JOURNAL OF GEOPHYSICAL RESEARCH: SOLID EARTH, VOL. 118, 1845–1861, doi:10.1002/jgrb.50151, 2013

Environmental magnetic record of paleoclimate, unroofing of the Transantarctic Mountains, and volcanism in late Eocene to early Miocene glaci-marine sediments from the Victoria Land Basin, Ross Sea, Antarctica

Andrew P. Roberts,[1] Leonardo Sagnotti,[2] Fabio Florindo,[2] Steven M. Bohaty,[3] Kenneth L. Verosub,[4] Gary S. Wilson,[5] and James C. Zachos[6]

Received 17 December 2012; revised 8 March 2013; accepted 10 March 2013; published 6 May 2013.

[1] We synthesize environmental magnetic results for sediments from the Victoria Land Basin (VLB), which span a total stratigraphic thickness of 2.6 km and a ~17 Myr age range. We assess how magnetic properties record paleoclimatic, tectonic, and provenance variations or mixtures of signals resulting from these processes. The magnetic properties are dominated by large-scale magnetite concentration variations. In the late Eocene and early Oligocene, magnetite concentration variations coincide with detrital smectite concentration and crystallinity variations, which reflect paleoclimatic control on magnetic properties through influence on weathering regime; high magnetite and smectite concentrations indicate warmer and wetter climates and vice versa. During the early Oligocene, accelerated uplift of the Transantarctic Mountains gave rise to magnetic signatures that reflect progressive erosion of the Precambrian-Mesozoic metamorphic, intrusive, and sedimentary stratigraphic cover succession associated with unroofing of the adjacent Transantarctic Mountains. From the early Oligocene to the early Miocene, a consistent fining upward of magnetite particles through the recovered composite record likely reflects increased physical weathering with glacial grinding contributing to progressively finer grained Ferrar Dolerite-sourced magnetite. After 24 Ma, the magnetic properties of VLB sediments are primarily controlled by the weathering and erosion of McMurdo Volcanic Group rocks; increased volcanic glass contents contribute to the fining upward of magnetite grain size. Overall, long-term magnetic property variations record the first-order geological processes that controlled sedimentation in the VLB, including paleoclimatic, tectonic, provenance, and volcanic influences.

Citation: Roberts, A. P., L. Sagnotti, F. Florindo, S. M. Bohaty, K. L. Verosub, G. S. Wilson, and J. C. Zachos (2013), Environmental magnetic record of paleoclimate, unroofing of the Transantarctic Mountains, and volcanism in late Eocene to early Miocene glaci-marine sediments from the Victoria Land Basin, Ross Sea, Antarctica, *J. Geophys. Res. Solid Earth*, *118*, 1845–1861, doi:10.1002/jgrb.50151.

[1]Research School of Earth Sciences, The Australian National University, Canberra, ACT, Australia.
[2]Istituto Nazionale di Geofisica e Vulcanologia, Rome, Italy.
[3]National Oceanography Centre, University of Southampton, Southampton, UK.
[4]Department of Geology, University of California, Davis, California, USA.
[5]Department of Marine Science, University of Otago, Dunedin, New Zealand.
[6]Department of Earth and Planetary Sciences, University of California, Santa Cruz, California, USA.

Corresponding author: A. P. Roberts, Research School of Earth Sciences, The Australian National University, Canberra, ACT 0200, Australia. (andrew.roberts@anu.edu.au)

2169-9313/13/10.1002/jgrb.50151

1. Introduction

[2] Environmental magnetism involves measurement of the magnetic properties of natural materials, such as sediments and soils, to investigate environmental processes. Environmental magnetism has been successfully used to analyze temporal trends in paleoclimatic processes, sediment provenance, weathering, transport pathways, and post-depositional diagenetic alteration of magnetic minerals in a wide range of environments [e.g., *Thompson and Oldfield*, 1986; *Verosub and Roberts*, 1995; *Maher and Thompson*, 1999; *Evans and Heller*, 2003; *Liu et al.*, 2012].

[3] Magnetic measurements of sediments spanning the Eocene-Oligocene (E-O) transition in the CIROS-1 core [*Barrett*, 1989] from the Victoria Land Basin (VLB), Ross

Sea, Antarctica, indicated that variations in magnetite concentration were synchronous with variations in the concentration and crystallinity of detrital smectite particles [*Sagnotti et al.*, 1998a]. Variations in smectite concentration and crystallinity have been widely observed across the E-O transition at sites around Antarctica [e.g., *Ehrmann and Mackensen*, 1992; *Diester-Haass et al.*, 1996; *Robert and Kennett*, 1997; *Ehrmann*, 1998; *Robert et al.*, 2002; *Ehrmann et al.*, 2005]. These variations are interpreted to represent paleoclimatic alternations between warm and humid periods, where chemical weathering of basic igneous rocks allowed smectite formation, and cool, dry periods, when chemical weathering was suppressed and physical (mechanical) weathering was dominant, largely through glacial processes. The shift from warmer to cooler conditions across the E-O transition is the most marked Cenozoic climatic transition, with onset of Antarctic glaciation and buildup of ice sheets driving progressive cooling from the early Paleogene greenhouse world to the late Paleogene-Quaternary icehouse world [e.g., *Miller et al.*, 1987; *Prothero*, 1994; *Zachos et al.*, 2001; *Coxall et al.*, 2005]. The potential sensitivity of magnetic parameters to weathering regime variations on the Antarctic continent and, therefore, to paleoclimate and ice-sheet history has motivated environmental magnetic studies of sediment cores from the VLB [*Sagnotti et al.*, 1998b, 2001a; *Verosub et al.*, 2000; *Roberts et al.*, 2003a]. There have been relatively few other environmental magnetic studies of paleoclimatic processes or of sediment provenance from Antarctic margin sediments [e.g., *Sagnotti et al.*, 2001b; *Brachfeld et al.*, 2002; *Kanfoush et al.*, 2002; *Pirrung et al.*, 2002; *Venuti et al.*, 2011].

[4] In this paper, we synthesize environmental magnetic results from a suite of sediment cores from the VLB (Figures 1 and 2 and Table 1). The studied cores represent a total stratigraphic thickness of about 2.6 km and span an age range of approximately 17 million years (across the E-O transition; Figure 3) [*Florindo et al.*, 2005]. They are unique because they represent the most comprehensive suite of continental shelf drillcores spanning this age range from the circum-Antarctic region. We examine critically whether the magnetic properties of these sediments record paleoclimatic, tectonic, diagenetic, or provenance variations or a mixture of signals resulting from these processes. Detailed rock magnetic data are presented for each core in our original studies and are not reproduced here [*Sagnotti et al.*, 1998a, 1998b, 2001a; *Verosub et al.*, 2000; *Roberts et al.*, 2003a]. We focus here on synthesizing temporal trends in environmental magnetic properties of VLB sediments. We only reproduce representative results where they contribute to a broader view of temporal trends, which has not been presented before.

2. Background

2.1. Geological Setting

[5] The VLB is the westernmost marine basin of the West Antarctic rift system in the Ross Sea (Figures 1a and 1c) [*Davey et al.*, 1982, 1983; *Cooper et al.*, 1987]. It consists of extended continental crust and is one of the world's largest active continental rift systems [*Cande et al.*, 2000]. Although it is largely aseismic, it is considered active because basic volcanism and extensional faulting have continued into the late Cenozoic [*Kyle*, 1990; *Behrendt et al.*, 1991]. The Transantarctic Mountains represent the shoulder of the West Antarctic rift system and rise to elevations of >4000 m at distances ~35 km from the coast [*Fitzgerald et al.*, 1986; *Stern and ten Brink*, 1989]. The East Antarctic Ice Sheet (EAIS) is grounded behind the Transantarctic Mountains, and mountain glaciers drain from the EAIS into McMurdo Sound (Figures 1b and 1c). VLB sediments therefore record both the uplift history of the Transantarctic Mountains and the waxing and waning of the EAIS [*Fielding et al.*, 2008]. Several scientific drilling projects have targeted VLB strata in McMurdo Sound since 1979 (Figure 1b and Table 1) to study sedimentation associated with Transantarctic Mountain uplift and EAIS fluctuations. Holes in a three-hole transect drilled in association with the Cape Roberts Project (CRP; Figure 1b) were spaced to recover progressively older strata by penetrating seaward-dipping seismic reflectors (Figures 2 and 3) [*Hamilton et al.*, 2001; *Henrys et al.*, 2001]. The studied cores were obtained from a drill-rig setup on a seasonal sea-ice platform. Use of high-speed diamond coring, with a mud circulation and riser system, generally enabled high core recovery (Table 1). On average, we sampled the studied 2.6 km of core at about 0.5 m intervals.

2.2. Chronology of the Studied Cores

[6] Large portions of the studied cores are dominated by glacial lithofacies (Figure 3), and construction of robust chronostratigraphies has been a major challenge. The glacigenic sedimentary sequences are truncated by numerous disconformities associated with ice-advance events. The discontinuous nature of the sedimentation makes it difficult to use magnetic polarity stratigraphy for dating without precise independent chronostratigraphic constraints. Rift-related volcanism was sporadic, and the oldest known contemporary volcanic products that are clearly not reworked are dated at 24 Ma [*McIntosh*, 2000]. Thus, radiometric dating of volcanic products is only partially useful for constraining the depositional age of VLB sediments. Matching of $^{87}Sr/^{86}Sr$ ratios from mollusk shells with the global curve for seawater strontium composition through geological time [*McArthur et al.*, 2001] provides numerical ages that are useful for constraining the age of these sediments [e.g., *Lavelle*, 1998, 2000, 2001]. However, $^{87}Sr/^{86}Sr$ dating is limited by the sporadic occurrence of mollusk shells, shell recrystallization, and poor age resolution in time intervals with minimal variation in the global seawater Sr-isotope curve [*McArthur et al.*, 2001]. Biostratigraphy is also hampered by sporadic preservation of microfossils and by low abundance of marker taxa with well-calibrated age ranges. While specimens from numerous microfossil groups are preserved in the cored VLB sediments, such as calcareous nannofossils [e.g., *Watkins and Villa*, 2000], dinoflagellate cysts and marine and terrestrial palynomorphs [*Hannah*, 1997; *Hannah et al.*, 2001; *Raine and Askin*, 2001], marine diatoms are the most abundant, best preserved and most age-diagnostic microfossil group in sediment cores from the VLB [e.g., *Harwood*, 1986, 1989; *Harwood et al.*, 1998; *Scherer et al.*, 2000; *Harwood and Bohaty*, 2001; *Olney et al.*, 2007]. Development of integrated chronologies, using all available age-diagnostic data, is the most fruitful way of

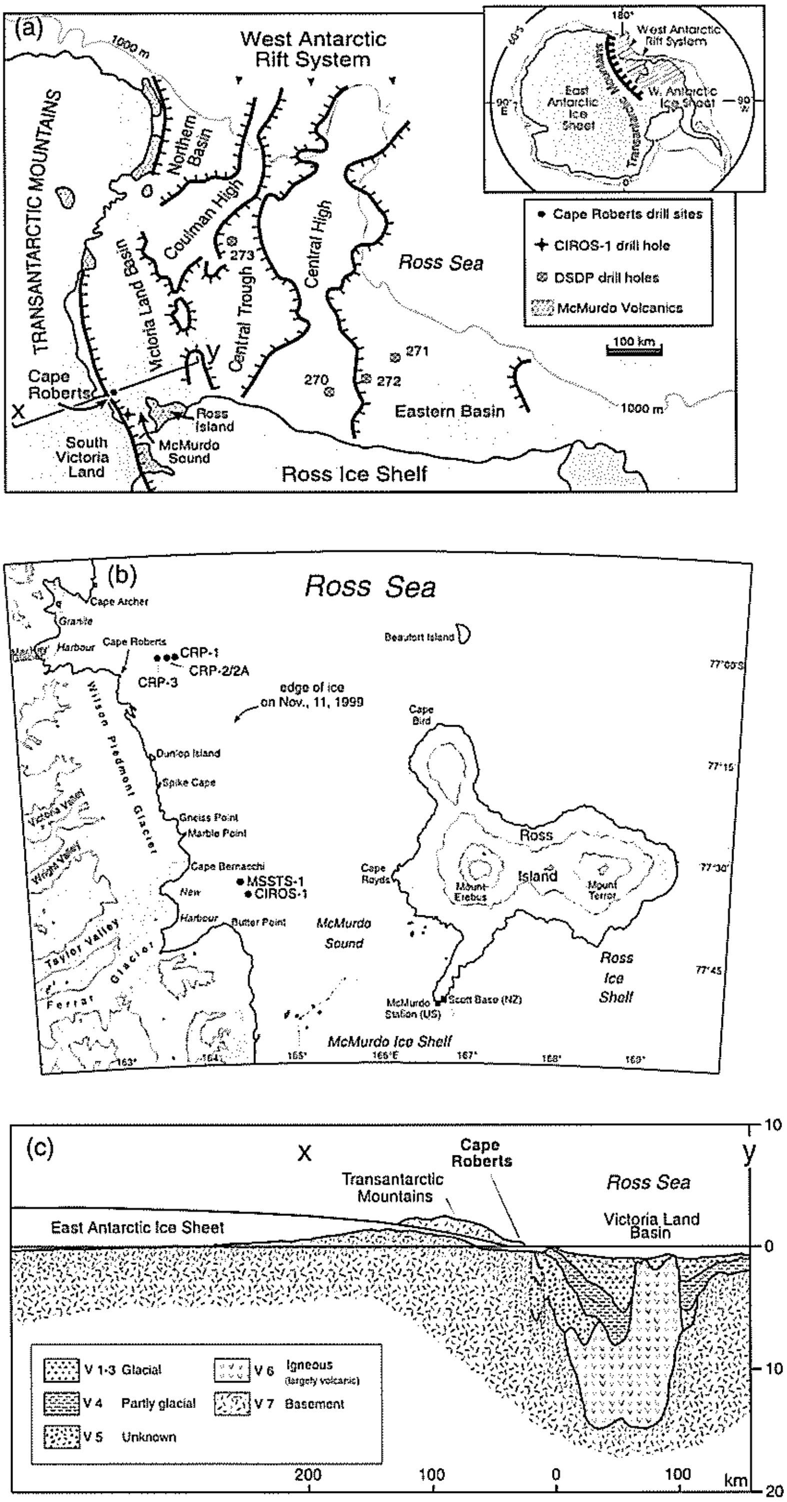

Figure 1. (a) Location map of the Victoria Land Basin with respect to the West Antarctic rift system (inset). (b) Map of McMurdo Sound with locations of the studied cores. (c) Schematic geological cross-section (along line X-Y in Figure 1a), with relationships between the Victoria Land Basin, Transantarctic Mountains, and East Antarctic Ice Sheet. Figures 1a and 1c are after *Barrett et al.* [1995].

