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Evidence for iceberg armadas from East Antarctica in the Southern Ocean during the late Miocene and early Pliocene

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ABSTRACT

Sediments from Ocean Drilling Program Site 1165 in the Indian Ocean sector of the Southern Ocean (off Prydz Bay) contain a series of layers that are rich in ice-rafted debris (IRD). Here we present evidence that IRD-rich layers at Site 1165 at 7, 4.8, and 3.5 Ma record short-lived, massive discharges of icebergs from Wilkes Land and Adélie Land, more than 1500 kilometers to the east of the depositional site. This distant source of icebergs is clearly defined by the presence of IRD hornblende grains with $^{40}Ar/^{39}Ar$ ages of 1200–1100 Ma and 1550–1500 Ma, ages that are not found on the East Antarctic continent in locations closer to Site 1165. This observation requires enormous amounts of detritus-carrying drifting icebergs, most likely in the form of large icebergs. These events probably reflect destabilization, surge, and break-up of ice streams on the Wilkes Land and Adélie Land margins of the East Antarctic Ice Sheet, in the vicinity of the low-lying Aurora and Wilkes Basins. They occurred under warming conditions, but each coast seems to have produced ice-rafting events independently, at different times. The data presented here constitute the first evidence of far-traveled icebergs from specific source areas around the East Antarctic perimeter. Launch of these icebergs may have happened during quite dramatic events, perhaps analogous to “Heinrich Events” in the North Atlantic.

1. Introduction

In the Heinrich events of the glacial North Atlantic, rapid deposition of ice-rafted debris (IRD) results from massive discharge of debris-rich icebergs from a distant source (reviewed in Hemming, 2004). These remarkable events provide insight into the dynamics of large ice sheets, and they illustrate that low-lying parts of large ice sheets can quickly become unstable, and release armadas of icebergs into the ocean.

Marine geological records of IRD in the Southern Ocean offer a similar potential to investigate dynamic behavior of the Antarctic ice sheet. The history of the Antarctic ice sheets since 34 Ma is documented by marine $\delta^{18}O$ records, sea-level estimates, ice-proximal geological records from the continent and its shelf, and from numerical simulations of the ice sheet (discussed below). These studies provide broad estimates and constraints about variations in Antarctic ice volume, but much uncertainty remains about how much, how fast, and how often the ice volume changed during the late Miocene and early Pliocene. Moreover, information about rapid ice retreat events and the specific locations where they occur is rare.

Today, the East Antarctic Ice Sheet (EAIS) has a volume equivalent to ~52 m of sea-level rise, and the West Antarctic Ice Sheet has ~5 m (Fig. 1; Lythe and Vaughan, 2001). An ice volume on East Antarctica similar to today's was probably reached at the mid-Miocene climate transition (~14 Ma), when deep ocean temperatures cooled by 2–3 °C (e.g., Lear et al., 2000; Billups and Schrag, 2003; Shevenell et al., 2008), surface ocean temperatures cooled by 6–7 °C (Shevenell et al., 2004), and summers cooled by at least 8 °C in the Dry Valleys of the Transantarctic Mountains (Lewis et al., 2008). Estimates for the sea-level fall resulting from the mid-Miocene ice buildup include 25 ± 5 m (Miller et al., 1998) and 50 ± 5 m (John et al., 2004). For the early Pliocene, with global temperatures about 3 °C warmer than today (Ravelo et al., 2004), marine $\delta^{18}O$ records and temperature considerations put a limit of 25 m for the highstands (Kennett & Hodell, 1993), although some sea-level highstand estimates from paleoshorelines reach as high as 35 m ± 18 m (Dowsett and Cronin, 1990). The

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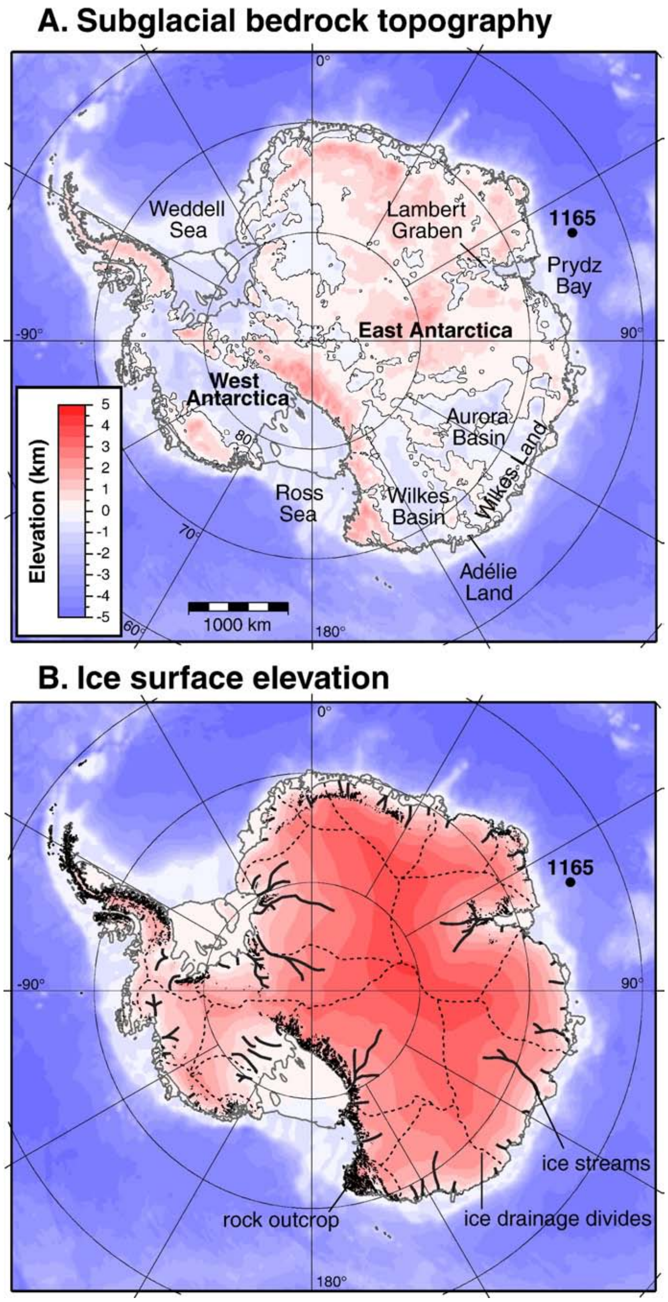

Fig. 1. A. Map showing subglacial bedrock elevation (Lythe and Vaughan, 2001) with the location of East Antarctic subglacial basins and seabed bathymetry (ETOPO-5). B. Map of ice surface elevation (ETOPO-5), rock outcrop (Antarctic Digital Database), ice drainage divides (Vaughan et al., 1999), and the approximate extent of major ice streams (Bamber et al., 2000).

IPCC (2007) reports a range of 15–25 m for mid-Pliocene sea-level rise, and this range of estimates presents differing scenarios for the stability of the East Antarctic Ice Sheet: 15 m would mean minimal melting (the less stable Greenland and West Antarctic ice sheets sum up to about 13 m), whereas 25 m would mean a reduction of about 23%. Even this could be an underestimate, as the Greenland and West Antarctic ice sheets are unlikely to melt entirely (e.g. Bamber et al., 2009).

