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Environmental magnetic record of Antarctic palaeoclimate from Eocene/Oligocene glaciomarine sediments, Victoria Land Basin

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SUMMARY

The onset of continent-wide glaciation in Antarctica is still poorly understood, despite being one of the most important palaeoclimatic events in the Cenozoic. The Eocene/Oligocene boundary interval has recently been recognized as a critical time for Antarctic climatic evolution, and it may mark the preglacial–glacial transition. Magnetic susceptibility, intensity of natural and artificial remanences, hysteresis parameters and magnetic anisotropy of the lower half (late Eocene/early Oligocene) of the CIROS-1 core (from McMurdo Sound, Antarctica) reveal alternating intervals of high and low magnetic mineral concentrations that do not correspond to lithostratigraphic units in the core. Pseudo-single-domain magnetite is the main magnetic mineral throughout the sequence, and sharp changes in magnetite concentration match changes in clay mineralogy beneath and at the Eocene/Oligocene boundary. The detrital magnetite originated from weathering of the Ferrar Group (which comprises basic extrusive and intrusive igneous rocks). Weathering processes and input of magnetite to the Victoria Land Basin were intense during periods when the Antarctic climate was warmer than today, but during intervals when the climate was relatively cool, chemical weathering of the Ferrar Group was suppressed and input of detrital magnetite to the Victoria Land Basin decreased. Our results also indicate that a cold and dry climate was not established in Antarctica until the Eocene/Oligocene boundary, with major ice sheet growth occurring at the early/late Oligocene boundary. Some earlier cold intervals are identified, which indicate that climate had begun to deteriorate by the middle/late Eocene boundary.

Key words: Cenozoic, palaeoclimate, palaeomagnetism, rock magnetism, sediments.

INTRODUCTION

An understanding of the glacial history of Antarctica is crucial to any reconstruction of global climatic changes during the Cenozoic; however, the onset of continent-wide glaciation in Antarctica is still poorly understood. Kennett (1977) suggested that the inception of Antarctic glaciation coincided with major palaeoceanographic modifications in the Southern Ocean that occurred in association with a plate tectonic redistribution of land masses at high southern latitudes. This redistribution is a direct expression of the break-up of Gondwana (Kennett 1977; Lawver, Gahagan & Coffin 1992). Integration of results from ocean drilling around Antarctica since the early 1970s has established the antiquity of glaciation in Antarctica. However, even though several independent data sets suggest that the onset of widespread glaciation over the East Antarctic craton occurred close to the Eocene/Oligocene boundary (e.g. Wei 1991; Ehrmann & Mackensen 1992; Hambrey & Barrett 1993; Diester-Haass, Robert & Chamley 1996), the character and extent of the Antarctic continental ice cap since the early Oligocene are still the subject of controversy (see Wise *et al.* 1991; van der Wateren & Verbers 1993; Bartek *et al.* 1996). Debate about the onset of glaciation has gained added significance because the middle Eocene to early Oligocene interval is now recognized as a critical period in Earth history, which was characterized by major changes in global climate and by significant turnovers in terrestrial and marine biota (Berggren & Prothero 1992).

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Cenozoic sedimentary sequences from Antarctic shelf basins have great potential for elucidating the history of Antarctic glaciation. The CIROS-1 drill-hole, which penetrated late Eocene to early Miocene strata of the southern Victoria Land Basin (VLB) beneath McMurdo Sound (Barrett 1989), represents the most complete sedimentary sequence obtained to date from the Antarctic continental margin.

Environmental magnetism (that is, the study of the concentration, grain size and mineralogy of the magnetic particles in a sedimentary sequence) has recently emerged as a discipline that can provide semi-quantitative proxies for palaeoenvironmental changes, with a speed, sensitivity and cost-effectiveness that most other techniques cannot match (Reynolds & King 1995; Verosub & Roberts 1995). To date, however, mineral magnetic and environmental magnetic techniques have not been extensively applied to Antarctic marine sediments. In this paper, we report on mineral magnetic properties of the CIROS-1 core and discuss their bearing on the reconstruction of regional and continent-wide environmental changes in the Ross Sea during late Eocene/early Oligocene time.

THE CIROS-1 CORE

The CIROS-1 core was drilled from a sea ice platform during October and November 1986, in 197 m of water, 12 km offshore from the western margin of McMurdo Sound, southwestern Ross Sea (Fig. 1). The aim of the CIROS project was to study the Cenozoic tectonic and glacial history of the region, as recorded at the western edge of the VLB (see Barrett *et al.* 1989). The drill-hole was continuously cored from 26 to 702 m below seafloor (mbsf), with a core recovery of 98 per cent. The archive half of the CIROS-1 core is currently stored at the Antarctic Marine Geology Research Facility at Florida State University, Tallahassee, and has been maintained at a constant temperature of 2 °C. The working half of the core is stored at the Institute of Geological and Nuclear Sciences, Lower Hutt, New Zealand.

The sedimentary sequence comprises alternating diamictites, sandstones, mudstones and conglomerates (Fig. 2): a glacial influence is interpreted to be present to varying degrees throughout the core (Hambrey, Barrett & Robinson 1989). The core was divided into 22 lithological units (Robinson *et al.* 1987; Hambrey *et al.* 1989), which can be grouped into two main sequences, separated by a major disconformity (Fig. 2).

The upper sequence (from the seafloor to 366 mbsf; units 1–17) was deposited under the influence of glacial activity; it comprises mostly massive to weakly stratified diamictites (water-lain and lodgement tills) and mudstones with isolated pebbles, which were deposited in a shallow-water environment (Hambrey *et al.* 1989). At least three major episodes of glacial advance and retreat are recognized in the upper sequence of the CIROS-1 core (Barrett *et al.* 1989).