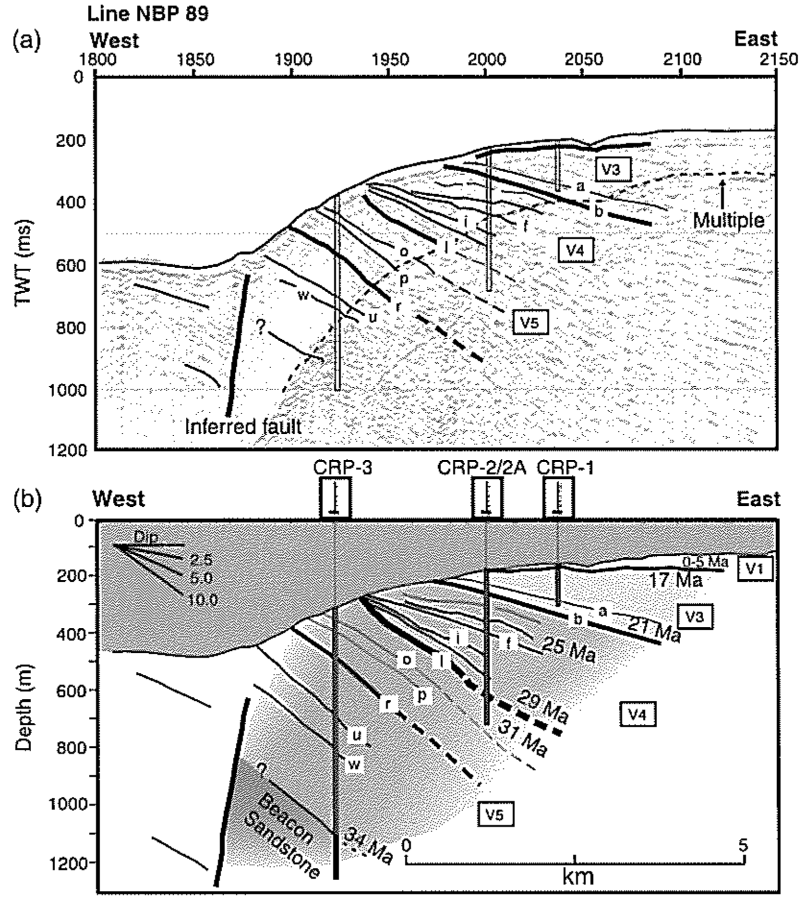

Figure 2. (a) Interpretation of single-channel seismic line NBP-89, from offshore of Cape Roberts (Figure 1b), with configuration of seaward-dipping seismic reflectors sampled in the CRP-1, CRP-2/2A, and CRP-3 drill holes, after *Hamilton et al.* [2001] and *Henrys et al.* [2001]. (b) Depth-migrated section with age relationships of reflectors from Figure 2a. Interpretations of ages of the sediment packages are derived from the respective core chronologies [*Florindo et al.*, 2005]. The V1–V5 designations for seismic stratigraphic sediment packages (see also Figure 1c) follow *Barrett et al.* [1995], while the a–w designations for seismic reflectors follow *Henrys et al.* [2001].

deriving accurate chronostratigraphies for these glacigene sediments. This work is complemented by sedimentological analysis, which provides evidence for glacial erosion and associated sedimentary hiatuses [e.g., *Fielding et al.*, 2000].

[7] Despite the above-mentioned difficulties, the Oligocene-Miocene boundary interval from the VLB has been dated with a temporal resolution of better than 100 kyr, which is the first time that this resolution has been achieved for Antarctic margin glacial sediments of this age [*Naish et al.*, 2001; *Wilson et al.*, 2002; *Roberts et al.*, 2003a]. We focus here on longer-term environmental magnetic trends; the resolution of available age models is

Table 1. Cores with Eocene-Miocene Sediments From the Victoria Land Basin[a]

Name of Core	Year Drilled	Latitude; Longitude (°S/°E)	Water Depth (m)	Depth to Base of Core (m)	Recovery (%)	Age Range of Sediments (Ma)
MSSTS-1	1979	77°33′55″; 164°29′56″	195	227	56	late Oligocene–early Miocene
CIROS-1	1986	77°34′26″; 163°23′13″	197.5	702	98	late Eocene–early Miocene
CRP-1	1997	77.008°; 163.755°	153.2	147	86	early Miocene
CRP-2/2A	1998	77.006°; 163.719°	177.9	624	95	early Oligocene–early Miocene
CRP-3	1999	77.0106°; 163.6404°	295	939	97	late Eocene–early Oligocene
Total	—	—	—	2639	—	late Eocene–early Miocene

[a]Note that most cores have a thin veneer of Quaternary sediments that are not considered here. In the CRP-3 core, late Eocene sediments lie unconformably on Devonian(?) basement rocks of the Beacon Supergroup (Figure 3).

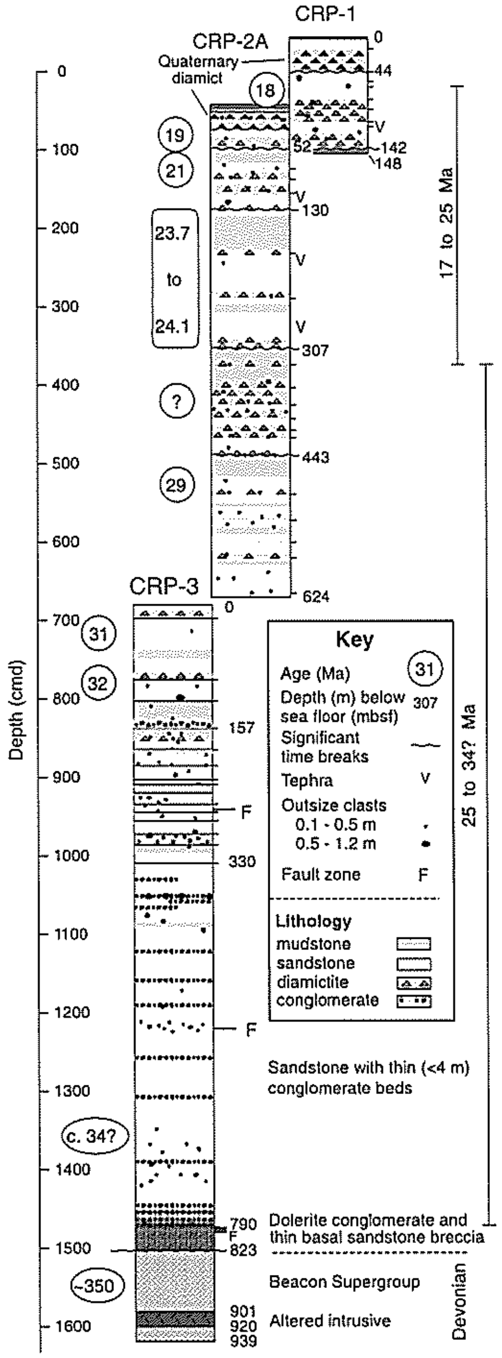

Figure 3. Summary of the stratigraphy of the CRP-1, CRP-2A, and CRP-3 cores from the Victoria Land Basin, Antarctica [after *Florindo et al.*, 2005]. Sediments span the age interval from the late Eocene to the early Miocene. Depths below seafloor are shown on the right-hand side of each stratigraphic column; total meters composite depth (mcd) for the three CRP cores is shown on the left-hand side of the figure. Stratigraphic details of the MSSTS-1 and CIROS-1 cores are given by *Barrett* [1986, 1989], respectively, and are not replicated here. Details of the total cumulative length of all studied cores (2.6 km) are given in Table 1.

adequate for our purposes. Details of the chronologies are given in the following papers for each core: CIROS-1 [*Wilson et al.*, 1998; *Roberts et al.*, 2003a], CRP-1 [*Roberts et al.*, 1998], CRP-2/2A [*Wilson et al.*, 2000a, 2000b], and CRP-3 [*Florindo et al.*, 2001; *Galeotti et al.*, 2012]. An overview of the chronology for CRP-1, 2/2A, and 3 (Figure 3) is provided by *Florindo et al.* [2005]. We use these chronologies without further comment unless explicit discussion is warranted.

2.3. Source Rocks for Victoria Land Basin Sediments

[8] The Transantarctic Mountains and local volcanic outcrops of southern McMurdo Sound are the dominant sources for sediment deposited in the VLB (Figure 1). We briefly describe the regional geology to provide an overview of the sources of magnetic particles into the VLB. Generalized geological relationships are shown in a schematic cross-section in Figure 4.

[9] The oldest rocks in southern Victoria Land consist of multiply deformed Precambrian metasediments of the Koettlitz Group [*Grindley and Warren*, 1964], which crop out along the foothills of the Transantarctic Mountains in coastal southern Victoria Land. The metamorphic rocks are intruded by lower Paleozoic plutonic rocks (granitoids), which are collectively referred to as the Granite Harbor Intrusive Complex [*Gunn and Warren*, 1962]. Basement metamorphics and the Granite Harbor intrusives record events associated with the Paleozoic Ross Orogen; the intrusives mark a long-lasting phase of Cambrian-Ordovician subduction at the culmination of the orogen [*Stump*, 1995]. After cessation of plutonic activity associated with the Granite Harbor intrusives, these rocks were exhumed and eroded. During this post-tectonic phase, basement rocks were cut by a sharp erosion surface [*Gunn and Warren*, 1962], which is referred to as the Kukri Peneplain [*Barrett et al.*, 1972]. The Kukri Peneplain is overlain by flat-lying Devonian to Jurassic continental sediments of the Beacon Supergroup [*Barrett et al.*, 1972; *Barrett*, 1981]. Beacon sediments, which can reach thicknesses of 3000 m, contain large amounts of cross-bedded quartzose sandstone, interbedded with red and green siltstones. These sediments were deposited in a variety of settings, including eolian, fluvial, lacustrine, swamp, and subglacial environments. The distribution of Beacon rocks suggests that they formed a fringe around the margin of East Antarctica and were derived from erosion of quartzose basement rocks in the continental interior. The Beacon Supergroup provides important evidence for widespread Permian glaciation of Gondwana [*Barrett et al.*, 1972; *Barrett and McKelvey*, 1981].

[10] Basement and Beacon Supergroup rocks were intruded by thick dolerite sills in the middle Jurassic. Some of this material erupted subaerially as basaltic lavas. The intrusives are referred to as the Ferrar Dolerites [*Harrington*, 1958], while the eruptive rocks are referred to as the Kirkpatrick Basalts [*Grindley*, 1963]; together, these units form the Ferrar Group [*Grapes et al.*, 1974]. The Ferrar Group crops out along 3000 km of the Transantarctic Mountains and comprises a short-lived continental flood basalt province (duration of ~1 Myr at about 176 Ma) that is temporally related to Gondwana breakup [*Heimann et al.*, 1994; *Encarnación et al.*, 1996; *Elliot et al.*, 1999; *Elliot and Fleming*, 2000].

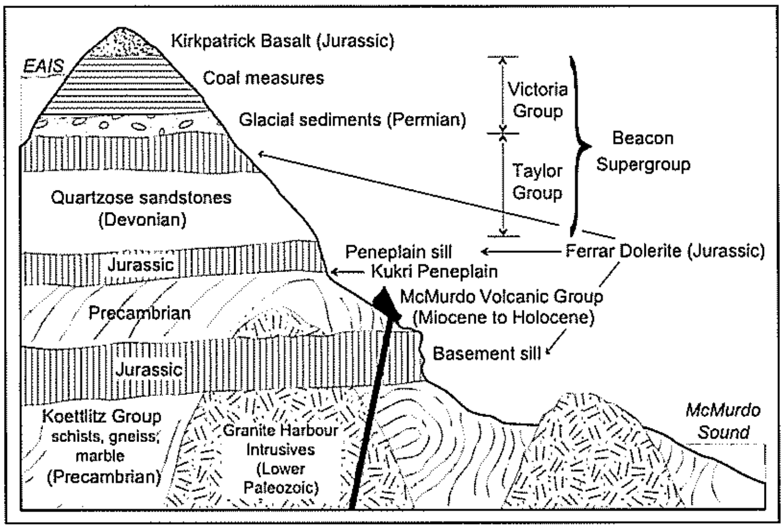

Figure 4. Generalized geological cross-section from the East Antarctic Ice Sheet (EAIS), through the Transantarctic Mountains to McMurdo Sound. After *Campbell and Claridge* [1987]. The stratigraphic subdivision of the Beacon Supergroup is after *McKelvey et al.* [1970].