The Lambert Graben and Wilkes Land ice margins are thought to be among the least stable parts of the East Antarctic Ice Sheet, and probably would be the first parts of the East Antarctic Ice Sheet to melt (e.g. Hambrey and McKelvey, 2000; Pollard and DeConto, 2009). There are few estimates of Antarctic temperature conditions during ice retreat, but in the Pliocene in the Southern Ocean north of Prydz Bay, sea surface temperatures reached over 4 °C warmer than present (Bohaty and Harwood, 1998; Whitehead and Bohaty, 2003), and winter sea ice was reduced up to 78% compared to present day (Whitehead et al., 2005).

Direct evidence for dynamic changes in ice sheet extent is documented in sediment sequences at ice-proximal locations. Drill cores in the northwest part of the Ross Ice Shelf taken by the ANDRILL program show 38 cycles of advance and retreat of the West Antarctic ice sheet over the last 5 Ma (Naish, et al., 2009). Glacial sediments of the Lambert Graben indicate at least four retreats of the ice sheet several hundred kms inland during the Miocene–Pliocene (Hambrey and McKelvey, 2000; Whitehead et al., 2006). On the other hand, the behavior of the ice sheet on the Wilkes Land margin of Antarctica is not yet well defined: numerical ice sheet models, under Pliocene conditions, suggest either very little ice margin retreat (Pollard and DeConto, 2009), or instability resulting in up to 500 km retreat inland along the entire Wilkes Land margin (Hill et al., 2007).

Large changes in an ice margin, such as the surge and break-up of an ice stream, would produce icebergs and lead to deposition of ice-rafted debris (IRD) on the sea bed. The marine geological record of IRD has the advantage of being more continuous and more easily dated than the more ice-proximal glaciogenic sediments. For example, the first occurrence of IRD in marine sediment cores in the Southern Ocean is the first unambiguous evidence for glacial inception on Antarctica (Hayes et al., 1975; Zachos et al., 1992). Knowledge of provenance of IRD (the location of the source areas where the debris was glacially eroded from) allows us to determine which areas of the ice sheet became unstable and calved icebergs, and what iceberg drift tracks looked like in the past. This approach has been successfully demonstrated in the North Atlantic Heinrich events of the last glacial cycle, where the provenance of the IRD has been established by its petrography (detrital carbonate, Bond et al., 1992), and geochemistry (argon, lead, and neodymium isotopes; e.g., Grousset et al., 1993; Gwiazda et al., 1996a, b, c; Hemming et al., 1998, 2002; reviewed by Hemming, 2004). A similar approach was used by Licht et al. (2005) and Farmer et al. (2006) to determine the provenance of till in the Ross Sea, and use it to describe the flowlines of the Ross ice sheet from East and West at the last glacial maximum.

In a series of previous studies on 30 core-top samples from all around Antarctica it was demonstrated that $^{40}Ar/^{39}Ar$ analyses of individual hornblende mineral grains from IRD layers, as well as Nd, Sr and Hf isotopes in the fine fraction of the same samples closely match the Antarctic continental source areas proximal to the sediment sites (Roy et al., 2007; Hemming et al., 2007; van de Flierdt et al., 2007, 2008). Here we apply this approach for the first time to study the dynamics of the East Antarctic Ice Sheet in the Miocene to Pliocene. By establishing the geochemical fingerprint of the IRD, we infer the provenance of the icebergs that carried it, and hence the ice margin that produced the bergs. Our results indicate major ice-rafting events in the late Miocene and Pliocene Southern Ocean from the Wilkes Land and Adélie Land coasts of East Antarctica. Large amounts of IRD from these coasts traveled over 1500 km, suggesting dramatic release mechanisms, reminiscent of North Atlantic Heinrich events.

2. ODP Site 1165

Site 1165 was drilled as part of ODP Leg 188, whose aim was to provide direct evidence for long- and short-term changes in Cenozoic paleoenvironments and glacial history in the Prydz Bay region (Cooper and O'Brien, 2004). Site 1165 reached 999 m below the seabed, and covers the last 21 Myr without major hiatuses (Florindo et al., 2003) (Fig. 2).

The IRD concentration increases through time, partly because of the reduction in sediment accumulation rates from more than 10 cm/kyr in the Early Miocene to less than 1 cm/kyr in the Pleistocene

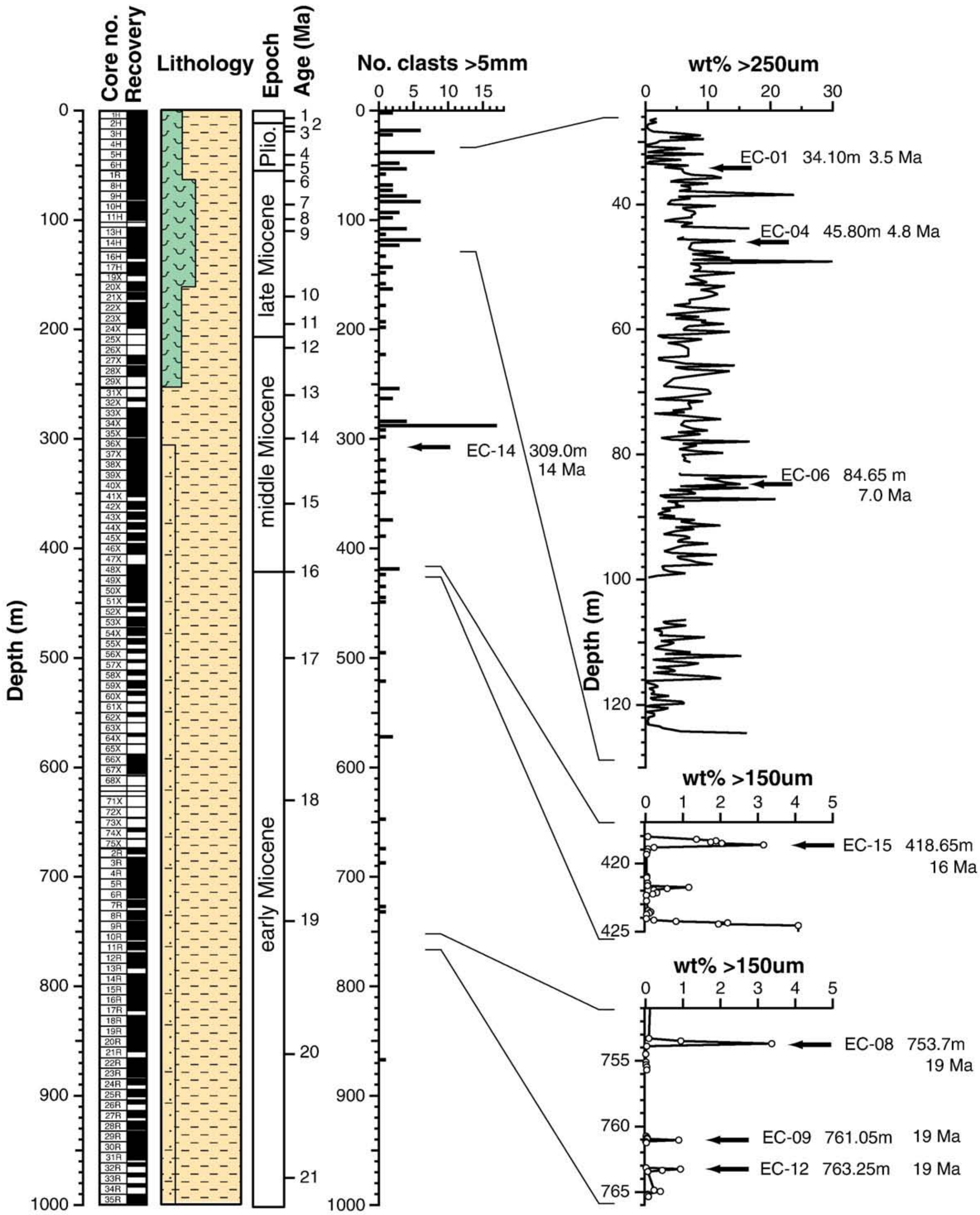

Fig. 2. Samples for this study from ODP Site 1165 in relation to core recovery, lithology (O'Brien et al., 2001), age (Florindo et al., 2003, transferred to the Lourens et al., 2004, time scale), and weight % coarse fraction (Grützner, 2003, Williams and Handwerger, 2005).