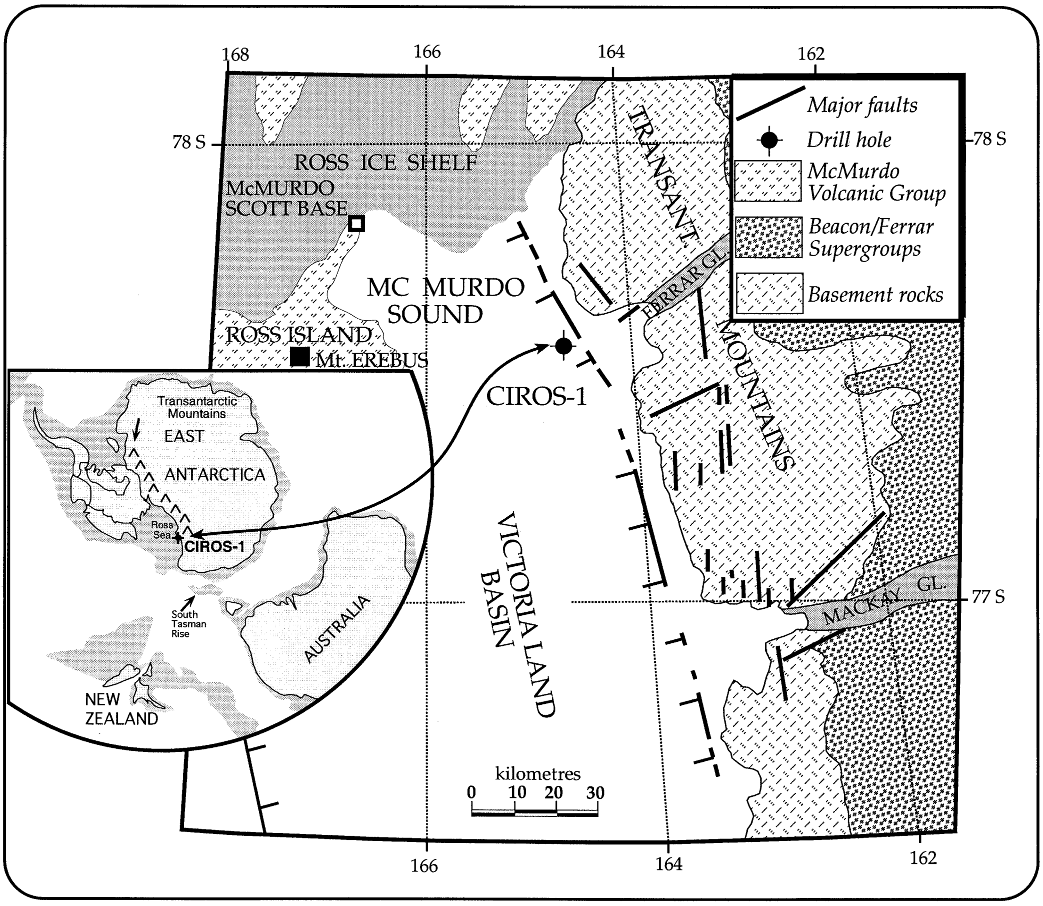

Figure 1. Map of the Ross Sea sector of Antarctica, with location of the CIROS-1 drill site. A palaeogeographic reconstruction of southern Gondwana is shown for the late Eocene (after Lawver *et al.* 1992) around the time of deposition of the basal sediments in the CIROS-1 core; shaded areas indicate continental margins.

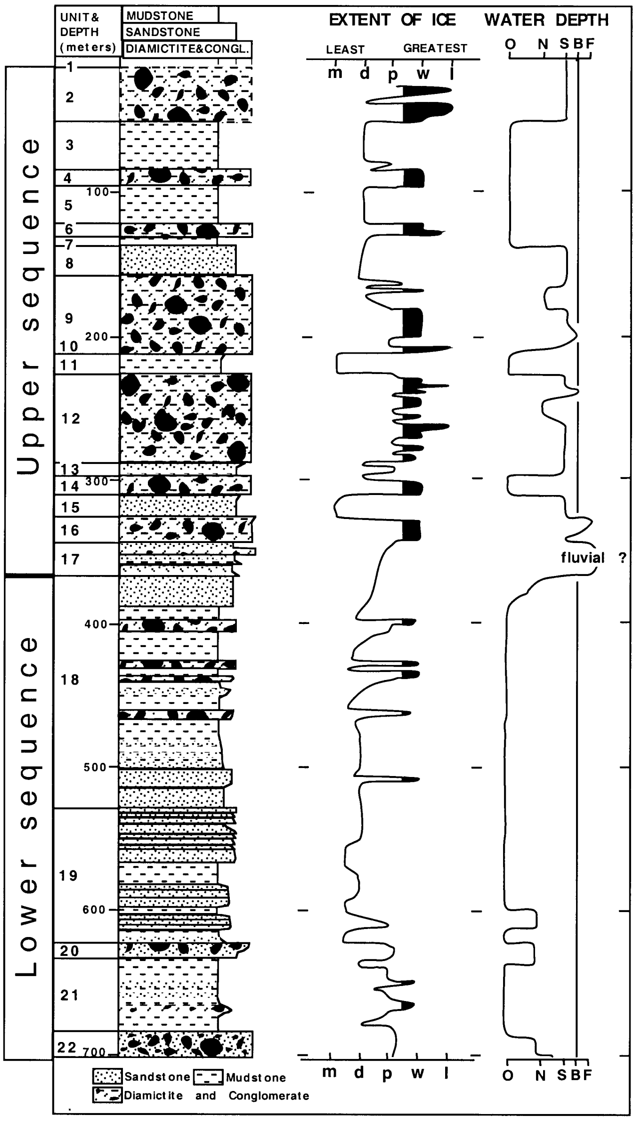

Figure 2. Summary lithological log of the CIROS-1 core with interpretation of extent of ice and water depth at the drill site. Ice front position: m = marine, d = distal, p = proximal, w = water-lain till, l = lodgement till. Water depth: O = offshore, N = nearshore, S = shoreface, B = beach, F = fluvial (after Hambrey *et al.* 1989).

The lower sequence (366–702 mbsf; units 18–22) mainly comprises bioturbated mudstones and sandstones that were deposited in an offshore pro-delta environment, with occasional glacial influence represented by water-lain diamictite horizons. Glacially derived dropstones, which are scattered through the lower sequence, indicate intermittent, but regular, deposition from floating ice (Hambrey *et al.* 1989). The mudstone intervals represent background sedimentation in a deep-water offshore setting. Numerous thin (0.1–1 m) sandstone beds occur between ≈530 and 620 mbsf (Unit 19) and were probably deposited by sediment gravity flows near the foot of the delta slope. Hambrey *et al.* (1989) interpreted two conglomerate intervals towards the base of the sequence (623–634 mbsf, Unit 20, and 686–702 mbsf, Unit 22) to represent debris flow deposits.

The CIROS-1 core was originally inferred to range in age from early Oligocene (or possibly latest Eocene) through to early Miocene (Harwood *et al.* 1989). However, the age of the lower part of the core was recently revised, and was attributed first to the middle Eocene (Hannah 1994) and, more recently, to the late Eocene (Wilson *et al.* 1998).

METHODS

Magnetostratigraphic analysis of the CIROS-1 core was originally performed by Rieck (1989); this work was limited to the upper 382.7 mbsf of the core, and the samples were not fully demagnetized. In the summer of 1995, we resampled the lower sequence in order to extend the magnetostratigraphic analysis to the base of the core and to analyse the rock magnetic properties, which have not been studied before. Cylindrical cores (25 mm diameter) were collected at an average stratigraphic interval of ≈2 m from 366 to 702 mbsf, wherever the sediment texture was suitable for palaeomagnetic and rock magnetic studies. Magnetostratigraphic results from the CIROS-1 core, and their implications for Antarctic glacial history, are discussed elsewhere (Wilson *et al.* 1998).

