63. CALCAREOUS NANNOFOSSIL STRATIGRAPHY AND REASSESSMENT OF THE EOCENE GLACIAL RECORD IN SUBANTARCTIC PISTON CORES OF THE SOUTHEAST PACIFIC[1]

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ABSTRACT

This study provides the first detailed documentation of calcareous nannofossil assemblages in Eocene subantarctic *Eltanin* piston cores recovered from the southeast Pacific Ocean. These *Eltanin* cores are important because they have been reported to contain ice-rafted quartz. The present study confirms early and middle Eocene ages for the cores and dates them more precisely using calcareous nannofossils. Semiquantitative study of the nannoflora indicates that they are of a warmer water character than those from the higher latitudes (such as Falkland Plateau, Maud Rise, and Kerguelen Plateau). This study concludes that it is unlikely for the ice-rafted quartz in the Eocene sections to be downcore contaminants from the overlying Neogene sediment and suggests that the grains are probably the result of Eocene ice-rafting from Antarctica when the Drake Passage was closed and water circulation patterns were different from those of today.

INTRODUCTION

One of the major objectives of Ocean Drilling Program (ODP) Leg 120 was to study the paleoceanographic history of the Indian Ocean sector of the Southern Ocean, including the shift of the polar front, and the initiation and development of Circumpolar and Antarctic Bottom Water circulation (Schlich, Wise, et al., 1989), all of which are directly related to the development of the Antarctic ice sheet. During Leg 120 to the Central Kerguelen Plateau, abundant lower Oligocene ice-rafted debris (IRD) was recovered at Site 748 (Fig. 1; Breza and Wise, this volume). Coincident with the IRD is a sharp positive excursion in foraminifer $\delta^{18}O$ values at Site 748 and other sites (Zachos et al., this volume). This led Breza and Wise (this volume) to postulate the existence of an earliest Oligocene ice sheet on the Antarctic continent.

Previously, ODP Leg 113 drilling at Site 693 on the East Antarctic margin in the Weddell Sea (Fig. 1) recovered ice-rafted sands and dropstones within a sequence of lower Oligocene diatomaceous muds dated at 33 Ma (Barker, Kennett, et al., 1988, 1990). Abundant IRD was also found at ODP Sites 738 and 744 on the Southern Kerguelen Plateau (Barron, Larsen, et al., 1989, 1991; Ehrmann, 1991). Extensive lowest Oligocene glacial diamictites were recovered at the CIROS-1 site in the Ross Sea (Barrett, 1989; Wei, this volume) and in Prydz Bay, East Antarctica (Fig. 1; Barron, Larsen, et al., 1989, 1991). All these recent discoveries suggest the existence of an ice sheet on Antarctica during the earliest Oligocene, pushing the age back some 20 m.y. earlier than the long-held hypothesis that the onset of the East Antarctic ice sheet was in the middle Miocene (Shackleton and Kennett, 1975; Savin et al., 1975; Kennett, 1986).

Although no complete consensus has been reached, most investigators believe that an ice sheet was present on East Antarctica during the Oligocene, including the earliest Oligocene (Wise et al., 1991). A major question now is whether an ice sheet could have been present there during the Eocene. Even before Deep Sea Drilling Project (DSDP) drilling in the Southern Ocean, there were reports of ice-rafting during the Eocene. Margolis and Kennett (1971) reported Eocene ice-rafted quartz grains in several *Eltanin* piston cores taken in the subantarctic region of the Pacific Ocean. This ice-rafted debris remains the oldest ever reported for the Cenozoic. Currently, many workers do not accept these quartz particles as Eocene ice-rafted grains, mostly because there is very little other evidence available now that suggests significant ice on Antarctica during the early and middle Eocene. These quartz grains were identified as ice-rafted by their surface texture and were dated as Eocene on the basis of planktonic foraminifers before biozonation schemes had been established for the early and middle Cenozoic of the subantarctic-antarctic regions (Margolis and Kennett, 1971). It is, therefore, important to redate these cores using calcareous nannofossils, as suggested recently by S. Margolis (J. Barron, written comm., 1989), and to reassess the Eocene glacial record.

The present study was undertaken to examine and document the calcareous nannofossil assemblages in *Eltanin* Cores 13-4, 24-8, 24-9, and 24-10, which Margolis and Kennett (1971) dated as Eocene and from which they reported the ice-rafted quartz. In addition, an effort was made to determine whether there is any significant core mixing that may have brought the younger ice-rafted quartz into older core sections. This study should help date the cores more precisely and clarify some of the questions concerning contamination.