[11] Apart from sporadic outcrops of the Sirius Group [e.g., *Webb and Harwood*, 1991], there are no rock outcrops younger than the Jurassic Ferrar Group in the VLB catchment area until the onset of activity associated with the McMurdo Volcanic Group. In southern Victoria Land, the earliest products of the McMurdo Volcanic Group are dated at 24 Ma [*McIntosh*, 2000]. The McMurdo volcanics [*Harrington*, 1958] consist of varied basic extrusives that occur along the eastern front of the Transantarctic Mountains and result from crustal extension associated with the West Antarctic rift system [*Kyle*, 1990]. The Ferrar Group and the McMurdo Volcanic Group are the most likely sources of fine-grained magnetite in VLB sediments. In contrast, clean quartzose sediments of the Beacon Supergroup would not be expected to contribute significantly to the flux of magnetic minerals to the VLB.

2.4. Weathering in Antarctica

[12] *Campbell and Claridge* [1987] provided a detailed synthesis of the present weathering regime and soil formation in the cold, arid climate of Antarctica. This is useful for our environmental magnetic investigation in which we test whether magnetic properties can provide a proxy for intensity of chemical weathering in the Transantarctic Mountains during the Eocene greenhouse versus physical weathering in the current icehouse environment. As is the case in other parts of the world, rock type strongly influences the rate at which rocks weather and the end products of weathering processes [*Leeder*, 2011]. However, climate is the most important factor for determining the rate of weathering; the importance of parent rock variations is much reduced by the weak present-day weathering environment of Antarctica. The climate of Antarctica is the coldest and driest in the world. Virtually all water is frozen and is unavailable to contribute to weathering processes (except for subglacial processes). Chemical weathering reactions are inhibited and occur at extremely slow rates. Low moisture availability and low temperatures mean that the major process by which rocks disintegrate in Antarctica is physical weathering via glacial and wind action, insolation, salt weathering, and action of water in various forms (liquid, ice, and vapor) [*Campbell and Claridge*, 1987].

[13] Fission track data from apatite grains indicate that substantial crustal denudation occurred since the onset of Transantarctic Mountain uplift at ~55 Ma [*Fitzgerald*, 1992]. Much of the development of the present landscape in the Transantarctic Mountains probably occurred via fluvial processes before buildup of continental ice sheets [*Sugden et al.*, 1995, 1999]. It is argued that subsequent glacial activity had a limited impact on the landscape and that many elements of a preglacial fluvial landscape have been preserved [*Sugden et al.*, 1999]. While there has been heated debate about EAIS dynamics [e.g., *Webb and Harwood*, 1991; *Marchant et al.*, 1993], glacial landscape modification has been minimal in many areas. For example, cosmogenic isotopes provide late Miocene exposure ages for some geomorphic surfaces, which suggest that parts of the Dry Valleys in the Transantarctic Mountains are the oldest exposed landscapes on Earth [*Schäfer et al.*, 1999]. Thus, while physical weathering dominates the process of rock decay in Antarctica, it is much less intensive than in alpine or subalpine zones of more temperate areas such as New Zealand or South America where frost cracking, exfoliation, and rock disintegration are more effective because of abundant moisture supply [*Campbell and Claridge*, 1987].

[14] Evidence of chemical weathering is also present in Antarctica, mainly in the form of red- or brown-colored staining on rocks. When such stainings are examined closely, the effects of chemical weathering are not evident. For example, *Kelly and Zumberge* [1961] studied variably weathered quartz diorite samples from Marble Point, McMurdo Sound (Figure 1b). Despite marked differences in hardness, consistency, and color of samples, bulk chemical compositions were nearly constant for hard, fresh samples compared to weathered, crumbling, and iron-stained rock. No evidence was found for weathering-related clay mineral formation; the only observed chemical change was

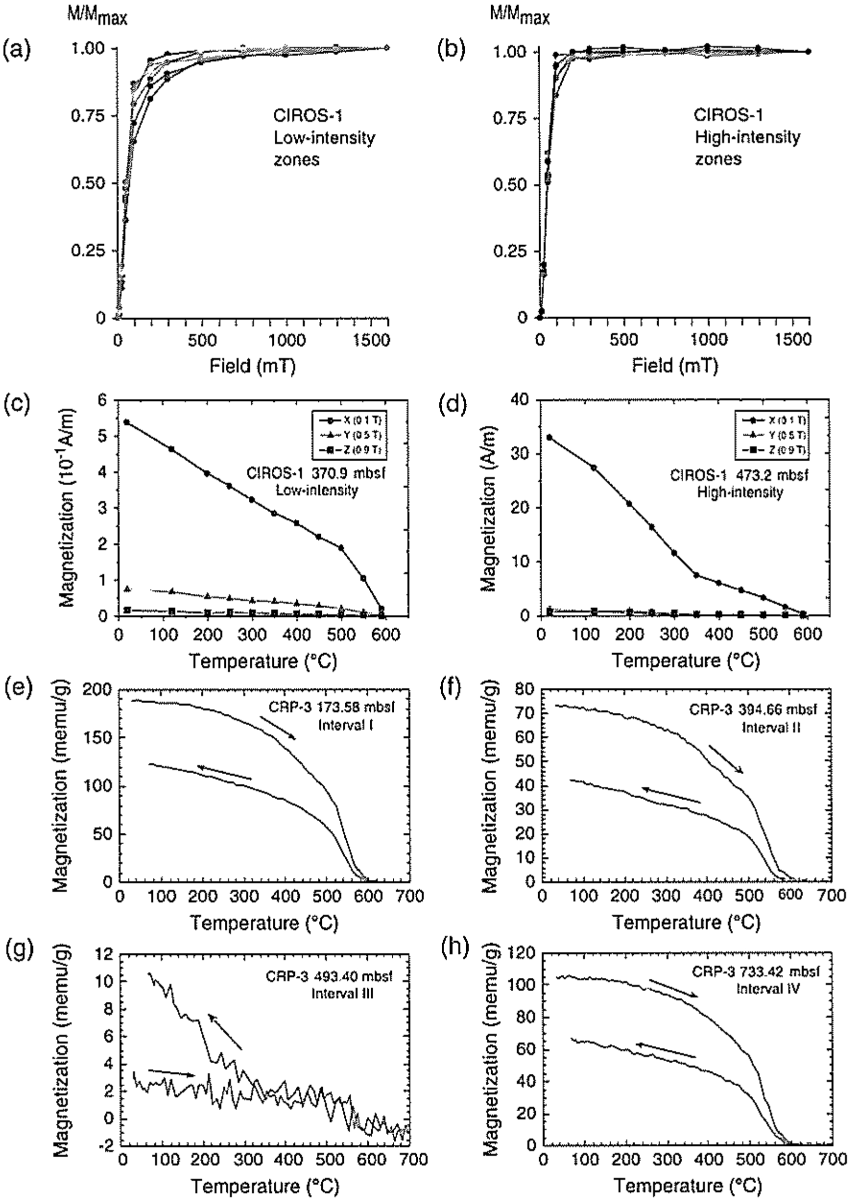

Figure 5. Representative data concerning the dominant magnetic minerals in studied sediments from the Victoria Land Basin. The magnetic properties of most of the studied cores alternate between low magnetic intensity and high magnetic intensity zones. For the CIROS-1 core, IRM acquisition data indicate (a) the combined presence of a low- and high-coercivity component in low magnetic intensity zones and (b) a dominant low-coercivity component in high magnetic intensity zones. Thermal demagnetization of a composite 3-axis IRM [*Lowrie*, 1990] indicates that the magnetization is dominated by magnetite in both the low-intensity and high-intensity zones. A hematite component is also likely in the low-intensity zones (see Figure 5a), but this is not evident in Figure 5c. The inflection at 350°C in the low-coercivity component for samples from (d) the high-intensity zones indicates the likely presence of maghemite (probably due to surficial oxidation of magnetite grains). The CRP-3 core is subdivided into four zones. (e–h) Samples from all zones yield thermomagnetic curves (heating and cooling in air) with Curie temperatures at 580°C, which indicates that magnetite dominates the magnetic properties of the CRP-3 core.

precipitation of ferric oxides as surficial coatings. This staining makes the rocks appear more weathered. Investigations of thin sections from other rock types, including dolerite and sandstone, confirm the thin and surficial nature of the coatings [*Campbell and Claridge*, 1987]. *Glasby et al.* [1981] used electron probe microanalysis to confirm the enrichment of iron in weathered surface layers; however, specific minerals were not positively identified. *Campbell and Claridge* [1987]

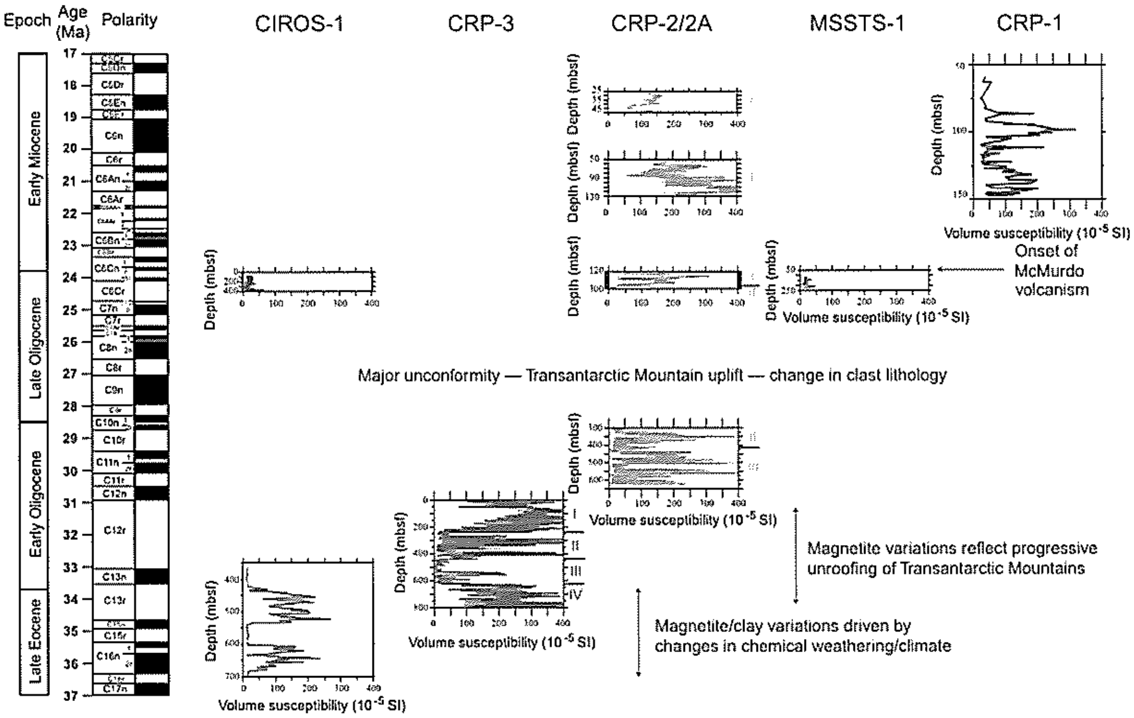

Figure 6. Summary of magnetic mineral concentration variations (indicated by the low-field volume susceptibility, κ) with respect to age for the studied cores from the Victoria Land Basin. The geomagnetic polarity timescale of *Cande and Kent* [1995] is shown on the left-hand side. Each stratigraphic record is truncated by multiple hiatuses due to ice-advance events; we only indicate age offsets for the largest hiatuses. Age models are from *Florindo et al.* [2005] for the CRP cores, *Wilson et al.* [1998] for the lower CIROS-1 core, *Roberts et al.* [2003a] for the upper CIROS-1 core, and unpublished diatom evidence for MSSTS-1 that indicates that it records a narrow Oligocene-Miocene boundary interval as recorded in CIROS-1 and CRP-2/2A.

state that the iron oxides are likely to be amorphous, with the colors of the staining suggesting the presence of goethite-like to anhydrous hematite-like minerals. These stainings are best developed in the most arid regions of Antarctica and on surfaces that have been exposed for the longest periods.

[15] The most important implication of modern weathering studies for environmental magnetism is that the VLB could have received fine-grained hematite and goethite inputs that could have resulted from long-term (millions of years) weathering. Such minerals would therefore provide a time-integrated signal and are unlikely to be representative of continental climate at the time of deposition. Magnetic properties of all of the studied VLB cores undergo marked alternations between intervals with high and low detrital magnetite concentrations [*Sagnotti et al.*, 1998a, 1998b, 2001a; *Verosub et al.*, 2000; *Roberts et al.*, 2003a]. High-coercivity minerals (hematite and goethite) have weak susceptibility and magnetization compared to magnetite. Thus, when magnetite is present, large quantities of hematite or goethite are necessary to be magnetically detectable with most magnetic measurements. We only observe significant hematite concentrations in VLB cores in intervals where magnetite concentrations are low [e.g., *Verosub et al.*, 2000]. Alternations between zones with high and low susceptibilities and magnetizations in VLB cores are linked to magnetite concentration variations [*Sagnotti et al.*, 1998a, 1998b, 2001a; *Verosub et al.*, 2000; *Roberts et al.*, 2003a]. By focussing on environmental signals that result from detrital magnetite content variations, we avoid complications that could have resulted from the long time spans inherent to the present Antarctic weathering regime, for which the main magnetic by-products appear to be hematite and goethite [*Campbell and Claridge*, 1987].