Rock magnetic analyses were performed in order to obtain a quantitative estimate of the downcore variation in the composition, concentration and grain size of magnetic minerals, and to determine the magnetic fabric of the sediments. Measurements were made at the University of California, Davis, and at the Istituto Nazionale di Geofisica, Rome, and included determination of:

(1) low-field magnetic susceptibility k;
(2) natural remanent magnetization (NRM);
(3) hysteresis properties, namely saturation magnetization M_s (up to peak fields of 1 T), saturation remanence M_r, coercive force H_c and coercivity of remanence H_{cr};
(4) stepwise acquisition and demagnetization of artificial remanences, namely anhysteretic remanent magnetization (ARM) imparted in a 100 mT alternating field, with a superimposed 0.1 mT bias field, and isothermal remanent magnetization (IRM) imparted in DC fields of up to 1.6 T;
(5) temperature dependence of magnetic susceptibility (thermomagnetic curves); and
(6) anisotropy of magnetic susceptibility (AMS).

Estimates of the concentration of magnetic minerals can be obtained from parameters such as k, ARM and IRM, while thermomagnetic curves are most diagnostic of magnetic mineral composition. Hysteresis loops and artificial remanences provide information about grain size and domain state of magnetic minerals, while AMS is an indicator of magnetic fabric.

RESULTS

In the lower sequence of the CIROS-1 core, there is a marked alternation of magnetic properties (Fig. 3) between zones of high magnetic intensity (where k generally exceeds 5×10^{-4} SI, and NRM generally exceeds 10^{-3} A m^{-1}) and zones of low magnetic intensity (where these parameters are at least two orders of magnitude lower). ARM and IRM, measured on representative specimens, display similar alternations between high and low intensities in these zones.

Five zones of contrasting magnetic properties exist in the lower sequence of the CIROS-1 core (Fig. 3), including two high-magnetic-intensity zones (between 430 and 540 mbsf, and

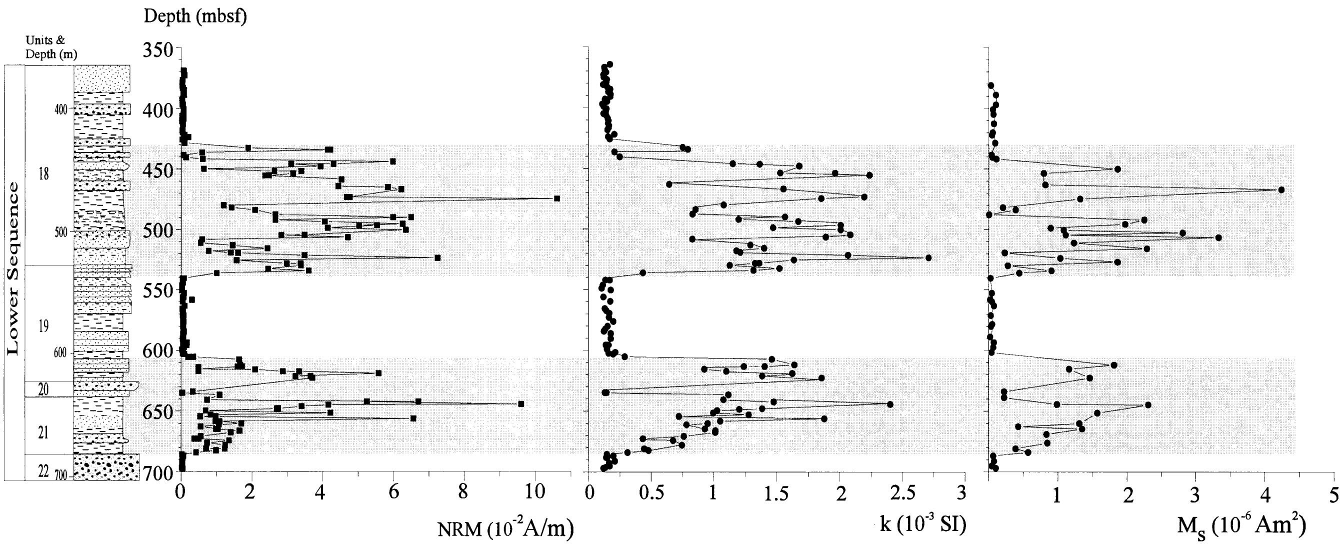

Figure 3. Downcore variation of the basic magnetic properties in the lower sequence of the CIROS-1 core. NRM = natural remanent magnetization; k = low-field magnetic susceptibility; M_S = saturation magnetization. All the parameters are consistent with a well-defined magnetic pattern that involves an alternation between low- and high-magnetic-intensity zones. The shaded areas mark the intervals with high concentrations of magnetic minerals.

between 605 and 685 mbsf) and three low-magnetic-intensity zones (above 430 mbsf, between 540 and 605 mbsf, and below 685 mbsf). This magnetic zonation is not clearly associated with lithostratigraphic boundaries within the CIROS-1 core.

Magnetite appears to be the main magnetic mineral throughout the sequence. For example, IRM acquisition curves generally approach saturation below 300 mT (Figs 4a and b), Curie temperatures lie between 550 and 580 °C (Fig. 4c) and the maximum unblocking temperature of the dominant low-coercivity component of a three-component IRM (*cf*. Lowrie 1990) lies between 550 and 600 °C (Fig. 4d). All of these properties are characteristic of magnetite (*cf*. Hunt, Moskowitz & Banerjee 1995). Saturation is not reached until 300–500 mT for some samples from the low-magnetic-intensity zones (Fig. 4a). This suggests the presence of a relatively high-coercivity phase. Thermal unblocking of a high-coercivity phase (coercivity between 500 and 900 mT) occurs over a broad range of temperatures up to 600 °C, which may be due to fine-grained haematite. Despite this possibility, it is clear that magnetite is the main magnetic component in the low-magnetic-intensity zones. In addition, some samples from the high-magnetic-intensity zones display a change in slope at about 350 °C during thermal demagnetization of a composite IRM (Fig. 4d). This may be due to the presence of maghemite, as is also suggested by a continuous decrease in the magnetic susceptibility after the 300 °C heating step, which may result from alteration of maghemite to haematite (e.g. De Boer & Dekkers 1996). However, as is the case in the low-magnetic-intensity zones, magnetite also appears to be the main magnetic mineral in the high-magnetic-intensity zones.

The grain size and domain state of the magnetic minerals also appear to be relatively constant. In particular, the hysteresis data indicate only a minor difference in grain size between high- and low-magnetic-intensity zones (Fig. 5). Apparent differences in grain size are suggested by variations in H_{cr}/H_c (Fig. 5a), and are consistent with the presence of slightly finer-grained magnetite in the high-magnetic-intensity zones. However, the hysteresis parameters indicate that the magnetite is dominantly in the pseudo-single-domain state (Fig. 5b).

Variations in magnetic fabric also alternate in a way that is consistent with variations in the other magnetic parameters. In the zones where the concentration of ferrimagnetic minerals is low, the magnetic fabric is mostly controlled by the paramagnetic clay matrix. For these samples, the degree of anisotropy ($P' < 1.05$) and the shape of the AMS ellipsoid (oblate, $T > 0$) (Jelinek 1981) are typical of undeformed to weakly deformed compacted sediments. In contrast, in the zones where the concentration of ferrimagnetic minerals is high, the magnetic fabric is controlled by the ferrimagnetic fraction, with P' values that are distinctly higher (up to 1.18) and AMS ellipsoids that have both oblate and prolate shapes (Fig. 6). A distinction can also be made between the two low-magnetic-intensity zones in terms of the range of P' values: for the interval between 366 and 430 mbsf, $1.001 < P' < 1.012$, while for the interval between 540 and 605 mbsf, $1.01 < P' < 1.04$.