(Fig. 2) (Handwerger and Jarrard, 2003). In terms of short-term sedimentation rate changes, preliminary ^{3}He measurements indicate that IRD-rich layers were probably deposited faster than the background sedimentation rate at Site 1165 (McAuley et al., 2004). The style of occurrence of the IRD also changes through time. In the early Miocene sediments (21–16 Ma), IRD-rich layers are distinct from the IRD-free silty clay of the intervening sediment (Williams and Handwerger, 2005). In contrast, in the late Miocene section (7.0–7.5 Ma) studied by Grützner et al. (2003), IRD is present in the background sediment as well as in IRD-rich layers, marking an overall increase in IRD content after the mid-Miocene transition. The IRD-rich layers occur in bioturbated sediments that have more biogenic material than the background sediment, and this facies is widely interpreted in Antarctic marine cores to represent warm intervals (Ehrmann and Grobe, 1991; Grobe and Mackensen, 1992; Cowan, 2001; Grützner et al., 2003; Williams and Handwerger, 2005). Both Grützner et al. (2003) and Williams and Handwerger (2005) inferred that IRD was deposited preferentially during climate warming and glacial retreat.

In the present day oceanographic setting, Site 1165 can receive IRD from the local Prydz Bay area, and from icebergs carried on the westward flowing Polar Current from areas further east. Sediment waves in seismic sections on the continental rise offshore of Prydz Bay indicate that prevailing bottom currents have flowed westwards over

Table 1
Information on samples from Site 1165, ODP Leg 188. Age and long-term sedimentation rate from Florindo et al., 2003.

Sample code	Hole–core-section (depth in section, cm)	Depth below sea floor (m)	Approx. age (Ma)	Weight % >150 μm %	Approx. long-term sed. rate (cm/kyr)	Approx. dry bulk density (g/cm^3)	IRD mass accum. rate (g/cm^2kyr)
EC-01	1165B-4H-6 (80)	34.1	3.5	5.5	3.0	0.8	13.2
EC-02	1165B-6H-1 (70)	45.5	4.8	4.5	0.8	1.9	6.6
EC-03	1165B-6H-1 (90)	45.7	4.8	8.7	0.8	1.9	12.8
EC-04	1165B-6H-1 (100)	45.8	4.8	12.1	0.8	1.9	17.7
EC-05	1165B-6H-1 (120)	46.0	4.8	6.2	0.8	1.9	9.1
EC-06	1165B-10H-2 (35)	84.7	7.0	20.7	3.7	1.1	84.3
EC-14	1165B-36X-7 (20)	309.0	14.0	1.3	5.0	1.1	7.0
EC-15	1165B-48X-3 (75)	418.7	16.0	3.2	8.0	1.1	27.7
EC-08	1165C-10R-3 (120)	753.7	19.0	3.4	10.0	1.5	50.5
EC-09	1165C-11R-2 (45)	761.5	19.0	0.9	10.0	1.5	13.3
EC-11	1165C-11R-3 (115)	763.3	19.0	0.9	10.0	1.5	14.0

Site 1165 since the Oligocene (Kuvaas and Leitchenkov, 1992). Furthermore, modeling of past ice sheets and wind fields under elevated temperature conditions by DeConto et al. (2007) indicates that even with a markedly smaller ice sheet, the winds that drive the westward polar current still follow the coast. These results suggest that Miocene to Pliocene iceberg trajectories were probably similar to today.

3. Samples and methods

For this study, eight IRD-rich layers were sampled to provide snapshots of provenance and ice sheet behavior from 19 to 3.5 Ma (Fig. 2; Table 1). Samples were chosen from intervals of the core that had been previously studied (Grützner et al., 2003; Warnke et al., 2004; Grützner, 2003; Williams and Handwerger, 2005; Passchier, 2007). Additionally, three samples from background sediment close to the IRD-rich layer at 45.8 m depth (~4.8 Ma) were analyzed to assess variability in IRD provenance across this specific IRD event. Two-cm-thick samples were taken, representing approximately 200 to 2000 years of deposition, based on the long-term sedimentation rates. This time window is long enough to provide debris from many icebergs and not to be biased by IRD from any one berg. Hornblende concentration varied between 2.5 and 7 grains (>150 μm) per gram of dry sediment, making up between 0.5 and 3.5% of the grains in the quartz-dominated terrigenous >150 μm fraction.

Approximately thirty hornblende grains were picked from the >150 μm fraction of each sample, in order to yield a statistically robust argon age distribution, with a total of 365 grains analyzed. Samples and monitor standards were co-irradiated in the Cd-lined in-core facility (CLICIT) at the Oregon State reactor, and are referenced to a McClure Mountain hornblende (Mmhb) age of 523 Ma (Renne et al., 1998). Single-step laser fusion $^{40}Ar/^{39}Ar$ analyses for individual grains were processed at the L-DEO Ar geochronology lab using a CO_2 laser for fusion and a VG5400 noble gas mass spectrometer (for results and analytical details see supplementary data file in Appendix A). Single-step fusion is used in order to date enough hornblende IRD grains to define the $^{40}Ar/^{39}Ar$ age populations. The analytical error is about 1% (see supplementary data file in Appendix A), which is much less than the age differences between the source provinces (see below). Excess argon cannot be detected on an individual single-step fusion age, but the well-separated age populations show that such effects do not compromise the provenance interpretation.

4. Source characterization — onshore and offshore evidence from East Antarctica

Interpreting the provenance of downcore IRD in the Southern Ocean requires a solid understanding of Antarctic geology and thermochronology: in order to match the isotope geochemical fingerprint of ice-rafted material with its source area, we need to know the geochemical fingerprint of potential source areas. Data from onshore rock outcrops are limited by ice cover and uncertainty in how far they can be interpolated to represent the sub-ice geology. Offshore core-top IRD from 30 locations all around Antarctica has been used to identify the $^{40}Ar/^{39}Ar$ and Sm–Nd geochemistry of the nearby coastal sector, and provides a view of the geology under the ice sheet (Roy et al., 2007). Here we build on this work, by adding a significant amount of onshore regional detail to better understand potential sources for IRD at ODP Site 1165. Figs. 3 and 4 summarize the geochronological and thermochronological data for the area of interest (between ~70°E and ~150°E), and an extended summary of the existing onshore data is provided in Appendix A.