Except for Geitzenauer et al. (1968), who recorded four Eocene nannofossil species in Core 13-4, no studies have been done on Eocene calcareous nannofossils from the other cores or from the southeast Pacific region. The detailed documentation of the calcareous nannofossils in the present paper should provide a useful reference in the southeast Pacific Ocean for future biostratigraphic, paleobiogeographic, and paleoecologic studies.

MATERIALS AND METHODS

The location and water-depth data of the four *Eltanin* piston cores investigated are given in Table 1. Core 13-4 was taken off southwest South America, whereas Cores 24-8, 24-9, and 24-10 are located in the South Pacific (Fig. 1), an area north of the present-day iceberg limit. Backtracking of the plate motion indicates that these core locations were no more than a few degrees south of their present positions during the Eocene (Margolis and Kennett, 1971).

[1] Wise, S. W., Jr., Schlich, R., et al., 1992. *Proc. ODP, Sci. Results*, 120: College Station, TX (Ocean Drilling Program).

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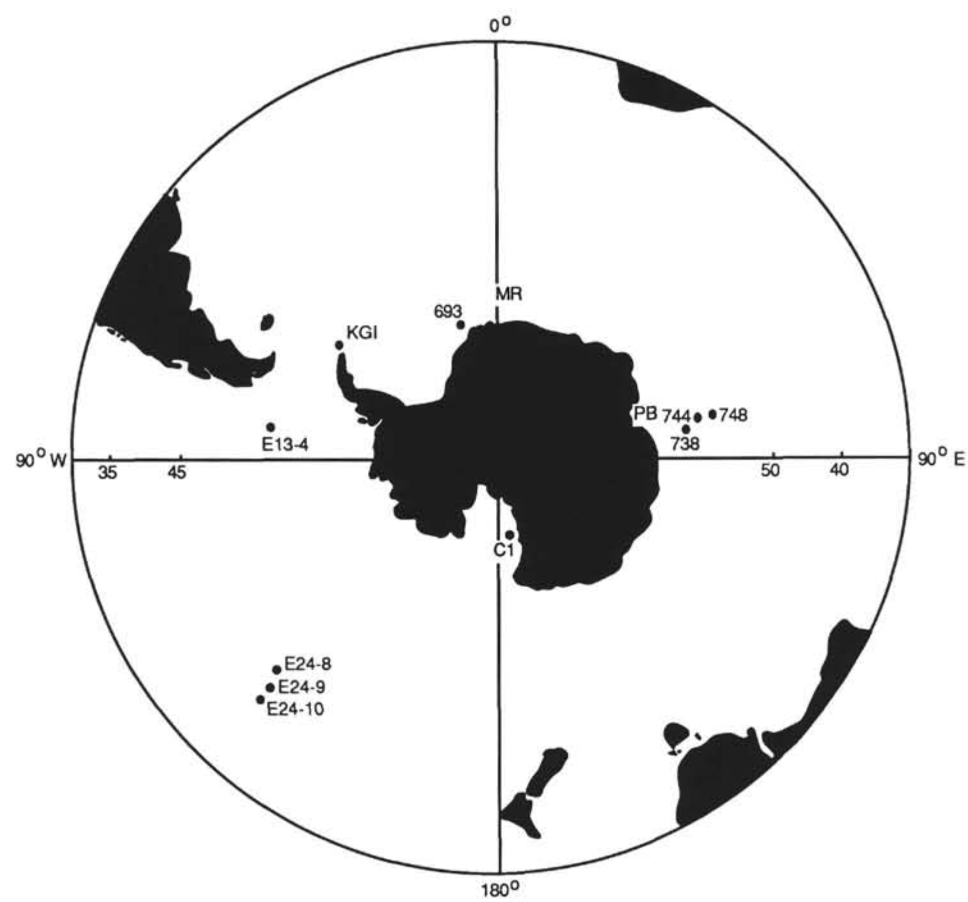


Figure 1. Location map of *Eltanin* piston Cores 13-4, 24-8, 24-9, and 24-10. Other sites discussed in the paper are also shown. C1 = CIROS-1, KGI = King George Island, MR = Maud Rise, and PB = Prydz Bay.

Table 1. Location and water depth data of the *Eltanin* piston cores studied.