3. Environmental Magnetism of Sediments From the Victoria Land Basin

3.1. Magnetic Mineralogy

[16] Extensive rock magnetic studies of Eocene-early Miocene VLB sediments indicate that magnetite dominates the magnetic properties of the sediments [*Sagnotti et al.*, 1998a, 1998b, 2001a; *Verosub et al.*, 2000; *Roberts et al.*, 2003a]. Stepwise acquisition of an isothermal remanent magnetization (IRM) indicates that the magnetic minerals consistently reach magnetic saturation at low applied fields (<300 mT; Figure 5b), although higher fields are often required to achieve saturation in the low magnetic intensity zones (Figure 5a). This indicates the presence of some hematite, although most of the magnetization in such samples decays to near-zero values below 600°C (Figure 5c), which indicates that magnetite dominates the magnetization even in the low magnetic

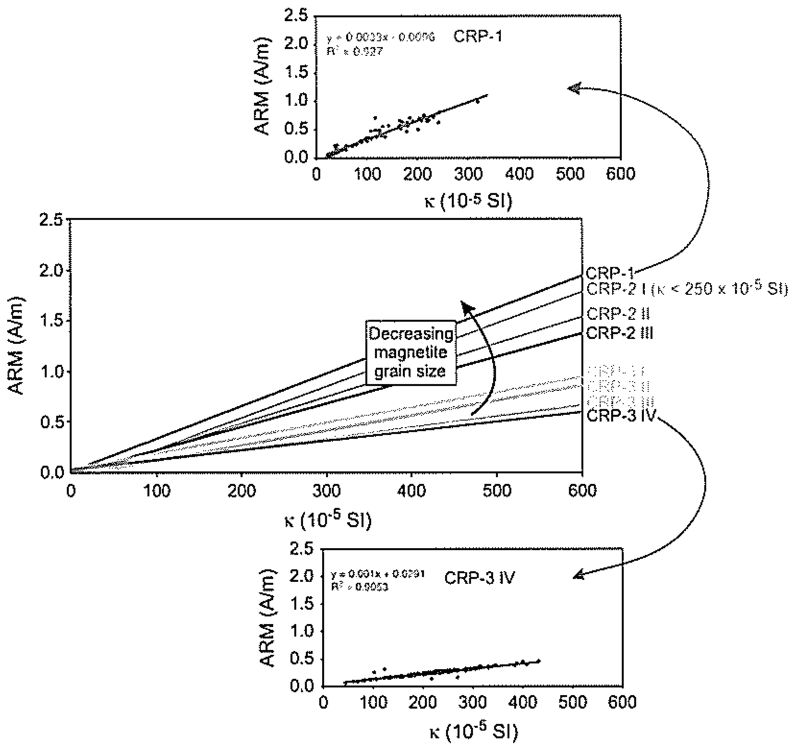

Figure 7. Evidence for progressive fining upward of magnetite grain size through the VLB successions recovered in CRP-3, CRP-2/2A, and CRP-1. The lower and upper plots of ARM versus susceptibility (κ) contain details of the well-correlated linear trends for CRP-3 zone IV and CRP-1, respectively, which illustrate the upward fining of magnetite grain size.

intensity zones. Maximum unblocking temperatures of 580°C are consistently observed in a range of high-temperature analyses for both high and low magnetic intensity zones (Figures 5c–5h). These properties are consistent with and indicative of those expected for magnetite. In samples from high-intensity zones in CIROS-1, inflections are observed in high-temperature data at 300–400°C (Figures 5d–5f and 5h), which indicates that magnetite has been partially maghemitized. The inflection is due to inversion of maghemite to hematite [*Deng et al.*, 2001; *Liu et al.*, 2005]; the weaker intrinsic magnetization of hematite probably also explains the decreased magnetizations of the thermomagnetic cooling curves compared to the heating curves (Figures 5e, 5f, and 5h). The presence of maghemite suggests that the more intense chemical weathering in these high magnetic intensity intervals [*Sagnotti et al.*, 1998a] caused oxidation of magnetite, which is not evident in the younger CRP cores that were deposited under conditions with much reduced or no chemical weathering. Magnetic properties of the studied cores undergo marked stratigraphic variations between periods with high and low magnetite concentrations (Figure 6). In stratigraphic intervals where magnetite concentrations are low, hematite often becomes relatively more important in terms of its contribution to the overall magnetization, particularly in CRP-1 [*Sagnotti et al.*, 1998b] and CRP-2/2A [*Verosub et al.*, 2000], although this trend is not observed in CRP-3 [*Sagnotti et al.*, 2001a]. Magnetite remains the dominant magnetic mineral in the more weakly magnetized stratigraphic intervals (Figures 5c and 5g); the contribution of hematite to the total magnetization is never sufficient to equal or exceed that due to magnetite, as indicated by a lack of wasp-waisted hysteresis loops [cf. *Roberts et al.*, 1995]. In our remaining discussion, we focus on interpretation of stratigraphic variations in magnetite concentration and grain size because magnetite is the dominant magnetic mineral responsible for the major environmental magnetic variations observed in the studied cores. This approach also avoids complexities due to long-term weathering that is potentially associated with slow hematite formation in cold and arid Antarctic environments, as discussed above.

[17] In addition to abundant detrital magnetite, *Roberts et al.* [2012] recently documented evidence for the presence of biogenic magnetite in the studied VLB cores. While this is an interesting development in terms of the widespread occurrence of biogenic magnetite, and for documenting magnetotactic bacteria in such highly sediment-charged glaci-marine environments, the concentration of biogenic magnetite particles is likely to be low compared to detrital magnetite particles. We therefore do not consider biogenic magnetite further in our discussion below.

[18] Greigite was observed in a single restricted stratigraphic interval at the base of CRP-1 (143.8 to 147.69 m below seafloor (mbsf)) [*Sagnotti et al.*, 2005]. Detailed analyses indicate that the greigite formed during late diagenesis, which remagnetized sediments in this thin stratigraphic interval [*Sagnotti et al.*, 2005], probably

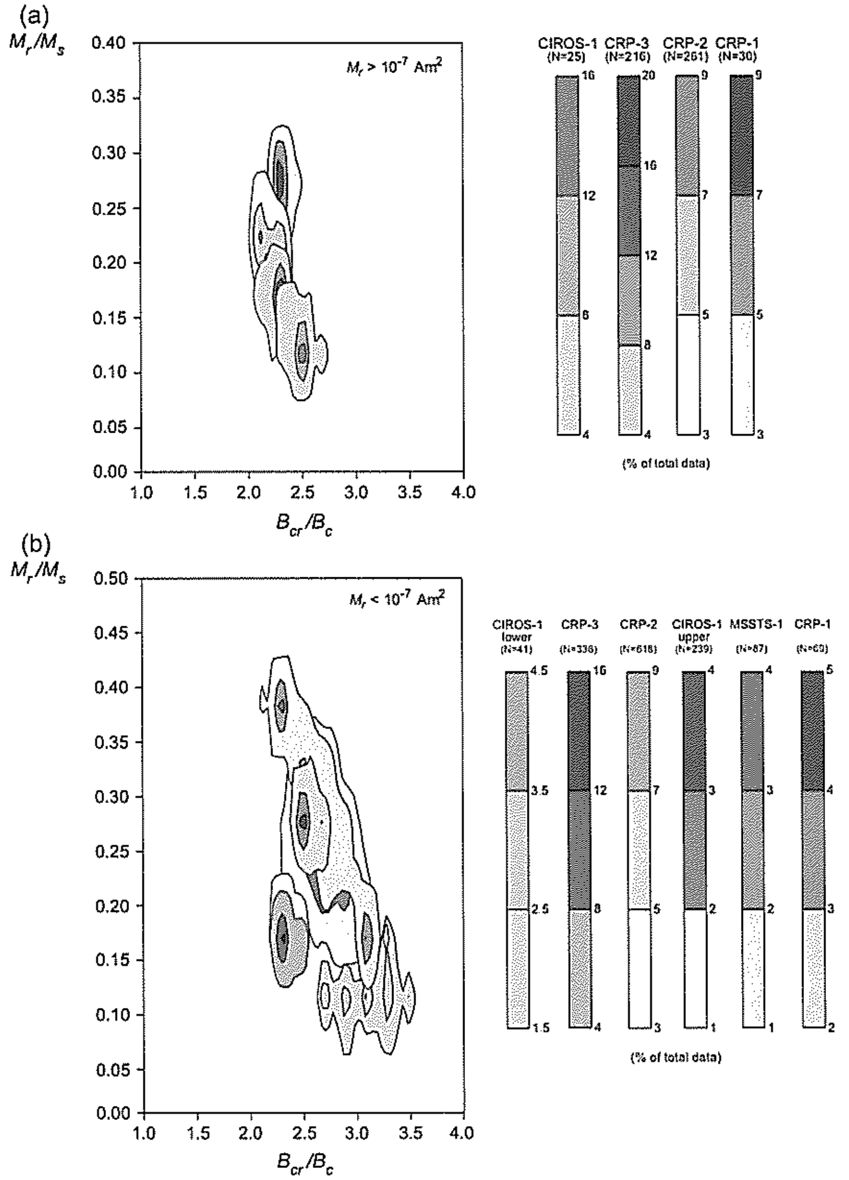

Figure 8. Contour plots of hysteresis ratio data (following *Day et al.* [1977]) for all studied holes from the VLB. For samples from (a) high magnetic intensity zones (using a cutoff value of 10^{-7} Am^2), the data are well clustered and have a clear trend with fining grain size from older to younger. Data from the (b) low-intensity zones are more scattered but still provide evidence for the overall fining from older to younger, although data from CIROS-1 and MSSTS-1 have far greater spread. Density distributions for hysteresis ratios were determined by counting the number of data points within cells with dimensions of 0.2 (for B_{cr}/B_c values between 1 and 5) $\times$ 0.02 (for M_r/M_s values between 0 and 0.5) on the Day plot. The number of data points in each cell is expressed as the percentage of the total number of data points for each core. The density distributions were then contoured using a Kriging algorithm. The color bars on the right-hand side indicate the percentage of data clustered in individual unit cells. For clarity, contours are only shown for values above a minimum threshold (data below these thresholds appear as a blank background in the plots).

through sulphidization associated with the former presence of methane hydrates (see *Roberts and Weaver* [2005] for details). This is the only interval in which greigite has been identified in the ~2.6 km stratigraphic thickness of sediment analyzed. We therefore do not consider this interval further.

3.2. Changes in Magnetite Concentration Through Time

[19] Variations in magnetite concentration dominate the environmental magnetic signal of VLB sediments and are best indicated by low-field magnetic susceptibility (κ) or

anhysteretic remanent magnetization (ARM). A synthesis of κ variations is plotted against age for the respective cores in Figure 6. Large-scale stratigraphic cycles are observed in most cores (Roman numerals denote the zones in CRP-2/2A and 3; *Verosub et al.* [2000] and *Sagnotti et al.* [2001a]). Higher frequency signals are also observed, although the nature and origin of such signals are not straightforward to understand because, as discussed above, Antarctic glaci-marine sediments of this age are notoriously difficult to date. As a result, we restrict our focus to large-scale (usually 100 m scale) stratigraphic variations in magnetic properties. While we do not attempt to represent stratigraphic gaps associated with multiple small hiatuses in Figure 6, a large angular unconformity is recorded at a depth of 306.65 mbsf in CRP-2/2A, which has produced an almost 5 Myr gap through the late Oligocene stratigraphic record [*Wilson et al.*, 2000a, 2000b]. A major change in clast lithology is recorded across this unconformity [*Talarico et al.*, 2000], which is interpreted to represent a major phase of uplift of the adjacent Transantarctic Mountains [*Smellie*, 2000]. As mentioned above, a major pulse of sedimentation is recorded in the VLB near the Oligocene-Miocene boundary [*Naish et al.*, 2001; *Wilson et al.*, 2002; *Roberts et al.*, 2003a], which is followed by a time interval for which no sediments are preserved in the studied cores (Figure 6). A period of deposition is then recorded between ~21 and 17 Ma [*Florindo et al.*, 2005]. We explore below (section 4) some of the likely causes of the large-scale magnetite concentration variations documented in Figure 6.