DISCUSSION AND CONCLUSIONS

Magnetic susceptibility is relatively insensitive to variations in magnetite grain size over a broad range of grain diameters (Heider, Zitzelsberger & Fabian 1996) and is commonly used to estimate the concentration of magnetic material in sediments. Given that magnetic grain size does not appear to vary markedly between the high- and low-magnetic-intensity zones, it can be inferred that the observed alternations between high and low values of k, NRM, ARM, IRM and M_s suggest that the zonation of magnetic properties is due primarily to variations in the abundances of ferrimagnetic grains (Fig. 3).

Bridle & Robinson (1989) studied diagenesis of the CIROS-1 core and found only minor *in situ* diagenetic alteration of detrital clays, some carbonate and zeolite cement formation, and ubiquitous pyrite formation within muddier horizons. Canfield & Berner (1987) demonstrated that magnetite dissolution should be ubiquitous in sedimentary environments that sup-

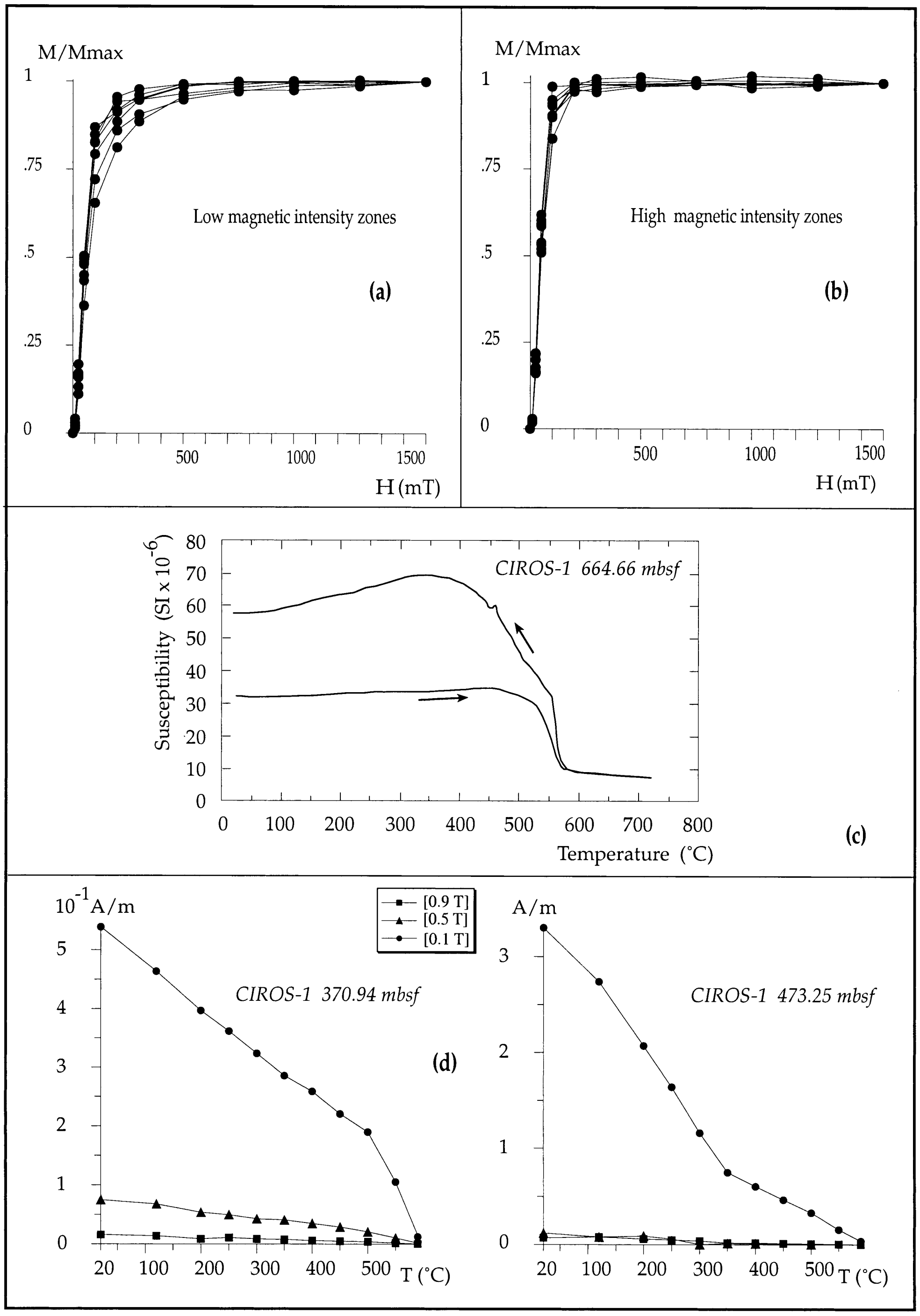

Figure 4. Representative plots providing evidence that magnetite is the main magnetic mineral in the lower sequence of the CIROS-1 core. (a) and (b) IRM acquisition curves for low- and high-magnetic-intensity zones, respectively. Saturation of IRM is reached in fields of 0.1–0.2 T for specimens from the high-magnetic-intensity zones and in fields of 0.3–0.5 T for specimens from the low-magnetic-intensity zones. (c) Variation of the magnetic susceptibility k in a heating–cooling cycle from room temperature to 700 °C. A Curie temperature is indicated at 550–580 °C. (d) Thermal demagnetization curves of a composite IRM (Lowrie 1990), for representative specimens from the low-magnetic-intensity zones (370.94 mbsf) and from the high-magnetic-intensity zones (473.25 mbsf). Both specimens are dominated by the low-coercivity component, with maximum unblocking temperatures between 550 and 600 °C. An inflection in the low-coercivity curve at 350 °C is often present in specimens from the high-magnetic-intensity zones.

port active sulphate reduction and pyrite formation. Despite the fact that the entire CIROS-1 sequence was deposited under reducing conditions, as suggested by widespread pyrite formation throughout the core, significant concentrations of detrital magnetite have survived early diagenetic dissolution and are apparently responsible for the magnetic signature of the CIROS-1 sediments. We therefore infer that the variations in the magnetic properties of the CIROS-1 core are not primarily controlled by post-depositional diagenetic processes, but are due to changes in the concentration of detrital material brought