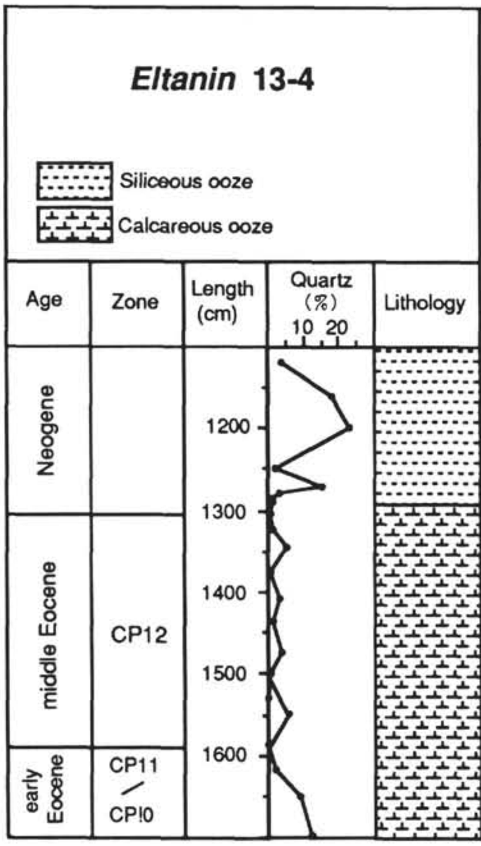
Core	Latitude	Longitude	Water depth (m)
13-4	57°46′S	90°48′W	4700
24-8	42°53′S	134°39′W	5011
24-9	40°35′S	135°16′W	4837
24-10	37°58′S	134°59′W	4874

Note: Data given here supersede those given in Margolis and Kennett (1971, table 1).

The upper 13 m of Core 13-4 is Neogene siliceous ooze devoid of calcareous microfossils, and ice-rafted quartz constitutes about 15% of the >63-μm grains in this section (Fig. 2). The lower 4 m of the sequence is clayey calcareous ooze, and there is even less ice-rafted quartz (about 5%). The lithology of Core 24-8 (Fig. 3) is similar to that of Core 13-4 in that it consists of a Neogene siliceous ooze section in the upper part of the core and a Paleogene calcareous ooze section in the lower part . There is strong indication of core flow-in below 5.4 m, so Margolis and Kennett (1971) did not examine quartz grains below this level. One sample from the Paleogene section of Core 24-8 yielded 10% of ice-rafted quartz, whereas most samples in the Neogene contained only about 5% ice-rafted quartz (Margolis and Kennett, 1971).

The lithology of Core 24-9 is given in Figure 4. The percentage of ice-rafted quartz increases upward in both the Paleogene and Neogene sections. Core 24-10 consists of only Paleogene calcareous ooze. Margolis and Kennett (1971) reported 1%–10% ice-rafted quartz in the five samples they examined from this core (Fig. 5).

Smear slides were made directly from unprocessed samples taken from the cores and examined using a light microscope at ×650 magnification. The abundance of calcareous nannofossils on each slide was estimated according to the following criteria: V = very abundant (>10 specimens per field of view); A = abundant (1–10 specimens per field of view); C = common (1 specimen per 2–10 fields of view); F = few (1 specimen per 11–50 fields of view); R = rare (1 specimen per 51–200 fields of view); and B = barren (no specimen was found in 200 fields of view). Preservation of the calcareous nannofossil assemblage is recorded as G = good (little evidence of etching or overgrowth); M = moderate (etching or overgrowth is apparent); and P = poor (there is significant etching or overgrowth and identification of some species is impaired). A complete list of the species considered in this paper is presented in the Appendix, and bibliographic references for these taxa can be found in Loeblich and Tappan (1966–1973), van Heck (1979–1983), or Steinmetz (1985–1989). All species identified in this study are documented with scanning electron microscope (SEM) and light micrographs in Plates 1–3. Calcareous nannofossil zones are given according to the zonation of Okada and Bukry (1980).