4.1. Offshore core-top $^{40}Ar/^{39}Ar$ dating of individual hornblende IRD grains

The $^{40}Ar/^{39}Ar$ age of hornblende grains documents the time when isotopic closure occurs, typically when the mineral cools below about

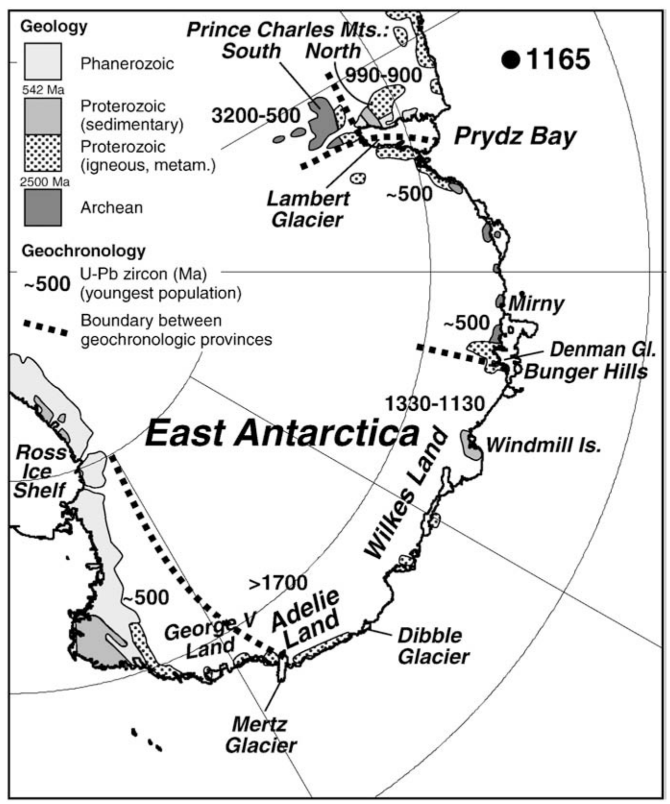

Fig. 3. Summary map of East Antarctica, showing geology (after Collins and Pisarevsky, 2005, and the Commission for the Geological Map of the World, 2000) and geochronological provinces based on the youngest population of zircon U–Pb ages in each area (Fitzsimons, 2000, 2003; see also Appendix A).

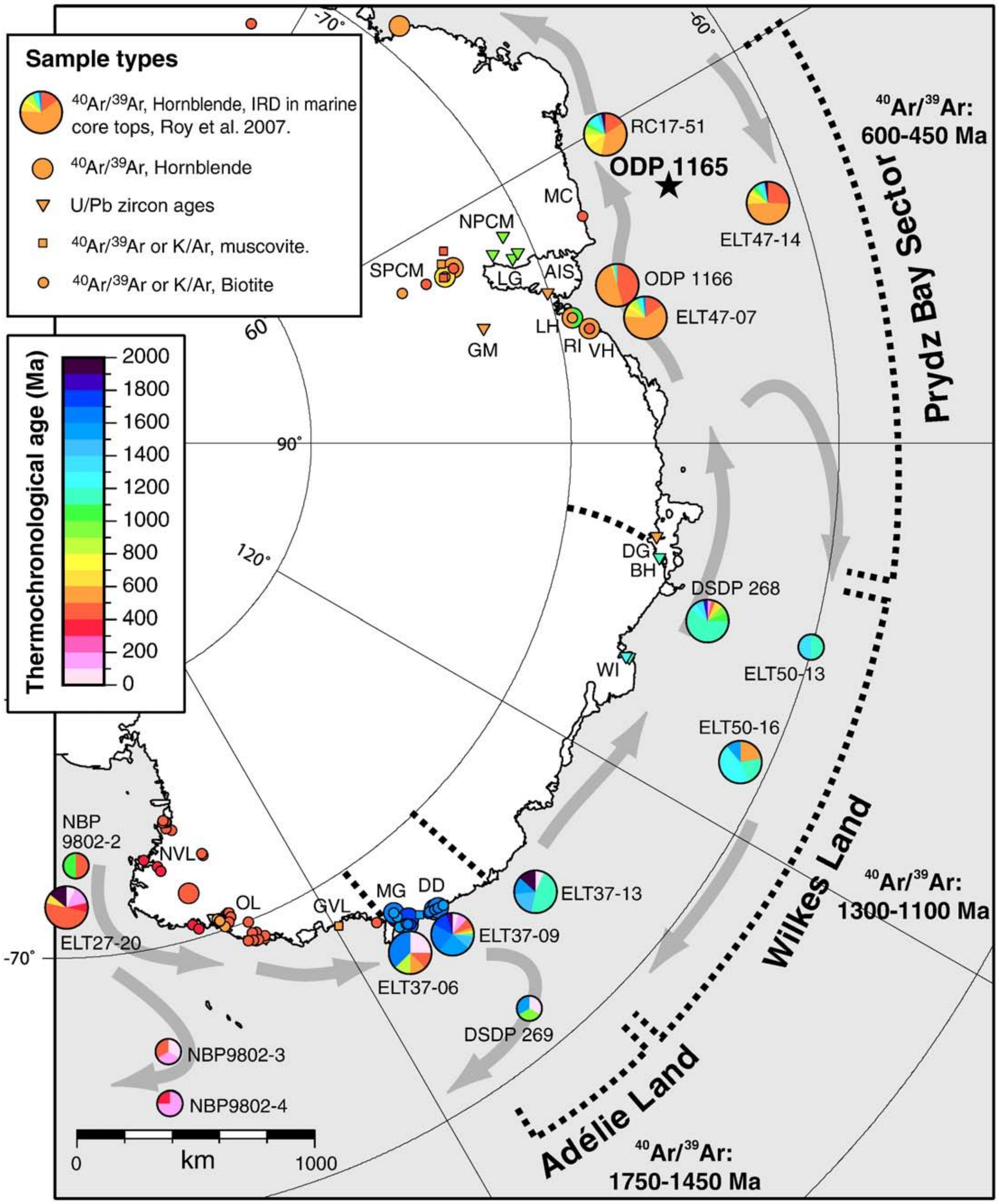

Fig. 4. Map of onshore and offshore thermochronological data of East Antarctica, 60°E to 170°E. Offshore $^{40}Ar/^{39}Ar$ data were measured on hornblende IRD grains (Roy et al., 2007) and are divided into three broad provinces: the Prydz Bay sector (main age peak 450–600 Ma), Wilkes Land (1100–1300 Ma), and Adélie Land (1450–1750 Ma). Analyses of DSDP Sites 268 and 269 are new for this paper. Each pie chart represents the population of ages at that site with each equal-area wedge representing one analysis; sites with less than five analyses are smaller pie charts. The onshore $^{40}Ar/^{39}Ar$ hornblende thermochronology (Phillips et al., 2007; Wilson et al., 2007; Tong et al., 1998, 2002; Duclaux et al., 2008; Di Vincenzo et al., 2007; Goodge, 2007) is supplemented with Ar/Ar biotite and muscovite data (Adams, 2006). In areas lacking Ar data zircon U–Pb data are plotted (Boger et al., 2000; Carson et al., 2000; Liu et al., 2006, 2007; Black et al., 1992; Sheraton et al., 1992; Post et al., 1996; Möller et al., 2002). The offshore IRD thermochronological signature is derived primarily from the adjacent coast, with a smaller admixture carried by the prevailing currents from the east (see text). AIS — Amery Ice Shelf, BH — Bunger Hills, DD — Dumont D'Urville, DG — Denman Glacier area, GM — Grove Mountains, GVL George V Land, LG — Lambert Glacier, LH — Larsemann Hills, MC — Mawson Coast, MG — Mertz Glacier area, NPCM — N. Prince Charles Mts., NVL — Northern Victoria Land, OL — Oates Land, RI — Rauer Islands, SPCM — S. Prince Charles Mts., VH — Vestfold Hills, WI — Windmill Islands.