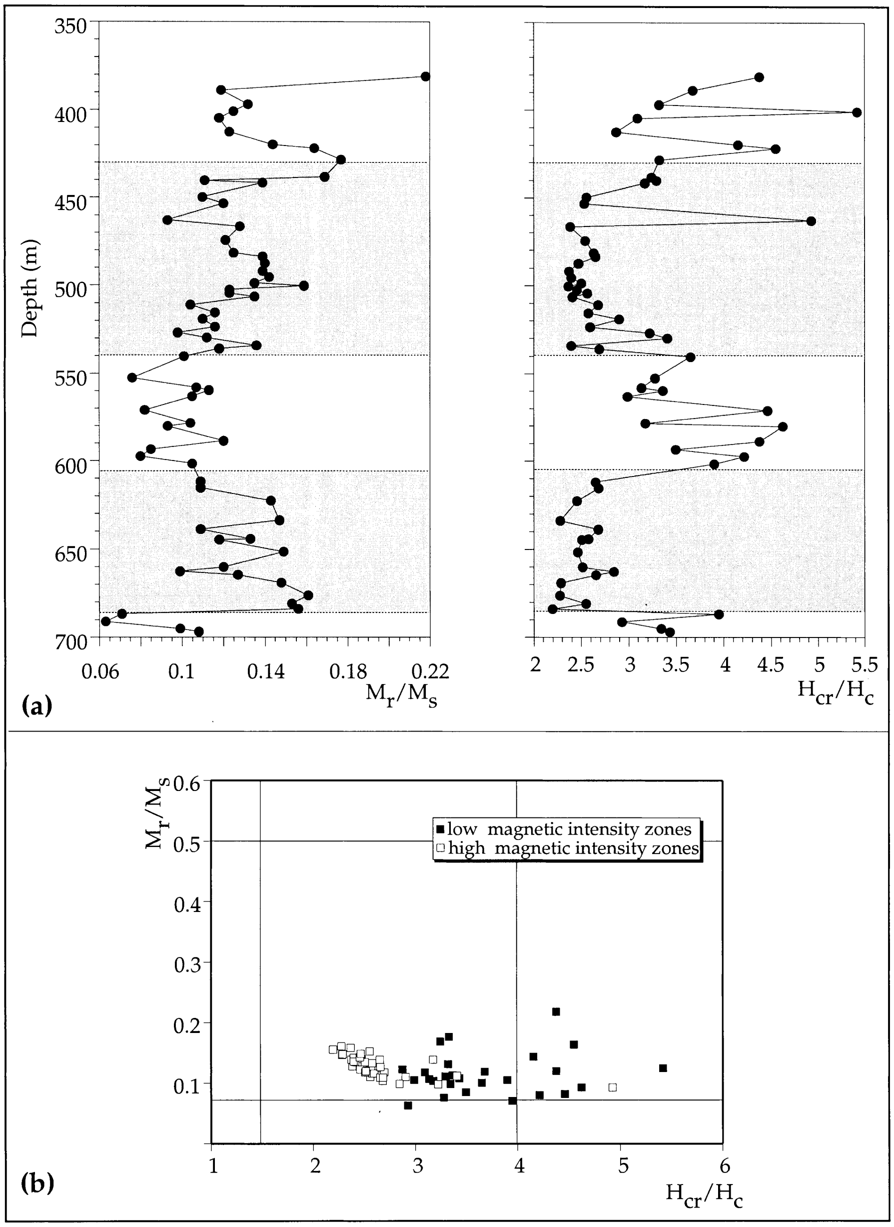

Figure 5. (a) Downcore variation of M_r/M_s and H_{cr}/H_c in the lower sequence of the CIROS-1 core. The shaded areas mark the intervals with high concentrations of magnetic minerals. (b) M_r/M_s vs. H_{cr}/H_c (after Day, Fuller & Schmidt 1977).

into the VLB. A possible explanation for the observed alternations in rock magnetic properties is that there were temporal variations in supply from the sedimentary source rocks (for example by progressive erosion through successive units, with contrasting magnetic properties, in the source area or by switching of sedimentary source regions). However, clast and sand-grain petrography and whole-rock geochemistry are almost uniform throughout the core (George 1989; Roser & Pyne 1989), and the slight variations that have been reported do not coincide with the observed changes in magnetic properties.

A second explanation for the alternating magnetic properties, which we prefer, is that the source rocks (crystalline granites and metamorphic rocks of the Precambrian basement, quartzose sediments of the Beacon Supergroup, and sills and dykes of the Ferrar Group) have remained similar throughout the time of transport and deposition of the lower sequence of the CIROS-1 core in the VLB, but that there was climatic modulation of weathering and transport mechanisms of the sediments. Clay mineralogy studies provide additional evidence for this interpretation (Claridge & Campbell 1989; Ehrmann 1997; Setti *et al.* 1997). The high-resolution studies of Ehrmann (1997) indicate that the observed changes in rock magnetic properties are paralleled by changes in clay mineralogy in the lower part of the CIROS-1 core (Fig. 7). In particular, the highest concentration of detrital smectites, and the lowest degree of detrital smectite crystallinity, coincide with zones of high magnetite concentration. The clay mineralogy in the high-magnetic-intensity zones is indicative of chemical weathering of source rocks under warmer and more humid conditions than those that currently exist in Antarctica. Claridge & Campbell (1989) identified ferriferous beidellite as the dominant smectite, which formed from acid leaching of dolerite and