3.3. Changes in Magnetite Grain Size Through Time

[20] In addition to the observed large-scale temporal changes in magnetite concentration, we observe a systematic fining of magnetite grain size as the VLB sediments get progressively younger. We illustrate this with two sets of magnetic parameters that are sensitive to magnetite grain size variations: ARM/κ [*Banerjee et al.*, 1981; *King et al.*, 1983] and hysteresis parameters [*Day et al.*, 1977; *Dunlop*, 2002]. In Figure 7, we plot ARM versus κ to illustrate the observed systematic fining of grain size with time in the studied CRP cores. Roman numerals denote zones with distinct magnetite concentration variations; the respective age relationships can be visualized in Figure 6. Data from CRP-3 zone IV and CRP-1 are shown to demonstrate the quality of the linear fits to the ARM-κ data. From the ages of core intervals in Figure 6, it can be seen that the slope of the linear fits to the raw ARM-κ data in Figure 7 progressively increases (indicating progressively finer magnetite grain size) from the older to the younger CRP core intervals. A similar fining-upward trend for magnetite is evident in plots of the ratios of saturation remanence (M_r) to saturation magnetization (M_s) versus coercivity of remanence (B_{cr}) to coercive force (B_c) [*Day et al.*, 1977] (Figure 8). We plot data separately for the strongly (Figure 8a) and weakly magnetized (Figure 8b) zones in the respective cores, where the cutoff M_r value for the two groups is 10^{-7} Am^2. Instead of replotting the individual data, which are plotted in the original studies of the respective cores, we illustrate overall data distributions for these datasets, with contours of the distributions to illustrate trends (Figure 8). There is greater data scatter for weakly magnetized samples (Figure 8b), but the trend to finer overall grain sizes with time is evident in both the weakly and strongly magnetized subgroups. All samples from MSSTS-1 and the upper part of CIROS-1 fall in the weakly magnetized subgroup, which means that fewer cores are represented in Figure 8a. Overall, the ARM-κ and hysteresis data indicate a progressive upward fining of magnetite grain size.

4. Discussion

4.1. Influence of Diagenesis on the VLB Sediments

[21] The gray muddy sediment matrix that was sampled for our studies consistently contains diagenetic pyrite. Pyrite formation occurs at the expense of detrital iron-bearing phases, and reductive dissolution of magnetite is ubiquitous in sediments in which pyrite forms [e.g., *Canfield and Berner*, 1987]. In most marine sedimentary environments, progressive sulphidization leads to dissolution of much of the detrital magnetite assemblage [e.g., *Karlin and Levi*, 1983, 1985; *Karlin*, 1990a, 1990b; *Leslie et al.*, 1990; *Channell and Hawthorne*, 1990; *Rowan et al.*, 2009]. *Karlin and Levi* [1985] and *Karlin* [1990a] argued that detrital magnetite can survive diagenetic dissolution if it occurs in sufficient initial concentrations to exhaust the sulphidization process. The studied VLB sediments were deposited in ice-proximal glaci-marine environments with almost completely siliciclastic compositions (with small biogenic siliceous and carbonate fossil contributions). Many of the lithologies from the source terrains that were eroded to provide detritus to the VLB contain high magnetite concentrations (Figure 4). Organic carbon contents in the studied cores are modest but nonzero because, unlike today, the Antarctic margin in Eocene-Miocene times was vegetated to varying extents [e.g., *Hill*, 1989; *Mildenhall*, 1989; *Raine*, 1998; *Roberts et al.*, 2003a; *Pross et al.*, 2012]. Thus, although organic carbon diagenesis has led to pyrite formation, which will have been accompanied by magnetite dissolution, magnetite concentrations in most of the VLB sediments are so high (Figure 6) that sulphidization would have been arrested by high concentrations of reactive iron [*Kao et al.*, 2004]. Magnetite therefore remains abundant in the studied sediments. Correlation between variations in magnetite concentration and the number of Ferrar Dolerite clasts in CRP-3 [e.g., *Sagnotti et al.*, 2001a] supports the interpretation that magnetite supply was high enough for magnetite dissolution not to have significantly modified the terrigenous signal carried by detrital magnetite. Additionally, even the lowest κ values illustrated in Figure 6 are higher than, or similar to, those in diagenetically unmodified modern continental margin sediments from the Oman and Northern California margins [*Rowan et al.*, 2009]. We therefore consider magnetite dissolution to be a secondary factor that has modified, but not controlled, magnetite concentration variations.

[22] As mentioned in section 3.1, we have documented a single thin stratigraphic interval at the base of CRP-1 in which greigite formed [*Sagnotti et al.*, 2005]. An additional interval in CRP-3 (at around 550 mbsf; zone III in Figure 6) with high magnetite concentrations is close to a fault where there is a pronounced local temperature anomaly [*Bücker et al.*, 2001]. *Sagnotti et al.* [2001a] considered it likely that localized fluid flow, which probably gave rise to siderite

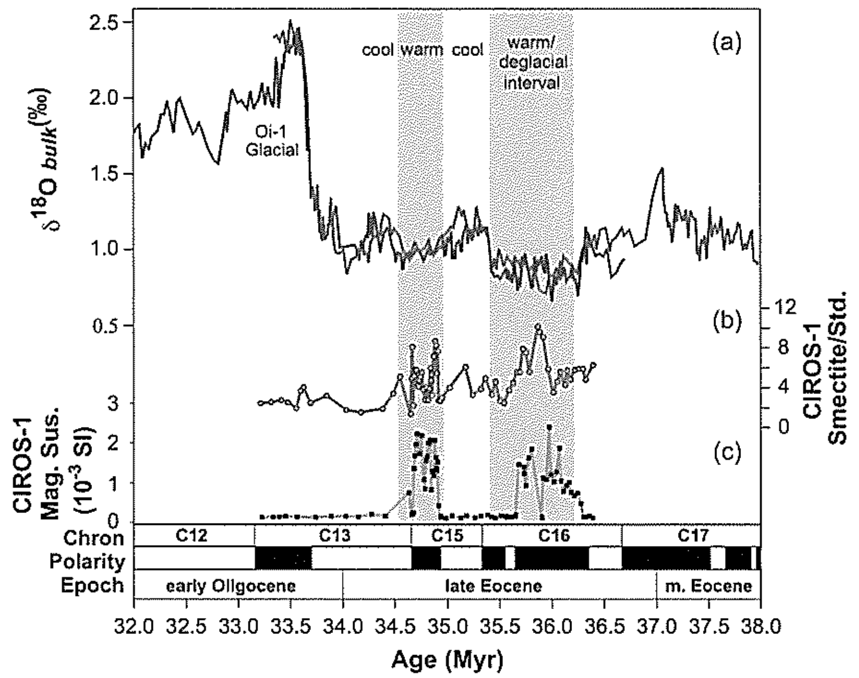

Figure 9. Verification that late Eocene smectite/magnetite variations from the CIROS-1 core and their inferred response to changes in Antarctic weathering regime [*Sagnotti et al.*, 1998a] are due to large-scale climatic variations. (a) Bulk carbonate $\delta^{18}O$ records from ODP holes 738B and 744A (Bohaty and Zachos, manuscript in preparation). (b) Smectite content (smectite/standard ratio) [*Ehrmann*, 1998] and (c) magnetic susceptibility record for the CIROS-1 core [*Sagnotti et al.*, 1998a]. The light blue and red vertical shading indicates fluctuations between cool and warm climatic periods that are interpreted to reflect Antarctic glacial and interglacial conditions, which provide independent confirmation of cyclical climatic changes that coincided with periods of increased physical and chemical weathering, respectively, on Antarctica [*Sagnotti et al.*, 1998a]. The geomagnetic polarity timescale at the base of the figure is from *Cande and Kent* [1995].

formation, could have caused diagenetic modification of the environmental magnetic signal in this interval. We do not consider these diagenetically modified intervals any further.

4.2. Magnetite Concentration as an Indicator of Antarctic Weathering Regime?

[23] *Sagnotti et al.* [1998a] observed a striking similarity between late Eocene variations in detrital smectite concentrations and crystallinity and magnetite concentrations (Figures 9b and 9c) in the lower part of the CIROS-1 core. They attributed the observed large-scale cycles to fluctuations between periods dominated by either chemical or physical weathering of basic igneous rocks (Ferrar Dolerite; Figure 4). During warmer, more humid, climatic periods, chemical weathering gave rise to formation of more highly crystalline smectite [*Ehrmann*, 1998]. Under these conditions, magnetite could be more readily released from the host rock. During cooler, drier periods, physical weathering was dominant and less effective in releasing magnetite from the source rocks and did not give rise to smectite formation. Further support for this hypothesis is provided by the fact that magnetite in the high magnetization intervals has been surficially oxidized to maghemite (Figures 5d–5f and 5h), which is more likely under warm/humid conditions than cold/arid conditions. While the hypothesized link between climate and weathering regime is attractive, the observations of *Sagnotti et al.* [1998a] have not yet been upheld by independent observations. A difficulty relates to the chronology of the CIROS-1 core. The magnetobiostratigraphic chronology of *Wilson et al.* [1998] was challenged by *Hannah et al.* [1997] and *Watkins* [2007], although neither set of age constraints have greater chronostratigraphic precision than those of *Wilson et al.* [1998]. The question concerning a link between smectite/magnetite cyclicity and weathering/climate, therefore, remains tantalizing but undemonstrated.

[24] In Figure 9, we use the chronology of *Wilson et al.* [1998] and correlate results from CIROS-1 with those from Ocean Drilling Program (ODP) sites 738 and 744 at Kerguelen Plateau, Southern Ocean, to further test this possibility. A bulk carbonate $\delta^{18}O$ record (Bohaty and Zachos, unpublished data) is presented using the paleomagnetic polarity record of *Roberts et al.* [2003b] to provide a stratigraphic framework for middle Eocene to early Oligocene climatic variations in the Southern Ocean. Periods with high smectite [*Ehrmann*, 1998] and magnetite [*Sagnotti et al.*, 1998a] concentrations in CIROS-1 (Figures 9b and 9c) coincide with two relatively long (~200–600 kyr) late Eocene warm periods indicated in the bulk $\delta^{18}O$ record (Figure 9a), while periods with low smectite and magnetite concentrations coincide with cooler intervals. These results, therefore, provide chronologically robust evidence to

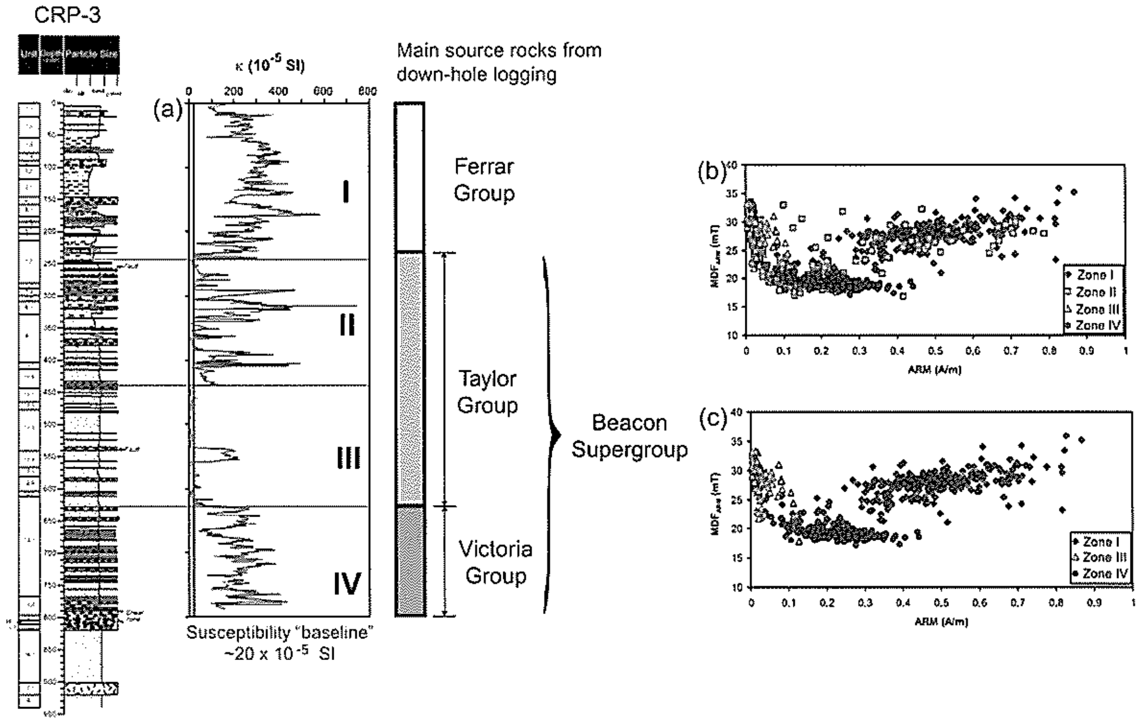

Figure 10. (a) Magnetic zonation of CRP-3. An approximately equivalent zonation is evident from multivariate statistical analysis of down-hole geophysical logs [*Bücker et al.*, 2001] and sandstone petrography [*Smellie*, 2001], which is interpreted to represent provenance variations from local source terrains associated with syn-sedimentary uplift and unroofing of the Transantarctic Mountains (Figure 4). The Victoria and Taylor groups represent the upper and lower parts of the Beacon Supergroup [*McKelvey et al.*, 1970]. (b, c) Plots of MDF_{ARM} versus ARM are used to discriminate the magnetic properties of zones I, III, and IV, with zone II representing a mixture of materials sourced from the other three zones. The environmental magnetic data reflect provenance variations associated with progressive unroofing of the Transantarctic Mountains. See text for discussion.

support the age model of *Wilson et al.* [1998] and the interpretation of *Sagnotti et al.* [1998a] concerning the climate/chemical weathering and magnetite concentration relationship in CIROS-1.