450–550 °C (Reiners and Brandon, 2006). It reflects cooling after crystallization or on uplift through the closure isotherm after major orogenic events and associated metamorphism. Hornblende is a primary phase in basic to intermediate igneous rocks, and in metamorphic rocks above amphibolite grade. $^{40}Ar/^{39}Ar$ ages of detrital hornblende grains separated from marine sediments reflect the tectono-metamorphic history of the source area they were eroded from. Sedimentary recycling of hornblende grains is unlikely due to relatively easy weatherability. Results obtained by Roy et al. (2007) are shown in Figs. 4 and 5B. Age determinations on the Holocene core-top IRD samples reveal three distinct sectors with $^{40}Ar/^{39}Ar$ ages at 1450–1750 Myr off Adélie Land/eastern George V Land (135–145°E), 1100–1300 Myr offshore Wilkes Land (100–135°E), and 450–600 Myr off the Prydz Bay sector (60–100°E). New data from the core top of Deep Sea Drilling Program Site 268 (105.2°E, 63.9°S), helps constrain the western extent of the Wilkes Land sector (see supplementary data file in Appendix A).

4.2. Onshore evidence from $^{40}Ar/^{39}Ar$ dating

The $^{40}Ar/^{39}Ar$ hornblende dataset for onshore outcrops in the study area is limited to about 20 analyses from Adélie Land (Di

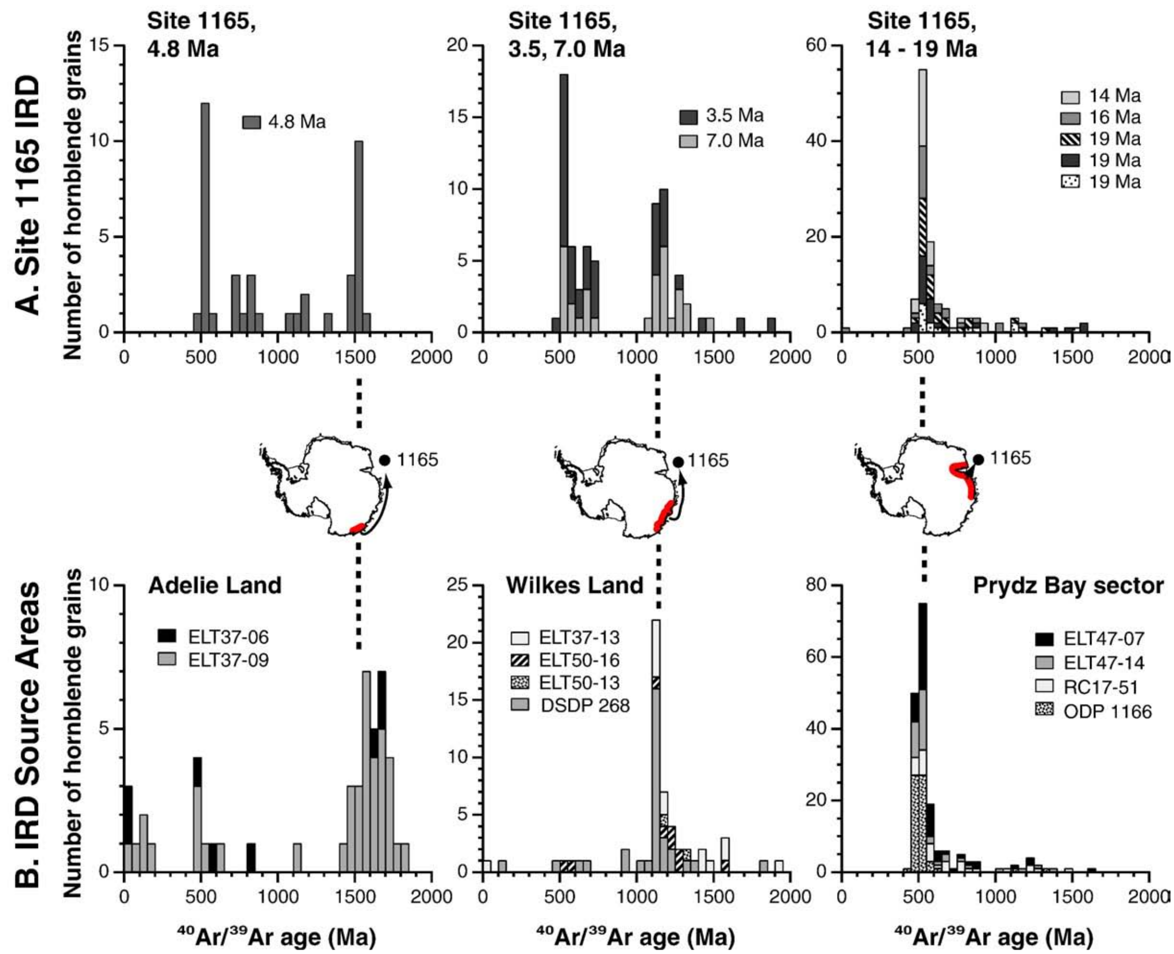

Fig. 5. Argon age histograms for: A. the samples from the IRD-rich layers at Site 1165 grouped by provenance sector; B. three provisional source sectors based on core-top $^{40}Ar/^{39}Ar$ data (Roy et al., 2007). All Site 1165 samples show $^{40}Ar/^{39}Ar$ ages of 500–600 Ma derived from the Prydz Bay sector; the 4.8 Ma sample shows a significant contribution of IRD from Adélie Land; and the 3.5 and 7.0 Ma samples have a significant contribution of IRD from Wilkes Land.

Vincenzo et al., 2007; Duclaux et al., 2007, 2008), the eastern coast of Prydz Bay (Zhao et al., 1992, 1997; Tong et al., 1998, 2002; Wilson et al., 2007), and the Southern Prince Charles Mountains (Phillips et al., 2007) (Fig. 4). Taken together with $^{40}Ar/^{39}Ar$ mica data presented in the same set of papers (biotite has an argon closure temperature of about 300–350 °C and muscovite 350 to 400 °C; Reiners and Brandon, 2006), we can identify ages of the last major tectono-metamorphic events from east to west as follows: (i) younger than 530 Ma to the east of the Mertz Shear Zone, (ii) 1500–1550 Myr in the Mertz Shear Zone, (iii) around ~ 1700 Myr in Adélie Land, (iv) 460 to 560 Myr in the Rauer Group (eastern Prydz Bay coast), (v) ~490 to 550 Ma in the Larsemann Hills (eastern Prydz Bay coast), and (vi) 486 to 524 Myr in the Southern Prince Charles Mountains (inland from Prydz Bay). No $^{40}Ar/^{39}Ar$ data are available for the coast between Adélie Land and the eastern part of the Prydz Bay coast (~80–135°E).

