (Figure 4a). The data density in the mid to late Eocene portion of the record is low due to poor representation of this stratigraphic interval. The records from the CRP sites, CIROS-1, and IODP Site U1356 partially overlap in time and also in paleotemperature trends (Figure 4a). Overall, a long-term cooling on the order of ≥8°C is inferred over a ~41 Myr time interval between the Early Eocene Climatic Optimum and the middle Miocene.

4. Discussion

4.1. Sediment Recycling

Thiry [2000] points out that inheritance of ancient kaolinite complicates the paleoclimatological interpretation of clay minerals in marine sediments. However, the methodology we use here is considerably different as we address the bulk geochemistry of mudrocks, which include a significant fine to coarse silt fraction. The methodology based on bulk sediment weathering ratios, pioneered by Nesbitt and Young [1982], considers the degree to which the feldspars in the mudrocks are altered. The central idea is that survival of fresh feldspar is related to the prevailing climate, with more silt-sized feldspar converted to clay minerals under warm and humid conditions, and survival of abundant feldspar during cold and dry conditions. As glaciers produce glacial rock flour through erosion of crystalline cratonic rocks, a steady supply of silt is maintained, which depending on the paleoclimatic conditions, will be affected by different degrees of chemical weathering in the coastal zone immediately before transport and deposition in the marine environment.

Although erosion of weathered sedimentary platform rocks can affect the composition of glacial sediments directly overlying these platform rocks, the effects are small in offshore sediment cores, as demonstrated by the CIA values of Quaternary diamicts from Antarctic drill cores, which are 47–56 [Passchier and Krisek, 2008; this study, and unpublished data], compared to 46 for average upper crust [Wedepohl, 1995]. Kaolinite reworking hence may result in a maximum of approximately 20% decrease in alkalinity of the bulk sediment, translating into a less than 20% increase in temperature and humidity.

4.2. Comparison With Vegetation-Based Climate Reconstructions

The high MAP of >1000 mm calculated for the Eocene at Sites U1356 and 1166 is in close agreement with the MAP for subtropical and temperate rain forest biomes on East Antarctica as reconstructed from pollen records in the same cores [Pross et al., 2012], and with plant macrofossil assemblages in Eocene strata near the Antarctic Peninsula and on subantarctic islands [Francis et al., 2007, 2009; and references therein]. A decrease in humidity is apparent from the early Oligocene onward. An estimated MAP of 500–800 mm for the late Oligocene and early Miocene is higher than the modern MAP of 150–400 mm for an Antarctic continent without vegetation. MAP levels <460 mm, however, are below the applicable limit of the climofunction (Figure 4b).

In agreement with our results, temperatures derived from plant macrofossils on the Antarctic Peninsula, on subantarctic islands, and in the Ross Sea are relatively warm in the early and middle Eocene, with MAT of 7–15°C (Table S1 and Figure 5).
Leaf fossils and pollen from Oligocene-Miocene drill cores in the Ross Sea suggest the presence of a low-diversity Nothofagus-dominated forest and shrub-land [Askin and Raine, 2000], typical of modern conditions along the southern tip of South America with MAT of 5–8°C [Prebble et al., 2006; Francis et al., 2007, 2009; Lewis et al., 2008; Macphail and Truswell, 2004; Poole et al., 2005; Prebble et al., 2006; Pross et al., 2012; Raine, 1998; Truswell and Macphail, 2009; Warny et al., 2009]. Error bars, where reported, indicate range based on standard errors for T (vegetation). The gray lines indicate the confidence intervals for the MAT based on the soil climofunction (see Figure 4a).

A discrepancy between vegetation-derived temperatures and the MAT derived from the soil climofunction, however, is apparent in the late Eocene at 37–35 Ma (Figure 5). The major element data for the upper Eocene are from the CIROS-1 drill hole (Figure 4a) for which the age model is controversial, whereas the vegetation data are from subantarctic islands [Francis et al., 2009], giving way to an impoverished shrub tundra and mossy tundra vegetation in the mid-Miocene [Lewis et al., 2008].