[25] The link between magnetite concentration and weathering/climate indicated by clay minerals persists through the lower part of CIROS-1 and CRP-3 (although the dominance of authigenic rather than detrital smectite in the lower part of CRP-3 complicates interpretation) [*Ehrmann*, 2001] and from the base of CRP-2/2A up to a depth of 485 mbsf [*Ehrmann*, 2000]. Above this depth, clay mineral assemblages become smectite-poor with decreased crystallinity [*Ehrmann*, 2000]. This suggests that increased Antarctic ice cover resulted in a weathering regime that is dominated by physical (mechanical) rather than chemical processes. Initiation of persistent glaciation at the E-O transition meant that it was never subsequently warm and humid enough (since the early Oligocene) for chemical weathering to contribute significantly to clay mineral or environmental magnetic signals. We have no convincing evidence for a link between climate and magnetic properties for the VLB sedimentary record since the early Oligocene.

4.3. Magnetite Concentration as an Indicator of Transantarctic Mountain Unroofing?

[26] Given that we have no convincing evidence for a link between climate and magnetic properties for the VLB sedimentary record since the early Oligocene, we now assess whether magnetite concentration can be used as a provenance indicator in these younger sediments. The CRP-3 environmental magnetic record is dominated by large-scale alternations between four major zones with high and low magnetizations [*Sagnotti et al.*, 2001a]. *Bücker et al.* [2001] presented a multivariate statistical analysis of down-hole geophysical logging data and argued that CRP-3 could be subdivided into three major stratigraphic intervals that reflect variations in sediment provenance (Figure 10a). This interpretation is supported by petrographic observations [*Smellie*, 2001]. Based on geophysical logging, sandstones in lowermost CRP-3 (788 to 630 mbsf) are discriminated from those in the overlying interval using radiogenic and magnetic properties [*Bücker et al.*, 2001]; this boundary corresponds precisely with that between magnetic zones IV and III (Figure 10a). These two sandstones are derived from different parts of the Beacon Supergroup [*McKelvey et al.*, 1970]; zone IV sandstones contain a recognizable contribution from Triassic volcanogenic sandstones of the Lashley Formation (uppermost Victoria Group), while the less strongly magnetic zone III consists of clean sands derived from the quartzose Devonian Taylor Group [*Smellie*, 2001]. This difference between "dirty" Victoria Group sources and "clean" Taylor Group sources has long been

known [e.g., *Korsch*, 1974]. There is also a distinct magnetic change at the boundary between zones III and II, with the change from clean quartzose sands to increased contributions from basement granitoids and metamorphics giving rise to stronger magnetizations in zone II. *Bücker et al.* [2001] argued that mudstones in the uppermost part of CRP-3 are derived primarily from erosion of the Ferrar Group. Consistently strong magnetizations in zone I (Figure 10) reflect the basic igneous composition of Ferrar Group rocks.

[27] Documentation of erosional detritus derived from the youngest source rocks (Victoria Group) at the base of CRP-3 to older Taylor Group rocks in the middle of CRP-3 to increasing contributions from basement rocks (Granite Harbor Intrusives) in the upper part of CRP-3 reflects progressive early Oligocene uplift and unroofing of the Transantarctic Mountain cover succession (Figure 4). The fact that the youngest erosional detritus in CRP-3 (Jurassic Ferrar Group) does not follow the same age progression expected for unroofing of an orogen reflects the fact that the Ferrar Dolerites intrude at multiple stratigraphic positions within the Transantarctic Mountains (Figure 4). The dolerites are volumetrically dominant in the lower cover sequence, which explains why they contribute significantly to the later erosion phase recorded in CRP-3. CRP-3 therefore provides a clear environmental magnetic record of early Oligocene unroofing of the stratigraphic succession exposed in the Transantarctic Mountains.

[28] The unroofing history of the Transantarctic Mountains can be further explored with more sophisticated magnetic parameters than κ alone, as shown in Figure 6. Plots of the median destructive field of the ARM (MDF_{ARM}; i.e., the alternating magnetic field at which half of the ARM has been destroyed) versus ARM (Figures 10b and 10c) contain three main data clusters from zones IV, III, and I. The more variable magnetizations in zone II represent a mixture between the data distributions of zones IV, III, and I. Mixing of materials from these zones within zone II is possible because the Taylor and Victoria groups had already been unroofed at that point, and Ferrar sources also intrude the younger rocks exposed in the Transantarctic Mountains (Figure 4).

[29] Magnetic discrimination of different zones in Figure 10 demonstrates the power of environmental magnetism for assessing provenance changes associated with progressive unroofing of an orogen. This is unlikely to be an isolated example. Regional metamorphic belts in particular will contain distinct magnetic mineral assemblages depending on the metamorphic grade of the rocks being exhumed, so magnetic analysis should be useful for studying orogenic unroofing. The Taiwan orogen is a good example, where erosion of rocks from the "pyrrhotite isograd" of *Rochette* [1987] can be detected easily through occurrence of strongly magnetic monoclinic pyrrhotite in marginal sedimentary basins around Taiwan [e.g., *Horng and Roberts*, 2006; *Horng et al.*, 2012].

4.4. Environmental Significance of Magnetite Grain Size Variations?

[30] The nearly constant detrital mode compositions of sandstones from the upper part of CRP-3 and throughout CRP-2/2A [*Smellie*, 2000, 2001] indicate that the entire cover succession and basement were being eroded from the Transantarctic rift margin (Figure 4) by the time these sediments were deposited. The environmental magnetic record of these and younger sediments is therefore likely to reflect mixed provenance. Nevertheless, a clear additional environmental magnetic signal is observed throughout CRP-3, CRP-2/2A, and CRP-1: progressive upward fining of magnetite grain size (Figures 7 and 8). It is not surprising that magnetite grain size fines through CRP-3 because zones IV and III are relatively coarse-grained, and the upper parts of CRP-3 (zones II and I) contain progressively more Ferrar-derived detritus, which is a good source of fine-grained magnetite. Nevertheless, progressive fining of magnetite is observed throughout CRP-2/2A and into CRP-1. The progressive fining of magnetite through CRP-3 zones II and I and CRP-2/2A zones III and II (Figure 7) is likely to be due to fining within Ferrar-sourced material [*Verosub et al.*, 2000]. We attribute this fining to increased glacial grinding of particles that started with initiation of permanent Antarctic glaciation at the E-O boundary above CRP-3 zone III (see Figure 6 for context). There is a marked shift above the major unconformity in CRP-2/2A (Figure 6) from zone II to zone I at 270 mbsf in which detritus sourced from the McMurdo Volcanic Group becomes dominant [*Verosub et al.*, 2000]. There is generally good agreement between magnetite concentration variations in CRP-2/2A and basalt clast, volcanic glass, and pumice concentrations from the McMurdo Volcanic Group, which indicates that the environmental magnetic signal in CRP-2/2A provides a record of variations in supply of fine-grained volcanic particles from the McMurdo Volcanic Group [*Verosub et al.*, 2000]. Continuation of the fining-upward trend within CRP-2/2A zone I and CRP-1 indicates that magnetite sourced from the McMurdo Volcanic Group is finer than that derived from the Ferrar Group. Magnetic parameters that are sensitive to grain size demonstrate that there are detectable differences between magnetite populations from the Ferrar Dolerite and McMurdo volcanics [*Verosub et al.*, 2000].

[31] Overall, the upward fining of magnetite grain size in the CRP cores (Figures 7 and 8) probably reflects three processes. First, basal sediments in CRP-3 are coarse-grained and represent a major phase of erosion of the nearby Transantarctic Mountains as they were being actively uplifted. Second, progressive fining of Ferrar-sourced magnetite in the upper parts of CRP-3 and lower parts of CRP-2/2A probably reflect glacial grinding of magnetite into progressively finer particles. Third, the fining of magnetite observed in the upper part of CRP-2/2A and CRP-1 reflects dominance of a finer source of magnetite associated with volcanic glass inputs from the McMurdo Volcanic Group.

5. Conclusions

[32] Environmental magnetic results from sediments spanning a total stratigraphic thickness of 2.6 km record paleoclimatic, tectonic, provenance, and volcanic processes in the VLB. The studied ~17 Myr time interval spanned the period prior to and during buildup of continental-scale Antarctic ice sheets, the period during which the Transantarctic Mountains were uplifted along the Ross Sea rift shoulder, and the subsequent period when widespread

volcanism developed in McMurdo Sound. The magnetic properties of the studied VLB cores are dominated by large variations in magnetite concentration. During the late Eocene and early Oligocene, magnetite concentration variations coincided with detrital smectite concentration and crystallinity variations, which indicates a paleoclimatic control on the magnetic properties through influence on the weathering regime (high magnetite and smectite concentrations indicate warmer and wetter climates and vice versa). Through the early Oligocene, magnetic parameters clearly record progressive erosion of geological units on the rift margin as the Transantarctic Mountains were uplifted and unroofed during a major phase of activity on the Ross Sea rift. A consistent overall fining upward of magnetite particles is observed from the late Eocene to early Miocene. This is interpreted to reflect the progressively increasing importance of mechanical grinding of erosional detritus due to glacial processes associated with glacial buildup through this time interval. After 24 Myr, McMurdo Group volcanics dominate the magnetic properties of VLB sediments. The finer grain size of magnetite in volcanic glass from the McMurdo Group contributed to the fining-upward trend observed for magnetite throughout the CRP cores. Overall, long-term magnetic property variations in the VLB record a range of paleoclimatic, tectonic, provenance, and volcanic influences. Magnetic properties have effectively recorded all of the important geological processes that governed sedimentation over a long period of time (17 Ma) in the Victoria Land Basin, Antarctica.

[33] **Acknowledgments.** We gratefully acknowledge the U.S. National Science Foundation Office of Polar Programs, the Programma Nazionale di Ricerche in Antartide (PNRA), the New Zealand Foundation for Research (Contract UOOX0207), and the New Zealand Ministry of Science and Innovation (Contract C05X1001), respectively, for supporting this work. We also acknowledge our colleagues from the Cape Roberts Project (led by Peter Barrett and Peter Webb) for numerous discussions that helped to develop some of the ideas expressed here. A.P.R. is grateful to the PNRA for funding a stay in Rome that enabled preparation of much of this paper. We thank Ted Evans and Stefanie Brachfeld for constructive review comments and Tom Parsons for efficient editorial handling.