4.3. Onshore evidence from U–Pb dating

The $^{40}Ar/^{39}Ar$ age provinces and boundaries generally have a good correspondence to the geochronological provinces of East Antarctica, established mainly from U–Pb dating of zircons (Post et al., 1996; Möller et al., 2002; Fitzsimons, 2003; Duclaux et al., 2008). U–Pb zircon dates are set at a higher temperature than $^{40}Ar/^{39}Ar$ in hornblende, during crystallization of whole grains or metamorphic rims. They will be older than $^{40}Ar/^{39}Ar$ dates, though they may be close, if cooling is relatively fast and there is no subsequent reheating. From east to west there are four major crustal units: (i) the areas to the west of the Mertz Shear Zone (Victoria Land, Transantarctic Mountains), which are dominated by ages characteristic of the ~500 Ma Ross orogeny, (ii) the Adélie craton, exposed on Adélie Land, characterized by ages of 1500 to 2450 Ma, (iii) the Wilkes Land coast with two age clusters of U–Pb ages of 1140–1240 Ma and 1300–1340 Ma, and (iv) the Prydz Bay area, which is characterized by a ~490 to 650 Ma overprint (Pan-African or Pan-Gondwanaland orogeny; Veevers, 2003) on Proterozoic and Archean basement rocks. The Pan-African event is evident in U–Pb zircon ages around ~550–490 Ma in the northern part of the southern Prince Charles Mountains, the Grove Mountains, and along the Prydz Bay coast (Liu et al., 2007) with peak metamorphic conditions at ~535–530 Ma (Fitzsimons, 2003). In contrast, the southern part of the southern Prince Charles Mountains is characterized by Achaean to Proterozoic U–Pb zircon ages (Phillips et al., 2006) and the northern Prince Charles Mountains and the Mawson Coast are dominated by Meso- to Neoproterozoic ages (990–900 Ma; e.g. the Rayner Complex; Boger et al., 2000).

In summary, the East Antarctic coast is composed of tectono-metamorphic provinces that can be distinguished on the basis of $^{40}Ar/^{39}Ar$ and U–Pb thermochronological data. Dominant $^{40}Ar/^{39}Ar$ age peaks are at 1750–1450 Ma for Adélie Land/eastern George V Land, 1300–1100 Ma for Wilkes Land, and 600–450 Ma for the Prydz Bay sector (Fig. 3). The offshore $^{40}Ar/^{39}Ar$ hornblende data on Holocene IRD presented by Roy et al. (2007) not only compares well to the data from the neighboring coast, but also to the crustal provinces inferred from U–Pb zircon ages (Fig. 3). These provinces define source areas for IRD deposited between 19 and 3.5 Ma at ODP Site 1165.

5. $^{40}Ar/^{39}Ar$ results and provenance of downcore IRD at ODP Site 1165

All eight of the IRD-rich layers investigated at Site 1165 show a main peak in $^{40}Ar/^{39}Ar$ ages around 600–450 Ma (Fig. 5; supplementary data table in Appendix A). It is striking that only the three samples from the IRD-rich layers deposited after the mid-Miocene transition at 3.5, 4.8 and 7.0 Ma have additional major peaks in $^{40}Ar/^{39}Ar$ ages older than 600 Ma (Fig. 5), accounting for up to 50% of the IRD in those layers. The 3.5 and 7.0 Myr old IRD-rich layers have prominent $^{40}Ar/^{39}Ar$ age peaks at 1300 to 1100 Ma, while the 4.8 Myr old IRD-rich layer has a prominent $^{40}Ar/^{39}Ar$ age peak at 1450 to 1600 Ma. Analyses of four samples across one 'IRD event' at 4.8 Ma reveal a main peak in $^{40}Ar/^{39}Ar$ ages around 600–450 Ma for all of the samples, regardless of whether they are in the 'background' signal, or whether they are enriched in IRD. Only the sample with the highest IRD concentration contains IRD with $^{40}Ar/^{39}Ar$ ages of 1450 to 1600 Ma (Fig. 6).

The dominant ~500 Ma age peak in the $^{40}Ar/^{39}Ar$ ages for all eight investigated IRD-rich layers at Site 1165 is consistent with a relatively local source of the IRD from the Lambert Glacier system draining into Prdyz Bay, and/or the coastal sector of Prydz Bay. Hence Site 1165 always received at least half of its IRD from nearby sources. For the IRD layers deposited prior to the mid-Miocene expansion of the East Antarctic Ice Sheet, the ~500 Ma age peak, from nearby sources, is the only clear peak in the age spectra (Fig. 5).

In contrast, for the IRD peaks after the mid-Miocene ice expansion we also find ages that are quite distinct from the prevailing Pan-African ages in the Prydz Bay area. The 1300–1100 Ma $^{40}Ar/^{39}Ar$ age peak found in the 3.5 and 7.0 Ma IRD-rich layers matches the core-top $^{40}Ar/^{39}Ar$ ages from the Wilkes Land margin (Figs. 4, 5), and the onshore thermochronology of Wilkes Land east of the Denman Glacier (100°E). This IRD is too old to be from the Rayner Complex exposed in the northern Prince Charles Mountains to the west of Prydz Bay (U–Pb zircon ages of 990–900 Ma, Boger et al., 2000), or from any isolated Grenvillian remnants younger than 1100 Ma in the Larsemann Hills (Tong et al., 2002). The geology at the head of the Lambert Glacier is relatively well exposed in the southern Prince Charles Mountains, and although a wide range of U–Pb thermochronological ages is represented, there is no particular age concentration at 1300–1100 Ma. Furthermore, $^{40}Ar/^{39}Ar$ results of Phillips et al. (2007) reveal a pervasive Pan-African metamorphic overprint in the southern Prince Charles Mountains. It is unlikely that this age population taps a hitherto unknown source in the continental interior, because glacial erosion is thought to occur predominantly where ice flow is fastest, close to the ice margins and under ice streams (Clark, 1987; Tulaczyk et al., 1998), in this case around the Lambert Graben and the Prydz Bay coast where the $^{40}Ar/^{39}Ar$ ages are relatively well defined (Fig. 4). Moreover, pre-glacial fluvial sediments from ODP Site 1166 in Prydz Bay, at the outlet from the Lambert drainage system, would be more likely to contain material eroded from further inland (e.g., the Gamburtsev Mountains), but show an age population tightly clustered around 505–512 Ma (213 hornblende grains analyzed, van de Flierdt et al., 2008).

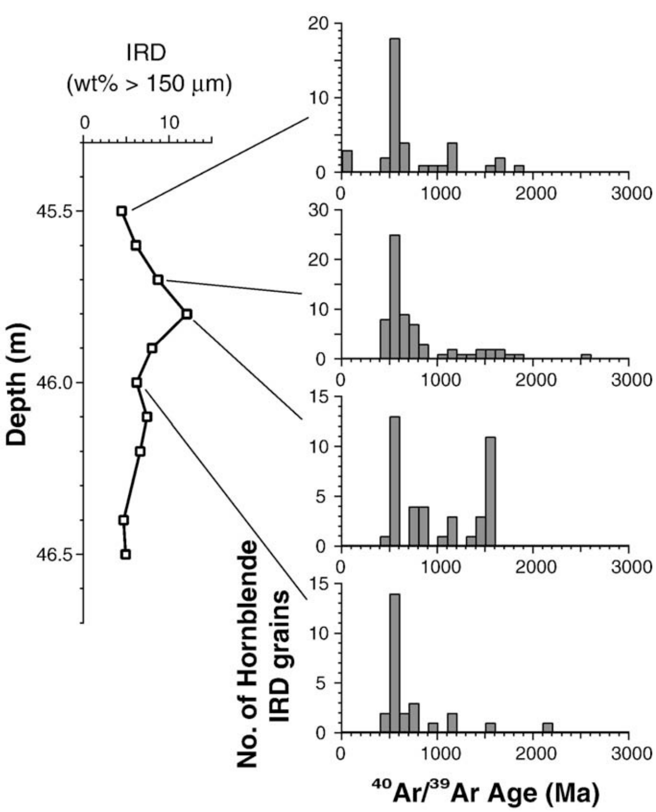

Fig. 6. $^{40}Ar/^{39}Ar$ age histograms of the 4.8 Ma IRD-rich layer centered at 45.8 m in Hole 1165A. IRD is represented here by weight % of the >150 µm grain size fraction. Note the abundance of IRD with $^{40}Ar/^{39}Ar$ age of 1500–1600 Ma in the middle of the IRD-rich layer, but its greatly reduced presence in the other samples.