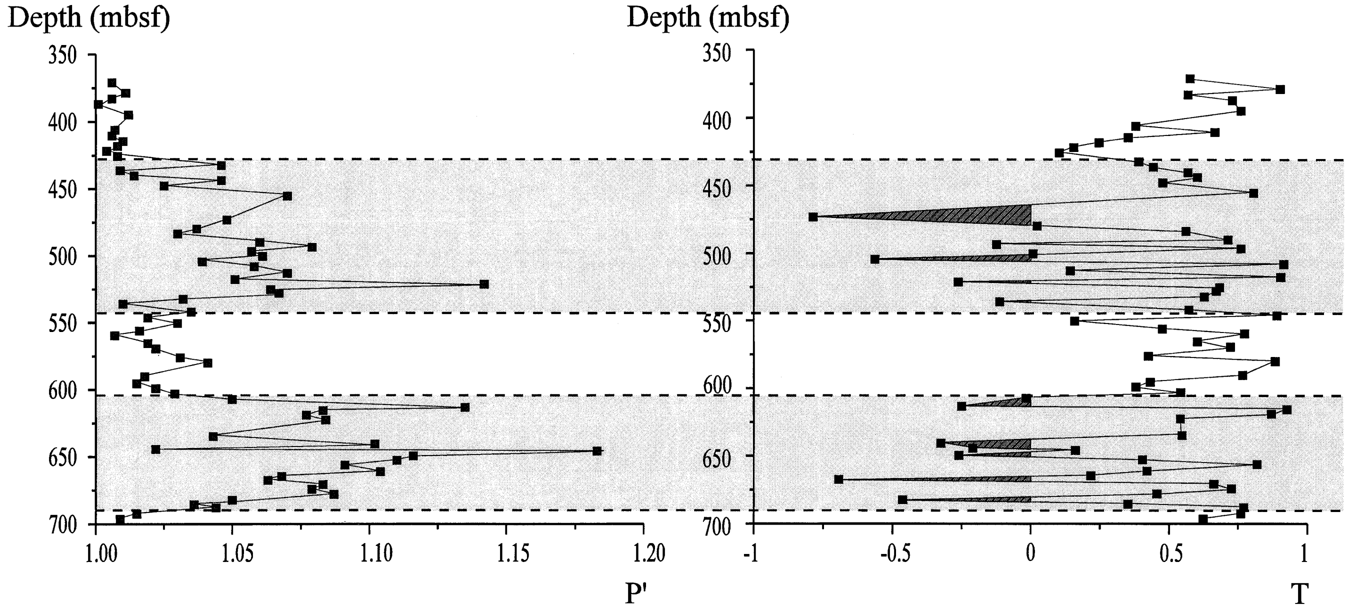

Figure 6. Downcore variation of the corrected anisotropy degree P' and the shape factor T (Jelinek 1981) in the lower sequence of the CIROS-1 core. $T=1$ for a perfect oblate ellipsoid, $T=-1$ for a perfect prolate ellipsoid and $T=0$ for neutral ellipsoids. The shaded areas mark the intervals with high concentrations of magnetic minerals.

other basic rocks of the Ferrar Group in podzolized soils in a cool to cold temperate climate (similar to that in Patagonia or Tierra del Fuego today). Furthermore, the absence of kaolin or halloysite demonstrates that the weathering regime was neither as strong nor as humid as in lower-latitude settings (Claridge & Campbell 1989). Above approximately 430 mbsf, the clay mineralogy is characterized by a smectite-poor assemblage, which is indicative of physical weathering of source rocks under a climatic regime that is similar to the present climate in Antarctica.

Evidence for this shift in clay mineralogy can be seen in the low-magnetic-intensity zones where the magnetic fabric appears to be determined by the paramagnetic clay matrix. As noted above, the susceptibility anisotropy, as measured by the P' parameter, is lower in the upper low-magnetic-intensity zone than in the middle low-magnetic-intensity zone. Because it is unlikely that the change in the degree of anisotropy could be due to different degrees of compaction or to different sedimentary processes, we suggest that it reflects a change in composition of the clay matrix.

The shift in clay mineralogy from smectite-rich to smectite-poor assemblages at other Antarctic locations has been inferred to represent the onset of continental glaciation in East Antarctica, and has been dated at the Eocene/Oligocene boundary (Ehrmann & Mackensen 1992; Ehrmann 1997). Rock magnetic properties demonstrate evidence for this shift at approximately 430 mbsf in the CIROS-1 core, which is within the Eocene/Oligocene transition interval, but which slightly precedes the Eocene/Oligocene boundary proper, as demonstrated by Wilson *et al.* (1998). Chemical weathering on the Antarctic craton, which resulted in the high smectite concentrations beneath approximately 430 mbsf, may also be responsible for the increased magnetite concentration beneath approximately 430 mbsf in the CIROS-1 core. As discussed above, Claridge & Campbell (1989) argue that the smectite originates from chemical weathering of rocks from the Ferrar Group (Ferrar Dolerite and Kirkpatrick Basalt). Increased chemical weathering of the dolerite and basalt would also liberate more ferrimagnetic minerals for erosion and transportation to the VLB. The likely presence of maghemite in the high-magnetic-intensity samples may be due to enhanced pedogenesis associated with the relatively warmer, wetter climate. Minor changes in source material could have occurred with little change to the total volume of sediment being eroded and deposited in the VLB. The proportion of sand-sized particles derived from the Ferrar Dolerite varies from 1 to 2 per cent in some intervals and from 4 to 11 per cent in other intervals (George 1989). Because the sampling interval was coarse and because rock magnetic measurements are sensitive to much finer (mud-sized) grains, the results of the sand petrography are not directly diagnostic of the source of the magnetic variations. However, they do indicate that the contribution of the Ferrar Group to the overall sediment load was relatively small. Because the Ferrar Group rocks have much higher concentrations of magnetic minerals than other possible sources (Heimann *et al.* 1994; Barrett, Elliot & Lindsay 1986), even a small increase (< 5 per cent) in the proportion of sediment derived from these rocks would probably increase the concentration of magnetic minerals that accumulated in the VLB sediments. Magnetostratigraphic analyses (Wilson *et al.* 1998) indicate that intervals of high magnetic mineral concentration are coincident with zones of slightly increased sedimentation rates in the CIROS-1 core.

The clear variations observed in the magnetic properties of the lower sequence of the CIROS-1 core are more pronounced than, but are also coincident with, the cyclic smectite variations reported by Ehrmann (1997). As discussed above, intervals of increased magnetite and smectite concentrations (≈430–540 mbsf and 605–685 mbsf) probably originated from increased chemical weathering of the Ferrar Group source rocks, which resulted in greater erosion of ferrimagnetic and smectite clay minerals into the VLB. Conversely, intervals of decreased magnetite and smectite concentrations in the CIROS-1 core