Calcareous nannofossil distribution

Sample depth (cm)	Abundance	Preservation	*Chiasmolithus eograndis*	*Chiasmolithus expansus*	*Chiasmolithus grandis*	*Chiasmolithus solitus*	*Coccolithus formosus*	*Coccolithus pelagicus*	*Discoaster barbadiensis*	*Discoaster kuepperi*	*Discoaster lodoensis*	*Discoaster saipanensis*	*Discoaster* spp.	*Discoaster sublodoensis*	*Helicosphaera* sp.	*Markalius inversus*	*Neococcolithes dubius*	*Reticulofenestra samodurovii*	*Sphenolithus moriformis*	*Toweius magnicrassus*	*Tribrachiatus orthostylus*
1210	B																				
1470	V	M	R	R	F	F	C	A		R	C	F	C	R			F	C	R	R	
1543	V	M	R		C	F	C	A		A	C		C	R			F			C	
1666	V	M			F	F	F	A	F	C	C		F		R	F	F		R		C
1698	A	G	R		C		F	A	F	C	C		F			F				R	C

Figure 2. Lithology, percentage of quartz and calcareous nannofossil biostratigraphy of *Eltanin* piston Core 13-4. The lithology and percentage of quartz are taken from Margolis and Kennett (1971). The abundance of nannofossils is characterized by V = very abundant, A = abundant, C = common, F = few, R = rare, and B = barren. For preservation, G = good and M = moderate.

RESULTS

Core 13-4

The distribution of calcareous nannofossils in Core 13-4 is presented in Figure 2. Nannofossils are very abundant and moderately well preserved below 1300 cm. Because of the high carbonate content, overgrowth of the nannofossils is more prominent than dissolution. Many six-rayed discoasters could not be differentiated at the species level. The very bottom section of the core contains more clay, and nannofossil preservation is good. This stratigraphic level coincides with the highest percentage (12%) of ice-rafted quartz found in the Paleogene section by Margolis and Kennett (1971).

The most age-diagnostic species in this core include *Discoaster kuepperi, Discoaster lodoensis, Discoaster sublodoensis,* and *Tribrachiatus orthostylus*. Common *D. kuepperi* and *D. lodoensis* and the absence of *D. sublodoensis* in the lowest two samples permit an age assignment of nannofossil Zones CP10–CP11. The last occurrence (LO) of *T. orthostylus* was observed at 1666 cm, where the species was common. This datum is conventionally used to define the top of Zone NP12 of Martini (1971), which is equivalent to the top of Zone CP10; however, the reliability of the LO of *T. orthostylus* has been questioned because the species seems to occur too high in some sections and sometimes co-occurs with *Nannotetrina* (Perch-Nielsen, 1985; Wei and Wise, 1989). The datum is, therefore, not applied for the top of Zone CP10 in this study.

The first occurrence (FO) of *D. sublodoensis* was found in Sample 1543. This defines the lower boundary of Zone CP12. Consequently, Samples 1470 and 1543 can be placed in Zone CP12 (early middle Eocene) based on the co-occurrence of *D. sublodoensis* with *D. kuepperi* and *D. lodoensis*. The results of this study agree with those of the initial study on this core by Geitzenauer et al. (1968), who recorded several age-diagnostic species and suggested a minimum age of middle Eocene for the sediment below 1296 cm.

Core 24-8

Two samples from the short calcareous ooze section in Core 24-8 were examined, and they yielded abundant to very abundant nannofossils (Fig. 3). *Discoaster lodoensis* is present in both samples, but its abundance is greatly reduced relative to that in Core 13-4. Rare *Discoaster sublodoensis* were found in Sample 487. The presence of this species suggests an age of CP12 or younger. The ubiquitous *Reticulofenestra umbilica,* the FO of which defines the bottom of Zone CP14, does not occur at this level. This indicates an age of CP13 or older. The presence of other species such as *Chiasmolithus grandis* and *Chiasmolithus solitus* are consistent with an age of CP12–CP13. Furthermore, *Discoaster kuepperi,* which is common to abundant in Core 13-4 taken at a higher latitude, is not present in Core 24-8. Therefore, the calcareous ooze section of Core 24-8 can be assigned to an undifferentiated CP12b–CP13 Zone (early middle Eocene). Further biostratigraphic subdivision is not possible because neither *Rhabdosphaera inflata, Nannotetrina fulgens,* nor *Chiasmolithus gigas* was found. The absence of these species is probably a result of enhanced dissolution in the deep water (present water depth = 5011 m). One specimen of *Isthmo-*