[23] A discrepancy between vegetation-derived temperatures and the MAT derived from the soil climofunction, however, is apparent in the late Eocene at 37–35 Ma (Figure 5). The major element data for the upper Eocene are from the CIROS-1 drill hole (Figure 4a) for which the age model is controversial, whereas the vegetation data are from subantarctic islands [Francis et al., 2009], giving way to an impoverished shrub tundra and mossy tundra vegetation in the mid-Miocene [Lewis et al., 2008].

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4.3. Climatic Implications

[23] The onset of continental-scale ice growth on Antarctica at ~34 Ma is relatively well established from deep-sea proxy records, glacigenic sediments, and ice sheet modeling [Hambrey et al., 1991; Coxall et al., 2005; DeConto et al., 2008]. The pronounced aridification we observe near the Eocene-Oligocene boundary (Figure 4) is in agreement with the growth of an ice sheet and its influence on atmospheric circulation patterns: The size of the ice sheet determines the strength of the polar high-pressure cell through an albedo feedback and promotes the onset of strong northward directed cold and dry katabatic winds. Following this reasoning, our record supports an absence of large continental ice sheets in Antarctica in early to mid-Eocene time [DeConto et al., 2008], as conditions remained relatively humid. Although we do not have
a continuous record across the Eocene-Oligocene boundary, the ~3°C drop in temperature observed in the running average of our data set (Figure 4b) is comparable to a 3–4°C drop in sea surface temperatures of the nearest high latitude ODP Site 277 [Liu et al., 2009].

[24] Midlatitude terrestrial paleoclimate records show similar aridification and cooling trends across the Eocene-Oligocene transition and of the same magnitude although there is considerable geographic variability. Paleoaccumulations from the west-central United States are geographically diverse and records from central Oregon show significant aridification, whereas the effects are smaller in the dryer areas of the continental interior of the United States due to dryer conditions in the Eocene [Sheldon and Retallack, 2004]. In contrast, terrestrial paleotemperature records show a significant drop in the midcontinent area of the United States [Zanazzi et al., 2007], whereas a gradual temperature change is observed in the more maritime regions of central Oregon [Sheldon and Retallack, 2004], Argentina [Kohn et al., 2004], and the Isle of Wight [Sheldon et al., 2009].

[25] In Antarctica, regional differences in humidity observed between the Oligocene records of the Wilkes Land (U1356) and the Ross Sea (CRP) coasts (Figure 4) are in agreement with the results from a regional climate model for a 2 × CO₂ scenario [DeConto et al., 2007]. The model, however, predicts year-round subfreezing coastal temperatures for a fully glaciated continent, in contrast to temperatures reconstructed from vegetation [Thorn and DeConto, 2006] and our geochemical records. The results of the proxies may be biased toward interglacial conditions when plant biomass production reached a maximum, and because the soil geochemical climofunctions are calibrated to peak interglacial MATs. Due to Antarctica’s strong seasonality, there is possibly also a bias toward the summer season. During portions of the Oligocene and Miocene, however, S index values are below average upper crustal compositions and the calculated MATs should be regarded as maximum temperatures (MAT < 7.7°C). Hence, model results for a fully glaciated continent are consistent with our proxy evidence for the earliest Oligocene and portions of the early to mid-Miocene.

5. Conclusions

[26] Geochemical signatures of marine sediments on the Antarctic continental margin together with vegetation reconstructions carry unique semiquantitative evidence of continental paleoclimatic conditions. The strong partitioning of alkaline elements in the detrital fraction of marine sediments allows application of soil climofunctions, which gauge alkaline element depletion in relation to mean annual temperature and precipitation in the source area. There is a strong agreement between chemical and vegetation proxies for Eocene-Miocene times in this study, indicating ≥8°C cooling from the early Eocene to middle Miocene. Although our data compare well with results from other proxy records and modeling, indicating the onset of continental-scale ice growth at the Eocene-Oligocene transition, further drilling on the Antarctic continental margin will be necessary to reconstruct continental aridity and inferred ice development during the mid to late Eocene (~45–37 Ma).

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