References

Banerjee, S. K., J. King, and J. Marvin (1981), A rapid method for magnetic granulometry with applications to environmental studies, *Geophys. Res. Lett.*, *4*, 333–336.
Barrett, P. J. (1981), History of the Ross Sea region during the deposition of the Beacon Supergroup 400–180 million years ago, *J. R. Soc. N. Z.*, *11*, 447–458.
Barrett, P. J. (Ed) (1986), *Antarctic Cenozoic History from the MSSTS-1 Drillhole, McMurdo Sound, Antarctica*, Dept. Sci. Ind. Res. Bull., 237 DSIR Publishing, Wellington, N.Z, 174 pp.
Barrett, P. J. (Ed) (1989), *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica*, Dept. Sci. Ind. Res. Bull., 245 DSIR Publishing, Wellington, N.Z, 254 pp.
Barrett, P. J., and B. C. McKelvey (1981), Permian tillites of South Victoria Land, Antarctica, in *Earth's Pre-Pleistocene Glacial Record*, edited by M. J. Hambrey, and W. B. Harland, pp. 233–236, Cambridge University Press.
Barrett, P. J., G. W. Grindley, and P. N. Webb (1972), The Beacon Supergroup of East Antarctica, in *Antarctic Geology and Geophysics*, edited by R. J. Adie, pp. 319–322, Universitetsforlaget, Oslo.
Barrett, P. J., S. A. Henrys, L. R. Bartek, G. Brancolini, M. Busetti, F. J. Davey, M. J. Hannah, and A. R. Pyne (1995), Geology off the margin of the Victoria Land Basin, Southwest Ross Sea, in *Geology and Seismic Stratigraphy of the Antarctic Margin*, edited by A. K. Cooper, P. F. Barker, and G. Brancolini, Am. Geophys. Un., Ant. Res. Ser., 68, pp. 183–207.
Behrendt, J. C., W. E. LeMasurier, A. K. Cooper, F. Tessensohn, A. Tréhu, and D. Damaske (1991), Geophysical studies of the West Antarctic rift system, *Tectonics*, *10*, 1257–1273.
Brachfeld, S. A., S. K. Banerjee, Y. Guyodo, and G. D. Acton (2002), A 13200 year history of century to millennial-scale paleoenvironmental change magnetically recorded in the Palmer Deep, western Antarctic Peninsula, *Earth Planet. Sci. Lett.*, *194*, 311–326.
Bücker, C. J., R. D. Jarrard, F. Niessen, and T. Wonik (2001), Statistical analysis of wireline logging data of the CRP-3 drillhole (Victoria Land Basin, Antarctica), *Terra Ant.*, *8*, 491–505.
Campbell, I. B., and G. G. C. Claridge (1987), Antarctica: Soils, weathering processes and environment, in *Developments in Soil Science*, vol. 16, p. 368, Elsevier, Amsterdam, The Netherlands.
Cande, S. C., and D. V. Kent (1995), Revised calibration of the geomagnetic polarity time scale for the late Cretaceous and Cenozoic, *J. Geophys. Res.*, *100*, 6093–6095.
Cande, S. C., J. M. Stock, R. D. Müller, and T. Ishihara (2000), Cenozoic motion between East and West Antarctica, *Nature*, *404*, 145–150.
Canfield, D. E., and R. A. Berner (1987), Dissolution and pyritization of magnetite in anoxic marine sediments, *Geochim. Cosmochim. Acta*, *51*, 645–659.
Channell, J. E. T., and T. Hawthorne (1990), Progressive dissolution of titanomagnetites at ODP site 653 (Tyrrhenian Sea), *Earth Planet. Sci. Lett.*, *96*, 469–480.
Cooper, A. K., F. J. Davey, and J. C. Behrendt (1987), Seismic stratigraphy and structure of the Victoria Land Basin, Western Ross Sea, Antarctica, in *The Antarctic Continental Margin: Geology and Geophysics of the Western Ross Sea*, edited by A. K. Cooper, and F. J. Davey, Earth Science Series, vol. 5B, pp. 27–76, Circum-Pacific Research Council, Houston, Texas.
Coxall, H. K., P. A. Wilson, H. Pälike, C. H. Lear, and J. Backman (2005), Rapid stepwise onset of Antarctic glaciation and deeper calcite compensation in the Pacific Ocean, *Nature*, *433*, 53–57.
Davey, F. J., D. J. Bennett, and R. E. Hout (1982), Sedimentary basins of the Ross Sea, Antarctica, *N. Z. J. Geol. Geophys.*, *25*, 245–255.
Davey, F. J., K. Hinz, and H. Schroeder (1983), Sedimentary basins of the Ross Sea, Antarctica, in *Antarctic Earth Science*, edited by R. L. Oliver, P. R. James, and J. B. Jago, pp. 533–588, Australian Academy of Science, Canberra.
Day, R., M. Fuller, and V. A. Schmidt (1977), Hysteresis properties of titanomagnetites: Grain size and composition dependence, *Phys. Earth Planet. Inter.*, *13*, 305–314.
Deng, C., R. X. Zhu, M. J. Jackson, K. L. Verosub, and M. J. Singer (2001), Variability of the temperature-dependent susceptibility of the Holocene eolian deposits in the Chinese loess plateau: A pedogenesis indicator, *Phys. Chem. Earth*, *26*, 873–878.
Diester-Haass, L. (1996), Late Eocene-Oligocene paleoceanography in the southern Indian Ocean (ODP Site 744), *Mar. Geol.*, *130*, 99–119.
Dunlop, D. J. (2002), Theory and application of the Day plot (M_{rs}/M_s versus H_{cr}/H_c) 1. Theoretical curves and tests using titanomagnetite data, *J. Geophys. Res.*, *107*(B3), 2056, doi:10.1029/2001JB000486.
Ehrmann, W. (1998), Implications of late Eocene to early Miocene clay mineral assemblages in McMurdo Sound (Ross Sea, Antarctica) on paleoclimate and ice dynamics, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *139*, 213–231.
Ehrmann, W. (2000), Variations in smectite content and crystallinity in sediments from CRP-2/2A, Victoria Land Basin, Antarctica, *Terra Ant.*, *7*, 575–580.
Ehrmann, W. (2001), Smectite content and crystallinity in sediments from CRP-3, Victoria Land Basin, Antarctica, *Terra Ant.*, *8*, 533–542.
Ehrmann, W. U., and A. Mackensen (1992), Sedimentological evidence for the formation of an East Antarctic ice sheet in Eocene/Oligocene time, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *93*, 85–112.
Ehrmann, W., M. Setti, and L. Marinoni (2005), Clay minerals in Cenozoic sediments off Cape Roberts (McMurdo Sound, Antarctica) reveal palaeoclimatic history, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *229*, 187–211.
Elliot, D. H., and T. H. Fleming (2000), Weddell triple junction: The principal focus of Ferrar and Karoo magmatism during initial breakup of Gondwana, *Geology*, *28*, 539–542.
Elliot, D. H., T. H. Fleming, P. R. Kyle, and K. A. Foland (1999), Long-distance transport of magmas in the Jurassic Ferrar Large Igneous Province, Antarctica, *Earth Planet. Sci. Lett.*, *167*, 89–104.
Encarnación, J., T. H. Fleming, D. H. Elliot, and H. V. Eales (1996), Synchronous emplacement of Ferrar and Karoo dolerites and the early breakup of Gondwana, *Geology*, *24*, 535–538.
Evans, M. E., and F. Heller (2003), *Environmental Magnetism: Principles and Applications of Enviromagnetics*, p. 299, Academic Press, San Diego, Calif.
Fielding, C. R., T. R. Naish, K. J. Woolfe, and M. A. Lavelle (2000), Facies analysis and sequence stratigraphy of CRP-2/2A, Victoria Land Basin, Antarctica, *Terra Ant.*, *7*, 323–338.
Fielding, C. R., J. Whittaker, S. A. Henrys, T. J. Wilson, and T. R. Naish (2008), Seismic facies and stratigraphy of the Cenozoic succession in McMurdo Sound, Antarctica: Implications for tectonic, climatic and glacial history, *Paleogeogr. Palaeoclimatol. Palaeoecol.*, *260*, 8–29.

Fitzgerald, P. G. (1992), The Transantarctic Mountains of southern Victoria Land: The application of apatite fission track analysis to a rift shoulder uplift, *Tectonics*, *11*, 634–662.

Fitzgerald, P. G., M. Sandiford, P. J. Barrett, and A. J. W. Gleadow (1986), Asymmetric extension associated with uplift and subsidence in the Transantarctic Mountains and Ross embayment, *Earth Planet. Sci. Lett.*, *81*, 67–78.

Florindo, F., G. S. Wilson, A. P. Roberts, L. Sagnotti, and K. L. Verosub (2001), Magnetostratigraphy of late Eocene-early Oligocene glaciomarine sediments from the CRP-3 Core, McMurdo Sound, Ross Sea, Antarctica, *Terra Ant.*, *8*, 599–613.

Florindo, F., G. S. Wilson, A. P. Roberts, L. Sagnotti, and K. L. Verosub (2005), Magnetostratigraphic chronology of a Late Eocene to Early Miocene glacimarine succession from the Victoria Land Basin, Ross Sea, Antarctica, *Global Planet. Change*, *45*, 207–236.

Galeotti, S., L. Lanci, F. Florindo, T. R. Naish, L. Sagnotti, S. Sandroni, and F. M. Talarico (2012), Cyclochronology of the Eocene-Oligocene transition from the Cape Roberts Project-3 core, Victoria Land basin, Antarctica, *Paleogeogr. Palaeoclimatol. Palaeoecol.*, *335*, 84–94.

Glasby, G. P., J. G. McPherson, B. P. Kohn, J. H. Johnston, J. R. Keys, A. G. Freeman, and M. J. Tricker (1981), Desert varnish in southern Victoria Land, Antarctica, *N. Z. J. Geol. Geophys.*, *24*, 389–397.

Grapes, R. H., D. L. Reid, and J. G. McPherson (1974), Shallow dolerite intrusion and phreatic eruption in the Allan Hills region, Antarctica, *N. Z. J. Geol. Geophys.*, *17*, 563–577.

Grindley, G. W. (1963), The geology of the Queen Alexandra Range, Beardmore Glacier, Ross Dependency, Antarctica, with notes on the correlation of Gondwana sequences, *N. Z. J. Geol. Geophys.*, *6*, 307–347.

Grindley, G. W., and G. Warren (1964), Stratigraphic nomenclature and correlation in the western Ross Sea region, in *Antarctic Geology*, edited by R. J. Adie, pp. 314–333, North Holland Publishing, Amsterdam, The Netherlands.

Gunn, B. M., and G. Warren (1962), Geology of Victoria Land between Mawson and Mullock Glaciers, Antarctica, *N. Z. Geol. Surv. Bull.*, *71*.

Hamilton, R. J., B. P. Luyendyk, C. C. Sorlien, and L. A. Bartek (2001), Cenozoic tectonics of the Cape Roberts Rift Basin and Transantarctic Mountain Front, southwestern Ross Sea, Antarctica, *Tectonics*, *20*, 325–342.

Hannah, M. J. (1997), Climate controlled dinoflagellate distribution in Late Eocene–Earliest Oligocene strata from CIROS-1 drillhole, McMurdo Sound, Antarctica, *Terra Ant.*, *4*, 73–78.

Hannah, M. J., M. B. Cita, R. Coccioni, and S. Monechi (1997), The Eocene/Oligocene boundary at 70°S, McMurdo Sound, Antarctica, *Terra Ant.*, *4*, 79–87.

Hannah, M. J., J. H. Wrenn, and G. J. Wilson (2001), Preliminary report on Early Oligocene and ?Latest Eocene marine palynomorphs from CRP-3 drillhole, Victoria Land Basin, Antarctica, *Terra Ant.*, *8*, 383–388.

Harrington, H. J. (1958), Nomenclature of rock units in the Ross Sea region, Antarctica, *Nature*, *182*, 290.

Harwood, D. M. (1986), Diatoms, in *Antarctic Cenozoic History from the MSSTS-1 Drillhole, McMurdo Sound, Antarctica*, edited by P. J. Barrett, Dept Sci. Ind. Res. Bull., 237, pp. 69–107, DSIR Publishing, Wellington, N.Z.

Harwood, D. M. (1989), Siliceous microfossils, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica*, edited by P. J. Barrett, Dept Sci. Ind. Res. Bull., 245, 67–97, DSIR Publishing, Wellington, N.Z.

Harwood, D. M., and S. M. Bohaty (2001), Early Oligocene siliceous microfossil biostratigraphy of Cape Roberts Project core CRP-3, Victoria Land Basin, Antarctica, *Terra Ant.*, *8*, 315–338.

Harwood, D. M., S. M. Bohaty, and R. P. Scherer (1998), Lower Miocene diatom biostratigraphy of the CRP-1 drillcore, McMurdo Sound, Antarctica, *Terra Ant.*, *5*, 499–514.

Heimann, A., T. H. Fleming, D. H. Elliot, and K. A. Foland (1994), A short interval of Jurassic continental flood basalt volcanism in Antarctica as demonstrated by $^{40}Ar/^{39}Ar$ geochronology, *Earth Planet. Sci. Lett.*, *121*, 19–41.

Henrys, S. A., C. J. Bücker, F. Niessen, and L. R. Bartek (2001), Correlation of seismic reflectors with the CRP-3 drillhole, Victoria Land Basin, Antarctica, *Terra Ant.*, *8*, 127–136.

Hill, R. S. (1989), Fossil leaf, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica*, edited by P. J. Barrett, Dept Sci. Ind. Res. Bull., 245, pp. 143–144, DSIR Publishing, Wellington, N.Z.

Horng, C. S., and A. P. Roberts (2006), Authigenic or detrital origin of pyrrhotite in sediments?: Resolving a paleomagnetic conundrum, *Earth Planet. Sci. Lett.*, *241*, 750–762.

Horng, C. S., C. A. Huh, K. H. Chen, C. H. Lin, K. S. Shea, and K. H. Hsiung (2012), Pyrrhotite as a tracer for denudation of the Taiwan orogen, *Geochem. Geophys. Geosyst.*, *13*, Q08Z47, doi:10.1029/2012GC004195.

Kanfoush, S. L., D. A. Hodell, C. D. Charles, T. R. Janecek, and F. R. Rack (2002), Comparison of ice-rafted debris and physical properties in ODP Site 1094 (South Atlantic) with the Vostok ice core over the last four climatic cycles, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *182*, 329–349.

Kao, S. J., C. S. Horng, A. P. Roberts, and K. K. Liu (2004), Carbon-sulfur-iron relationships in sedimentary rocks from southwestern Taiwan: Influence of geochemical environment on greigite and pyrrhotite formation, *Chem. Geol.*, *203*, 153–168.

Karlin, R. (1990a), Magnetite diagenesis in marine sediments from the Oregon continental margin, *J. Geophys. Res.*, *95*, 4405–4419.

Karlin, R. (1990b), Magnetic mineral diagenesis in suboxic sediments at Bettis Site W-N, NE Pacific Ocean, *J. Geophys. Res.*, *95*, 4421–4436.

Karlin, R., and S. Levi (1983), Diagenesis of magnetic minerals in recent haemipelagic sediments, *Nature*, *303*, 327–330.

Karlin, R., and S. Levi (1985), Geochemical and sedimentological control of the magnetic properties of hemipelagic sediments, *J. Geophys. Res.*, *90*, 10,373–10,392.

Kelly, W. C., and J. H. Zumberge (1961), Weathering of a quartz diorite at Marble Point, McMurdo Sound, Antarctica, *J. Geol.*, *69*, 433–446.

King, J. W., S. K. Banerjee, and J. Marvin (1983), A new rock-magnetic approach to selecting sediments for geomagnetic paleointensity studies: Application to paleointensity for the last 4000 years, *J. Geophys. Res.*, *88*, 5911–5921.

Korsch, R. J. (1974), Petrographic comparison of the Taylor and Victoria Groups (Devonian to Triassic) in South Victoria Land, Antarctica, *N. Z. J. Geol. Geophys.*, *17*, 523–541.

Kyle, P. R. (1990), McMurdo Volcanic Group-Western Ross Embayment: Introduction, *Am. Geophys. Un. Ant. Res. Ser.*, *48*, 19–25.

Lavelle, M. (1998), Strontium-isotope stratigraphy of the CRP-1 drill-hole, Ross Sea, Antarctica, *Terra Ant.*, *5*, 691–696.

Lavelle, M. (2000), Strontium-isotope stratigraphy and age model for CRP-2/2A, Victoria Land Basin, Antarctica, *Terra Ant.*, *7*, 611–620.

Lavelle, M. (2001), Strontium-isotope stratigraphy for CRP-3, Victoria Land Basin, Antarctica, *Terra Ant.*, *8*, 593–598.

Leeder, M. (2011), *Sedimentology and Sedimentary Basins: From Turbulence to Tectonics*, 2nd ed., p. 592, Wiley-Blackwell, Chichester, U.K.