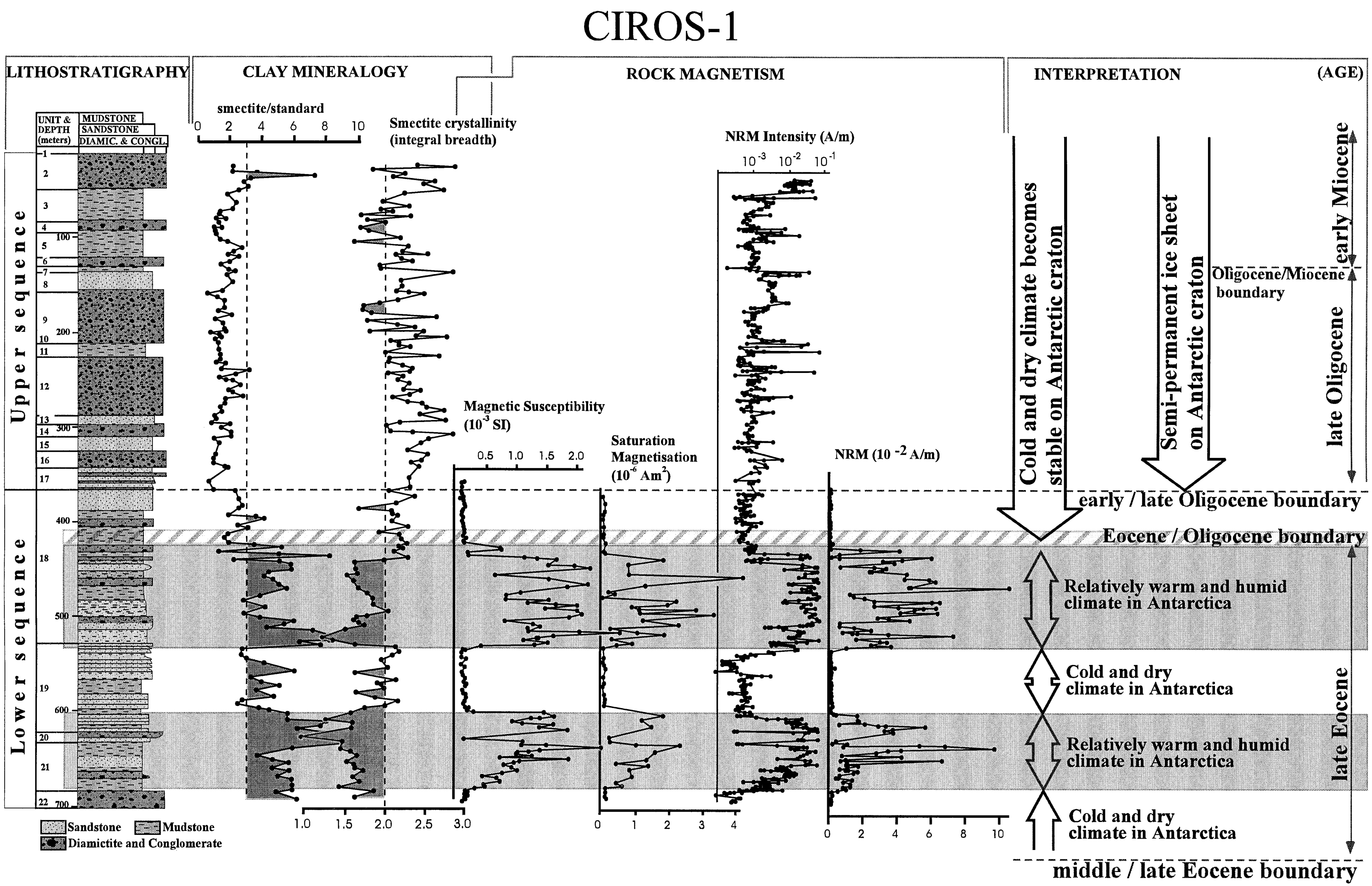

Figure 7. Downcore variation of the smectite/standard ratio and smectite crystallinity in the CIROS-1 core (redrawn from Ehrmann 1997) compared with rock magnetic parameters, including magnetic susceptibility, saturation magnetization and NRM intensity. The NRM record for the entire core is from Rieck (1989) and is plotted on a logarithmic scale, while the NRM record for the lower half of the core is from Wilson *et al.* (1998) and is plotted on a linear scale. The shaded areas mark the intervals with high concentrations of magnetic minerals. The cross-hatched area marks the Eocene/Oligocene boundary (after Wilson *et al.* 1998) and the dashed line marks the unconformity that separates the lithostratigraphy into upper and lower sequences (Hambrey *et al.* 1989). The climatic interpretation is given on the right-hand side of the figure.

(above approximately 430 mbsf, 540–605 mbsf, and below 685 mbsf) probably resulted from decreased chemical weathering of the Ferrar Group source rocks, with only low amounts of ferrimagnetic minerals and larger concentrations of clay minerals being shed into the VLB. This pattern of rock magnetic variation was previously observed by Rieck (1989), who reported NRM intensity variations for the whole CIROS-1 core. This observation, in conjunction with clay mineralogy and more detailed rock magnetic evidence, can be interpreted to be a manifestation of relatively cold and dry climatic conditions in Antarctica subsequent to the Eocene/Oligocene boundary. Rieck (1989) observed a slight increase in NRM intensity above 200 mbsf (Fig. 7). This, however, is due to increased input from the McMurdo Volcanic Group towards the top of the core.

Wilson *et al.* (1998) concluded from magnetobiostratigraphy and lithostratigraphy of the CIROS-1 core that major cooling of the Antarctic continent had begun by the Eocene/Oligocene boundary (33.4 Ma, ≈410–420 mbsf) and possibly as early as 34.5 Ma (500 mbsf). This cooling did not result in the formation of a major ice sheet on the Antarctic craton until the early/late Oligocene boundary (28.5 Ma, 365 mbsf). Major glaciation did not affect the CIROS-1 site until about 0.5 Myr later (≈28 Ma, 345 mbsf). Cooling and glacial intensification are recorded by the sharp decrease in magnetite concentration at approximately 430 mbsf. However, the environmental magnetic signal reveals two major cool episodes before the Eocene/Oligocene boundary. These earlier cool intervals (540–605 mbsf, 35–36 Ma; and >685 mbsf, >36.5 Ma) do not appear to coincide with any lithostratigraphic indicators of sea-level lowering or glacial advance. It is therefore likely that these intervals represent times when the Antarctic continent was cool, with perhaps limited ice cover; however, any ice cover was not as extensive as in the 28.5 Ma event on the Antarctic craton. The base of the CIROS-1 core therefore appears to represent the end of a cool interval that had begun at least by earliest late Eocene time.

This interpretation is consistent with palaeoclimatic trends that have been inferred from the global turnover in marine and terrestrial biota that occurred from late middle Eocene to early Oligocene time. The available palaeontological evidence, as reviewed by Berggren & Prothero (1992), indicates that stepwise-increasing extinctions occurred in this time interval, and were superimposed on a gradual cooling trend. Berggren & Prothero (1992) recognized the first major wave of extinctions at the middle/late Eocene boundary and related this to the inception of a continent-wide glaciation in Antarctica. The Eocene/Oligocene boundary was characterized by fewer extinctions and initiated a prolonged period of cooler and drier global climate that persisted throughout the Oligocene.

The environmental magnetic record in the CIROS-1 core supports the suggestion of Berggren & Prothero (1992) that the Antarctic craton became isolated and that major cooling of the continent had begun by the earliest late Eocene. Our data also support the interpretation of Lawver *et al.* (1992) that, although Australia had been separated from Antarctica for quite some time, final separation of the South Tasman Rise and development of major Southern Ocean circulation did not begin until around the middle/late Eocene boundary. However, we do not find firm evidence to demonstrate that major glaciation occurred in Antarctica in the late Eocene. It is more likely that the continent was subject to a period of climatic deterioration that began by the earliest late Eocene, prior to growth of a major ice sheet at about the time of the Eocene/Oligocene boundary.

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REFERENCES

Barrett, P.J., 1989. Introduction, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica,* pp. 5–6, ed. Barrett, P.J., Department of Scientific and Industrial Research Bulletin, **245,** DSIR Publishing, Wellington.

Barrett, P.J., Elliot, D.H. & Lindsay, J.F., 1986. The Beacon Supergroup (Devonian–Triassic) and Ferrar Group (Jurassic) in the Beardmore Glacier area, Antarctica, *Antarct. Res. Ser.*, **36,** 339–428.

Barrett, P.J., Hambrey, M.J., Harwood, D.M., Pyne, A.R. & Webb, P.N., 1989. Synthesis, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica,* pp. 241–251, ed. Barrett, P.J., Department of Scientific and Industrial Research Bulletin, **245,** DSIR Publishing, Wellington.

Bartek, L.R., Henrys, S.A., Anderson, J.B. & Barrett, P.J., 1996. Seismic stratigraphy of McMurdo Sound, Antarctica: implications for glacially influenced early Cenozoic eustatic changes?, *Mar. Geol.*, **130,** 79–98.