Therefore we consider it highly unlikely that the Prydz Bay sector was the source for the 1300–1100 Ma IRD deposited 3.5 and 7.0 Ma at Site 1165. Prevailing westward currents exclude any continental source areas to the west of Site 1165 (Kuvaas and Leitchenkov, 1992; DeConto et al., 2007). The Wilkes Land coast is the only known area between Prydz Bay and the Ross Sea that matches the thermochronological ages of 1330–1130 Ma found at Site 1165 (Figs. 3–5) (Post et al., 1996; Möller et al., 2002; Fitzsimons, 2003; Roy et al., 2007). We therefore suggest that IRD with ages between 1330 and 1130 Ma came from the Wilkes Land coast.

Following the same line of reasoning, we can identify the source of the hornblende grains, deposited at 4.8 Ma at Site 1165 with 1550–1500 Ma $^{40}Ar/^{39}Ar$ ages to be the Adélie craton. Adélie Land is the only extensive onshore exposure of thermochronological ages this old to the east of Site 1165, with $^{40}Ar/^{39}Ar$ hornblende ages ranging from 1682 to 1710 Ma, and biotite $^{40}Ar/^{39}Ar$ ages from 1502 to 1687 Ma (Di Vincenzo et al., 2007; Duclaux et al., 2008). The Mertz shear zone was active at ~1550–1500 Ma (Duclaux et al., 2008), suggesting further localization of the source area for IRD grains of this age. An Adélie Land source for the 4.8 Ma IRD-layer is also supported by the offshore IRD survey, which has hornblende $^{40}Ar/^{39}Ar$ ages in the range 1750–1450 Ma (Roy et al., 2007, Figs. 4, 5). Data from the catchment inland from Prydz Bay do not show a major metamorphic event around this age.

A minor component of the IRD population at Site 1165 has $^{40}Ar/^{39}Ar$ ages of 900 to 600 Ma, but at this stage, we cannot assign it to a unique source. Potential source areas include the northern Prince Charles Mountains, where such ages could represent either partial Pan-African resetting of the Grenville-age signature, excess argon, very slow post-Grenville cooling, or intermediate-aged metamorphic events, such as a possible 700 Ma event (Tong et al., 2002).

The implication of the IRD source characterization, described above, for the IRD layers at ODP Site 1165 is that the site has been supplied by IRD originating from areas more than 1500 km (Wilkes Land) and more than 3000 km (Adélie Land) away during specific events in the late Miocene and Pliocene. Up to 50% of the IRD during the 7.0, 4.8, and 3.5 Myr ice-rafting events had a distant source. Such a high proportion of distant IRD is especially notable because much of the iceberg mass will melt in transit when traveling 1500 to 3000 km. Moreover, many of the icebergs originating from the Wilkes Land and Adélie Coast will have been deflected northwards on their journey — only a fraction will reach Site 1165 (Fig. 4) (Gladstone et al., 2001). Far-traveled IRD is rare in the core-top (Holocene) survey (Fig. 4). The data signify ice-rafting conditions much different from the present day, and point to massive ice discharge events of debris-rich icebergs from Wilkes Land and Adélie Land in the late Miocene and early Pliocene (as discussed in the next section).

An interesting observation is that these ice-rafting events also point to diachronous dynamics of different parts of the ice sheet margin. The 4.8 Ma IRD-rich layer contains IRD from Adélie Land but not from Wilkes Land, suggesting that different coastal sectors did not necessarily release IRD-bearing icebergs at the same time. If oceanic conditions were favorable for icebergs from Adélie Land to travel all the way to Site 1165, so could icebergs from Wilkes Land, closer to Site 1165. The fact that we do not observe a mixed provenance signature of Wilkes Land and Adélie Land IRD in the sampled downcore layers at Site 1165 argues against synchronous retreat of East Antarctic coastal sectors. The presence of far-traveled IRD from individual coastlines suggests rapid ice-rafting events, and provides evidence that one ice margin can undergo retreat while another may remain relatively stable.

6. Southern Ocean icebergs, ice-rafted debris, and potential parallels with North Atlantic Heinrich events

Present day icebergs in the Southern Ocean typically do not carry IRD far from the source. Large tabular icebergs are mainly produced from ice shelves and have low debris concentrations. Debris-rich basal ice from the grounded ice sheet typically melts off when the ice becomes an ice shelf, leaving only englacial and supra-glacial debris in the icebergs when they calve. Although the larger icebergs can travel long distances before breaking up, their low levels of incorporated debris means they transport little IRD far from the calving site (Warnke, 1970). Tidewater glaciers on the other hand contain more concentrated debris (e.g. Powell, 1984; Dowdeswell and Murray, 1990), but icebergs generated by them tend to be smaller and generally do not transport debris far from the source.

After they calve from the main ice mass, icebergs drift westwards around the East Antarctic coast in the polar current (Fig. 3). This motion is observed in satellite images (Tchernia and Jeannin, 1984; Long et al., 2002) and in iceberg drift models (Gladstone et al., 2001; Silva et al., 2006). At locations where bathymetric highs and/or the wind patterns cause a northward component to the polar current, such as the Kerguelen Plateau, ~85°E, some icebergs drift north, and become entrained in the eastward-flowing Antarctic Circumpolar current (Fig. 3). More massive icebergs are more likely than small icebergs to travel further along the coast before deflecting northwards, because the Coriolis effect deflects icebergs to the left in the southern hemisphere, and acts more strongly on more massive objects. Hardly any of the giant icebergs (250 m thick, more than 18.5 km long) modeled by Silva et al. (2006), are deflected north, while many of the small icebergs (250 m thick, length less than 2.2 km) modeled by Gladstone et al. (2001) are. Icebergs large enough to be tracked by satellite tend to remain in the polar current and few are deflected northwards (Tchernia and Jeannin, 1984; Long et al., 2002).

In summary, in the present day, icebergs are transported westwards around the East Antarctic coast, but deposit the greater part of their IRD close to the source. This observation is in good agreement with the geochemical provenance analyses of modern IRD, which is dominated by material from the adjacent coast (Roy et al., 2007). Different conditions, in terms of iceberg production, debris content, and/or transport, must have prevailed in the past to produce the observed high proportion of far-traveled IRD in the late Miocene–Pliocene samples at Site 1165.