Leslie, B. W., S. P. Lund, and D. E. Hammond (1990), Rock magnetic evidence for the dissolution and authigenic growth of magnetic minerals within anoxic sediments of the California continental borderland, *J. Geophys. Res.*, *95*, 4437–4452.

Liu, Q. S., C. L. Deng, Y. J. Yu, J. Torrent, M. J. Jackson, S. K. Banerjee, and R. X. Zhu (2005), Temperature dependence of magnetic susceptibility in an argon environment: Implications for pedogenesis of Chinese loess/paleosols, *Geophys. J. Int.*, *161*, 102–112.

Liu, Q. S., A. P. Roberts, J. C. Larrasoaña, S. K. Banerjee, Y. Guyodo, L. Tauxe, and F. Oldfield (2012), Environmental magnetism: Principles and applications, *Rev. Geophys.*, RG4002, doi:10.1029/2012RG000393.

Lowrie, W. (1990), Identification of ferromagnetic minerals in a rock by coercivity and unblocking temperature properties, *Geophys. Res. Lett.*, *17*, 159–162.

Maher, B. A., and R. Thompson (Eds) (1999), *Quaternary Climates, Environments and Magnetism*, p. 402, Cambridge Univ. Press, New York.

Marchant, D. R., C. C. Swisher, D. R. Lux, D. P. West, and G. H. Denton (1993), Pliocene paleoclimate and East Antarctic ice-sheet history from surficial ash deposits, *Science*, *260*, 667–670.

McArthur, J. M., R. J. Howarth, and T. R. Bailey (2001), Strontium isotope stratigraphy: LOWESS version 3: Best fit to the marine Sr-isotope curve for 0–509 Ma and accompanying look-up table for deriving numerical age, *J. Geol.*, *109*, 155–170.

McIntosh, W. C. (2000), $^{40}Ar/^{39}Ar$ geochronology of tephra and volcanic clasts in CRP-2A, Victoria Land Basin, Antarctica, *Terra Ant.*, *7*, 621–630.

McKelvey, B. C., P.-N. Webb, M. P. Gorton, and B. P. Kohn (1970), Stratigraphy of the Beacon Supergroup between the Olympus and Boomerang Ranges, Victoria Land, Antarctica, *Nature*, *227*, 1126–1128.

Mildenhall, D. C. (1989), Terrestrial palynology, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica*, edited by P. J. Barrett, Dept Sci. Ind. Res. Bull., 245, pp. 119–127, DSIR Publishing, Wellington, N.Z.

Miller, K. G., R. G. Fairbanks, and G. S. Mountain (1987), Tertiary oxygen isotope synthesis, sea level history, and continental margin erosion, *Paleoceanography*, *2*, 1–19.

Naish, T. R. et al. (2001), Orbitally induced oscillations in the East Antarctic ice sheet at the Oligocene/Miocene boundary, *Nature*, *413*, 719–723.

Olney, M. P., R. P. Scherer, D. M. Harwood, and S. M. Bohaty (2007), Oligocene-early Miocene Antarctic nearshore diatom biostratigraphy, *Deep-Sea Res., II*, *54*, 2325–2349.

Pirrung, M., C. D. Hillenbrand, B. Diekmann, D. Fütterer, H. Grobe, and G. Kuhn (2002), Magnetic susceptibility and ice-rafted debris in surface sediments of the Atlantic sector of the Southern Ocean, *Geo. Mar. Lett.*, *22*, 170–180.

Pross, J. et al. (2012), Persistent near-tropical warmth on the Antarctic continent during the early Eocene epoch, *Nature*, *488*, 73–77.

Prothero, D. R. (1994), The Eocene–Oligocene extinctions, *Annu. Rev. Earth Planet. Sci.*, *22*, 145–165.
Raine, J. I. (1998), Terrestrial palynomorphs from Cape Roberts Project drillhole CRP-1, Ross Sea, Antarctica, *Terra Ant.*, *5*, 539–548.
Raine, J. I., and R. A. Askin (2001), Terrestrial palynology of Cape Roberts Project drillhole CRP-3, Victoria Land Basin, Antarctica, *Terra Ant.*, *8*, 389–400.
Robert, C., and J. P. Kennett (1997), Antarctic continental weathering changes during Eocene-Oligocene cryosphere expansion: Clay mineral and oxygen isotope evidence, *Geology*, *25*, 587–590.
Robert, C., L. Diester-Haass, and H. Chamley (2002), Late Eocene-Oligocene oceanographic development at southern high latitudes, from terrigenous and biogenic particles: A comparison of Kerguelen Plateau and Maud Rise, ODP Sites 744 and 689, *Mar. Geol.*, *191*, 37–54.
Roberts, A. P., and R. Weaver (2005), Multiple mechanisms of remagnetization involving sedimentary greigite (Fe_3S_4), *Earth Planet. Sci. Lett.*, *231*, 263–277.
Roberts, A. P., Y. L. Cui, and K. L. Verosub (1995), Wasp-waisted hysteresis loops: Mineral magnetic characteristics and discrimination of components in mixed magnetic systems, *J. Geophys. Res.*, *100*, 17,909–17,924.
Roberts, A. P., G. S. Wilson, F. Florindo, L. Sagnotti, K. L. Verosub, and D. M. Harwood (1998), Magnetostratigraphy of lower Miocene strata from the CRP-1 core, McMurdo Sound, Ross Sea, Antarctica, *Terra Ant.*, *5*, 703–713.
Roberts, A. P., G. S. Wilson, D. M. Harwood, and K. L. Verosub (2003a), Glaciation across the Oligocene-Miocene boundary in southern McMurdo Sound, Antarctica: New chronology from the CIROS-1 drill hole, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *198*, 113–130.
Roberts, A. P., S. Bicknell, J. Byatt, S. M. Bohaty, F. Florindo, and D. M. Harwood (2003b), Magnetostratigraphic calibration of Southern Ocean diatom biostratigraphic datums from the Eocene-Oligocene of Kerguelen Plateau (ODP sites 744 and 748), *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *198*, 145–168.
Roberts, A. P., L. Chang, D. Heslop, F. Florindo, and J. C. Larrasoaña (2012), Searching for single domain magnetite in the "pseudo-single-domain" sedimentary haystack: Implications of biogenic magnetite preservation for sediment magnetism and relative paleointensity determinations, *J. Geophys. Res.*, *117*, B08104, doi:10.1029/2012JB009412.
Rochette, P. (1987), Metamorphic control of the magnetic mineralogy of black shales in the Swiss Alps: Toward the use of "magnetic isogrades", *Earth Planet. Sci. Lett.*, *84*, 446–456.
Rowan, C. J., A. P. Roberts, and T. Broadbent (2009), Reductive diagenesis, magnetic dissolution, greigite growth, and paleomagnetic smoothing in marine sediments: A new view, *Earth Planet. Sci. Lett.*, *277*, 223–235.
Sagnotti, L., F. Florindo, K. L. Verosub, G. S. Wilson, and A. P. Roberts (1998a), Environmental magnetic record of Antarctic palaeoclimate from Eocene-Oligocene glaciomarine sediments, Victoria Land margin, *Geophys. J. Int.*, *134*, 653–662.
Sagnotti, L., F. Florindo, G. S. Wilson, A. P. Roberts, and K. L. Verosub (1998b), Environmental magnetism of lower Miocene strata from the CRP-1 core, McMurdo Sound, Antarctica, *Terra Ant.*, *5*, 661–667.
Sagnotti, L., K. L. Verosub, A. P. Roberts, F. Florindo, and G. S. Wilson (2001a), Environmental magnetic record of the Eocene-Oligocene transition in the CRP-3 drillcore, Victoria Land Basin, Antarctica, *Terra Ant.*, *8*, 507–516.
Sagnotti, L., P. Macrí, A. Camerlenghi, and M. Rebesco (2001b), Environmental magnetism of Antarctic Late Pleistocene sediments and interhemispheric correlation of climatic events, *Earth Planet. Sci. Lett.*, *192*, 65–80.
Sagnotti, L., A. P. Roberts, R. Weaver, K. L. Verosub, F. Florindo, G. S. Wilson, C. R. Pike, and T. Clayton (2005), Apparent high-frequency magnetic polarity reversals due to alternating remagnetizations resulting from late diagenetic growth of greigite from siderite, *Geophys. J. Int.*, *160*, 89–100.
Schäfer, J. M., S. Ivy-Ochs, R. Wieler, I. Leya, H. Baur, G. H. Denton, and C. Schlüchter (1999), Cosmogenic noble gas studied in the oldest landscape on earth: Surface exposure ages of the Dry Valleys, Antarctica, *Earth Planet. Sci. Lett.*, *167*, 215–226.
Scherer, R. P., S. M. Bohaty, and D. M. Harwood (2000), Oligocene and lower Miocene siliceous microfossil biostratigraphy of Cape Roberts Project Core CRP-2/2A, Victoria Land Basin, Antarctica, *Terra Ant.*, *7*, 417–442.
Smellie, J. L. (2000), Erosional history of the Transantarctic Mountains deduced from sand grain detrital modes in CRP-2/2A, Victoria Land Basin, Antarctica, *Terra Ant.*, *7*, 545–552.
Smellie, J. L. (2001), History of Oligocene erosion, uplift and unroofing of the Transantarctic Mountains deduced from sandstone detrital modes in CRP-3 drillcore, Victoria Land Basin, Antarctica, *Terra Ant.*, *8*, 481–489.
Stern, T. A., and U. S. ten Brink (1989), Flexural uplift of the Transantarctic Mountains, *J. Geophys. Res.*, *94*, 10,315–10,330.
Stump, E. (1995), *The Ross Orogen of the Transantarctic Mountains*, Cambridge Univ. Press, Cambridge, pp. 284 .
Sugden, D. E., G. H. Denton, and D. R. Marchant (1995), Landscape evolution of the Dry Valleys, Transantarctic Mountains: Tectonic implications, *J. Geophys. Res.*, *100*, 9949–9967.
Sugden, D. E., M. A. Summerfield, G. H. Denton, T. I. Wilch, W. C. McIntosh, D. R. Marchant, and R. H. Rutford (1999), Landscape development in the Royal Society Range, southern Victoria Land, Antarctica: Stability since the mid-Miocene, *Geomorphology*, *28*, 181–200.
Talarico, F., S. Sandroni, C. Fielding, and C. Atkins (2000), Variability, petrography and provenance of basement clasts in core from CRP-2/2A, Victoria Land basin, Antarctica, *Terra Ant.*, *7*, 529–544.
Thompson, R., and F. Oldfield (1986), *Environmental Magnetism*, p. 227, Allen and Unwin, Winchester, Mass.
Venuti, A., F. Florindo, A. Carburlotto, M. W. Hounslow, C.-D. Hillenbrand, E. Strada, F. M. Talarico, and A. Cavallo (2011), Late Quaternary sediments from deep-sea sediment drifts on the Antarctic Peninsula Pacific margin: Climatic control on the provenance of minerals, *J. Geophys. Res.*, *116*, B06104, doi:10.1029/2010JB007952.
Verosub, K. L., and A. P. Roberts (1995), Environmental magnetism: Past, present, and future, *J. Geophys. Res.*, *100*, 2175–2192.
Verosub, K. L., F. Florindo, L. Sagnotti, A. P. Roberts, and G. S. Wilson (2000), Environmental magnetism of Oligocene–Miocene glaciomarine strata from the CRP-2/2A drillcore, Victoria Land Basin, Antarctica, *Terra Ant.*, *7*, 599–608.
Watkins, D. K. (2007), Quantitative analysis of the calcareous nannofossil assemblages from CIROS-1, Victoria Land Basin, Antarctica, *J. Nannoplankton Res.*, *29*, 130–137.
Watkins, D. K., and G. Villa (2000), Palaeogene calcareous nannofossils from CRP-2/2A, Victoria Land Basin, Antarctica, *Terra Ant.*, *7*, 443–452.
Webb, P.-N., and D. M. Harwood (1991), Late Cenozoic glacial history of the Ross embayment, Antarctica, *Quat. Sci. Rev.*, *10*, 215–223.
Wilson, G. S., A. P. Roberts, K. L. Verosub, F. Florindo, and L. Sagnotti (1998), Magnetobiostratigraphic chronology of the Eocene-Oligocene transition in the CIROS-1 core, Victoria Land margin, Antarctica: Implications for Antarctic glacial history, *Geol. Soc. Am. Bull.*, *110*, 35–47.
Wilson, G. S. et al. (2000a), Chronostratigraphy of the CRP-2/2A drillhole, Ross Sea, Antarctica, *Terra Ant.*, *7*, 647–655.
Wilson, G. S., F. Florindo, L. Sagnotti, K. L. Verosub, and A. P. Roberts (2000b), Magnetostratigraphy of Oligocene–Miocene glaciomarine strata from the CRP-2/2A core, Victoria Land Basin, Antarctica, *Terra Ant.*, *7*, 631–646.
Wilson, G. S. et al. (2002), Integrated chronostratigraphic calibration of the Oligocene-Miocene boundary at 24.0 ± 0.1 Ma from the CRP-2A drill core, Ross Sea, Antarctica, *Geology*, *30*, 1043–1046.
Zachos, J., M. Pagani, L. Sloan, E. Thomas, and K. Billups (2001), Trends, rhythms, and aberrations in global climate 65 Ma to present, *Science*, *292*, 686–693.