Berggren, W.A. & Prothero, D.R., 1992. Eocene–Oligocene climatic and biotic evolution: an overview, in *Eocene–Oligocene Climatic and Biotic Evolution,* pp. 1–28, eds Prothero, D.R. & Berggren W.A., Princeton University Press, Princeton, NJ.

Bridle, I.M. & Robinson, P.H., 1989. Diagenesis, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica,* pp. 201–207, ed. Barrett, P.J., Department of Scientific and Industrial Research Bulletin, **245,** DSIR Publishing, Wellington.

Canfield, D.E. & Berner, R.A., 1987. Dissolution and pyritization of magnetite in anoxic marine environments, *Geochim. cosmochim. Acta,* **51,** 645–659.

Claridge, G.G.C. & Campbell, I.B., 1989. Clay mineralogy, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica,* pp. 185–193, ed. Barrett, P.J., Department of Scientific and Industrial Research Bulletin, **245,** DSIR Publishing, Wellington.

Day, R., Fuller, M. & Schmidt, V.A., 1977. Hysteresis properties of titanomagnetites: grain-size and compositional dependence, *Phys. Earth planet. Inter.*, **13,** 160–267.

De Boer, C.B. & Dekkers, M.J., 1996. Grain-size dependence of the rock magnetic properties for a natural maghemite, *Geophys. Res. Lett.*, **23,** 2815–2818.

Diester-Haass, L., Robert, C. & Chamley, H., 1996. The Eocene–Oligocene preglacial–glacial transition in the Atlantic sector of the Southern Ocean (ODP Site 690), *Mar. Geol.*, **131,** 123–149.

Ehrmann, W.U., 1997. Smectite concentrations and crystallinities: indications for Eocene age of glaciomarine sediments in the CIROS-1 drill hole, McMurdo Sound, Antarctica, in *The Antarctic Region,* *Geological Evolution and Processes,* pp. 771–780, ed. Ricci, C.A., Museo Nazionale dell'Antartide, Siena, Italy.

Ehrmann, W.U. & Mackensen, A., 1992. Sedimentological evidence for the formation of an East Antarctic ice sheet in Eocene/Oligocene time, *Palaeogeog. Palaeoclimat. Palaeoecol.*, **93,** 85–112.

George, A., 1989. Sand provenance, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica,* pp. 159–167, ed. Barrett, P.J., Department of Scientific and Industrial Research Bulletin, **245,** DSIR Publishing, Wellington.

Hambrey, M.J. & Barrett, P.J., 1993. Cenozoic sedimentary and climatic record, Ross Sea region, Antarctica, in *The Antarctic Paleoenvironment: a Perspective on Global Change,* pp. 91–124, eds Kennett, J.P. & Warnke, D.A., Antarctic Research Series, **60,** AGU, Washington, DC.

Hambrey, M.J., Barrett, P.J. & Robinson, P.J., 1989. Stratigraphy, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica,* pp. 23–48, ed. Barrett, P.J., Department of Scientific and Industrial Research Bulletin, **245,** DSIR Publishing, Wellington.

Hannah, M.J., 1994. Eocene dinoflagellates from CIROS-1 drill hole, McMurdo Sound, *Terra Antarctica,* **1,** 371.

Harwood, D.M., Barrett, P.J., Edwards, A.R., Rieck, H.J. & Webb, P.N., 1989. Biostratigraphy and chronology, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica,* pp. 231–239, ed. Barrett, P.J., Department of Scientific and Industrial Research Bulletin, **245,** DSIR Publishing, Wellington.

Heider, F., Zitzelsberger, A. & Fabian, K., 1996. Magnetic susceptibility and remanent coercive force in grown magnetite crystals from 0.1 μm to 6 mm, *Phys. Earth planet. Inter.*, **93,** 239–256.

Heimann, A., Fleming, T.H., Elliot, D.H. & Foland, K.A., 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.

Hunt, C., Moskowitz, B.M. & Banerjee, S.K., 1995. Magnetic properties of rocks and minerals, *Rock Physics Phase Relations, AGU Reference Shelf,* **3,** 189–204, AGU, Washington, DC.

Jelinek, V., 1981. Characterization of the magnetic fabrics of rocks, *Tectonophysics,* **79,** 63–67.

Kennett, J.P., 1977. Cenozoic evolution of Antarctic glaciation, the Circum-Antarctic Ocean, and their impact on global paleoceanography, *J. geophys. Res.*, **82,** 3843–3860.

Lawver, L.A., Gahagan, L.M. & Coffin, M.F., 1992. The development of paleoseaways around Antarctica, in *The Antarctic Paleoenvironment: a Perspective on Global Change,* pp. 7–30, eds Kennett, J.P. & Warnke, D.A., Antarctic Research Series, **56,** AGU, Washington, DC.

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

Reynolds, R.L. & King, J.W., 1995. Magnetic record of climate change, *Rev. Geophys., Suppl., US Natl Rep. to IUGG, 1991–94,* 101–110.

Rieck, H.J., 1989. Paleomagnetic stratigraphy, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica,* pp. 153–157, ed. Barrett, P.J., Department of Scientific and Industrial Research Bulletin, **245,** DSIR Publishing, Wellington.

Robinson, P.H., Pyne, A.R., Hambrey, M.J., Hall, K.J. & Barrett, P.J., 1987. *Core Log, Photographs and Grain Size Analyses from the CIROS-1 Drillhole, Western McMurdo Sound, Antarctica,* Antarctica Data Series, **14,** Victoria University of Wellington.

Roser, B.P. & Pyne, A.R., 1989. Whole rock geochemistry, in *Antarctic Cenozoic History from the CIROS-1 Drillhole, McMurdo Sound, Antarctica,* pp. 175–184, ed. Barrett, P.J., Department of Scientific and Industrial Research Bulletin, **245,** DSIR Publishing, Wellington.

Setti, M., Marinoni, L., López-Galindo, A. & Ben Aboud, A., 1997. XRD, SEM and TEM investigation of smectites of the core CIROS-1 (Ross Sea, Antarctica), *Terra Antarctica,* **4(2),** 119–125.

Verosub, K.L. & Roberts, A.P., 1995. Environmental magnetism: past, present and future, *J. geophys. Res.*, **100,** 2175–2192.

van der Wateren, F.M. & Verbers, A.L.L.M., 1993. Climate change, rifting and landscape evolution in the Ross embayment, *EOS, Trans. Am. geophys. Un.*, **74,** 490–491.

Wei, W., 1991. Evidence for an earliest Oligocene abrupt cooling in the surface waters of the Southern Ocean, *Geology,* **19,** 780–783.

Wilson, G.S., Roberts, A.P., Verosub, K.L., Florindo, F. & Sagnotti, L., 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.

Wise, S.W. Jr, Breza, J.R., Harwood, D.M. & Wei, W., 1991. Paleogene glacial history of Antarctica, in *Controversies in Modern Geology,* pp. 133–171, ed. Muller, D.W., Academic Press, New York.