Evidence that ice sheet and iceberg conditions were different in the past includes deep (>500 m) iceberg scour marks on the sea bed. Such scour marks result from icebergs from fast-flowing ice outlets, or a pulse of large icebergs released during major deglaciation or ice sheet collapse (Dowdeswell and Bamber, 2007). O'Cofaigh et al. (2008) characterized rates of retreat of paleo ice streams based on the bedforms they leave. Rapid retreat leaves few or no recessional moraines or grounding zone wedges, and icebergs produced in this way may be expected to retain more debris than from episodic or slow retreat, where debris is left on the shelf. Both these studies imply that ice stream destabilization events can produce large volumes of icebergs, which are may exceed 500 m thick, and carry larger amounts of detritus than seen in today's icebergs. From glacial deposits in the Lambert Graben, Hambrey and McKelvey (2000) infer greater numbers of debris-bearing icebergs during rapid ice stream recession.

There are parallels between the late Miocene to Pliocene Southern Ocean ice-rafting events, described here, with late Pleistocene North Atlantic Heinrich events. Heinrich events are characterized by short-lived but massive discharges of debris-rich icebergs into the North Atlantic (Hemming, 2004). Binge/purge cycles of the ice sheet (e.g. MacAyeal, 1993) and jökulhlaup (outburst flooding) activity (Johnson and Lauritzen, 1995; Roberts, 2005; Alley et al., 2006) have been proposed to cause ice stream surges and generate large masses of icebergs. It has been a puzzle how to incorporate enough debris in the icebergs to produce the observed thicknesses of debris layers across the North Atlantic. Proposed methods include injection of dirty subglacial floodwater up into the ice sheet (Roberts, 2005), and freezing-on of the debris-laden ice at the bed of the ice stream during surges (Alley and MacAyeal, 1994). Elements of these models may play a part in the generation of debris-bearing icebergs that caused the IRD layers at Site 1165. Additionally, a low-lying sub-glacial basin is a prerequisite for ice destabilization in the Heinrich models, and in the East Antarctic case, the Wilkes and Aurora sub-glacial basins (Fig. 1) would be notable candidates. Thus the high proportion of far-traveled IRD in the IRD layers at Site 1165 is consistent with independent evidence from both Antarctica and the Northern Hemisphere of relatively rapid ice destabilization that produces large IRD-bearing icebergs.

In terms of the iceberg drift paths in the past, the upstream-migrating sediment waves in the area of Site 1165 attest to westward bottom-water flow since at least the early Miocene (Kuvaas and Leitchenkov, 1992). In models where there is a moderately reduced ice sheet volume, wind directions are similar to today's, blowing westward along the coast (DeConto et al., 2007). The polar current is wind-driven, so the iceberg drift tracks also would be expected to be similar to the present. It takes a substantial reduction in ice sheet topography, to levels probably not seen since the mid-Miocene transition, to change the wind patterns significantly, to give weaker wind strengths and mixed wind directions at the coast (DeConto et al., 2007).

An overall reduced ice volume before the mid-Miocene climate transition is our favored explanation for the lack of far-traveled debris in the five sampled IRD layers older than 14 Ma. This would have a twofold effect on ice rafting. Firstly, fewer IRD-bearing icebergs would be produced from Wilkes Land and Adélie Land because there needs to be a sufficient amount of ice on the low-lying parts of the continent to enable ice stream surge and iceberg discharge. Secondly, the reduced coastal wind strength and non-uniform wind direction would reduce the strength of the polar current, so that icebergs would be dispersed more widely and fewer would be transported to Site 1165. This interpretation of reduced ice volume before the mid-Miocene is in agreement with the documented reduction in ice volume from marine oxygen isotope records (Lear et al., 2000; Zachos et al., 2001; Billups and Schrag, 2003; Shevenell et al., 2008) and matches our finding that pre-mid-Miocene IRD at Site 1165 are sourced from the Prydz Bay area.

The timing of IRD deposition relative to warm-cold climatic cycles is important in terms of understanding the processes by which the IRD-carrying icebergs were produced. For example, IRD-bearing icebergs can be produced when an ice sheet reaches the edge of the continental shelf during glacial maxima (Marshall and Koutnik, 2006);

during glacial advance (McManus et al., 1999), during glacial retreat (McManus et al., 1999), or due to instabilities in the ice sheet leading to ice rafting like the Heinrich events of the North Atlantic in the last glacial (Hemming, 2004). In contrast to Heinrich events, which coincide with times of unusually cold temperatures in the North Atlantic, the ice-rafting events described here occur under warming conditions (Grützner et al., 2003; Williams and Handwerger, 2005), suggesting a key role for rising temperatures in destabilizing ice streams, in turn leading to iceberg generation. Two of the ice-rafting events, at 3.5 Ma and 4.8 Ma, occurred when the West Antarctic ice sheet was much reduced, as recorded in the ANDRILL sediment record under the McMurdo ice shelf (Naish et al., 2009), and when sea ice at Site 1165 was reduced by approximately 55% relative to the present day (Whitehead et al., 2005). Our results indicate that the Wilkes Land and Adélie Land ice margins were also unstable at these times.

Our data agree with the view of an East Antarctic Ice Sheet whose dynamic margins have retreated several hundred kms at times in the late Miocene and Pliocene, as Hambrey and McKelvey (2000), and Whitehead et al. (2006) have shown for the Prydz Bay area. Our data imply similar dynamism for the Wilkes and Adélie ice margins, and moreover that the ice retreats were probably rapid. The results from Site 1165 are compatible with the ice sheet model of Hill et al. (2007), which shows a maximum retreat of several hundred kms for Pliocene climate conditions. On the other hand, DeConto and Pollard's (2009) ice sheet model suggests that even under the warmest Pliocene conditions, the Wilkes and Adélie ice margins remain relatively stable, with East Antarctic ice as a whole contributing only about 2 m equivalent of sea-level rise relative to today. Our data imply that the ice sheet is more dynamic than this model suggests, and highlights the known issue that despite recent advances, ice sheet models do not yet fully represent the physics of the ice streams and ice margins (Alley et al., 2005; Pollard and DeConto, 2007). More IRD provenance data, in tandem with continued ice sheet modeling, will advance the understanding of the frequency, rate, and magnitude of the ice margin retreats described in this paper.

7. Conclusions

We presented downcore $^{40}Ar/^{39}Ar$ data on individual hornblende grains from eight IRD-rich layers at ODP Site 1165 (Prydz Bay) covering the time interval from 3.5 to 19 Ma. When compared to the $^{40}Ar/^{39}Ar$ ages of continental source areas, the results reveal that IRD-rich layers after the mid-Miocene climate transition show a provenance that it is hard to reconcile with local sources alone. About half the IRD comes from the Wilkes Land coast or the Adélie Coast, 1500 km and 3000 km away from the site of IRD deposition, implying massive discharges of debris-rich icebergs from the Wilkes and Adélie coasts. Such long travel paths of debris-rich icebergs are not observed in the present day Southern Ocean, and imply different ice sheet dynamics at times in the late Miocene and Pliocene. Our data are consistent with Heinrich-like events supplying armadas of icebergs to the Southern Ocean. In contrast to Heinrich events however, the iceberg production seems to have occurred in warming phases, suggesting surges of large, IRD-bearing icebergs in times of rapid ice retreat. In order to fully understand the underlying mechanisms and the extent of ice break-up, and consequent release of iceberg armadas to the Southern Ocean, future studies on marine sediment cores off East Antarctica as well as coupling of these results to ice sheet models will be vital.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl.2009.12.031.

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