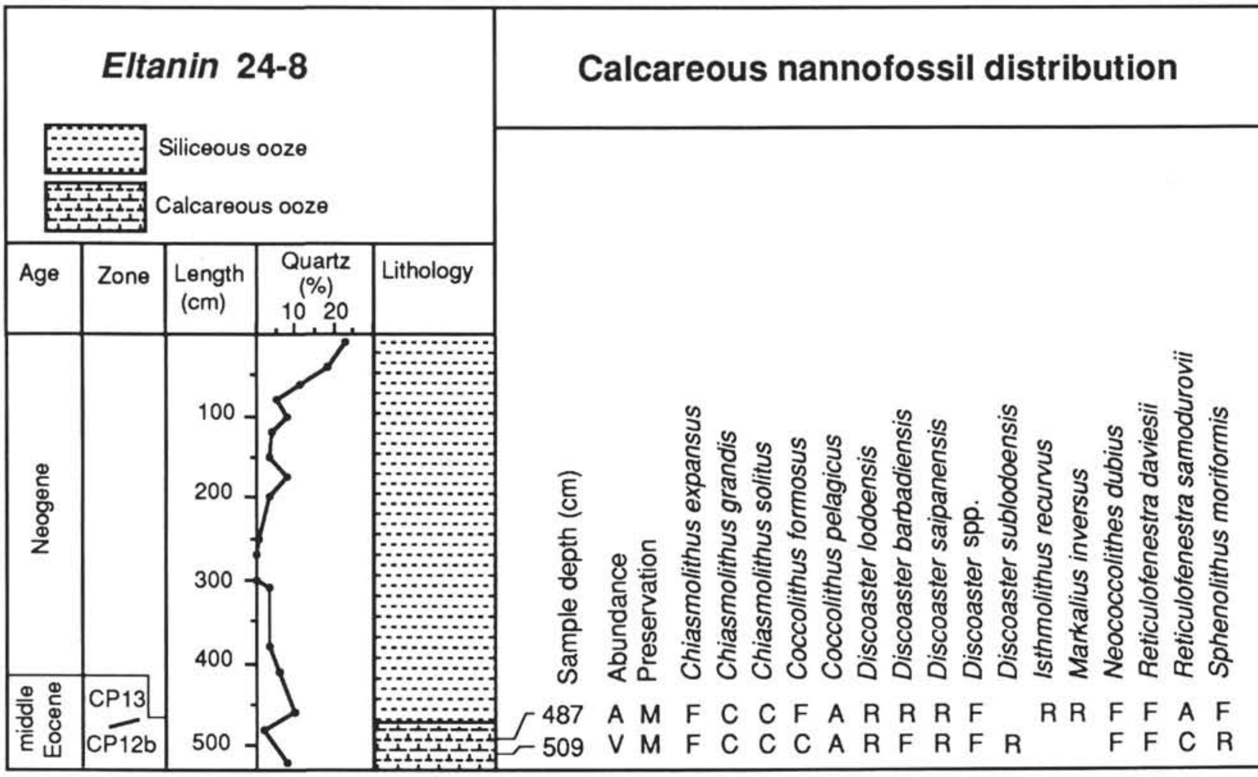


Figure 3. Lithology, percentage of quartz, and calcareous nannofossil biostratigraphy of *Eltanin* piston Core 24-8. The lithology and percentage of quartz are taken from Margolis and Kennett (1971). The abundance of nannofossils is characterized by V = very abundant, A = abundant, C = common, F = few, and R = rare. For preservation, M = moderate.

lithus recurvus was observed in Sample 487 (Pl. 2, Fig. 11). As *I. recurvus* normally ranges from Zones CP15 to CP16, this single specimen is believed to be a contaminant. The initial study of Margolis and Kennett (1971) assigned an early Eocene age to the calcareous section. This age is now revised to early middle Eocene.

Core 24-9

Calcareous nannofossils are very abundant but poorly preserved (overgrown and fragmented) in the calcareous ooze section of Core 24-9 (Fig. 4). The most age-diagnostic species present include *Chiasmolithus expansus, Chiasmolithus grandis,* and *Reticulofenestra samodurovii*. The presence of these species with the absence of *Discoaster lodoensis* and *Discoaster kuepperi* suggests Subzone CP12b or younger, and the absence of *Reticulofenestra umbilica* indicates Zone CP13 or older. The calcareous ooze section can, therefore, be placed in a combined CP12b–CP13 Zone (early middle Eocene). This assignment is only slightly older than the late middle Eocene age given by Margolis and Kennett (1971). Two specimens of *Tribrachiatus orthostylus* were found in Sample 488. The abnormal stratigraphic occurrence of *T. orthostylus* in this core could be the result of reworking; alternatively, this may be an example of the case discussed above in which *T. orthostylus* is thought to range higher in some sections.

Core 24-10

Four samples were taken from this 230-cm-long core and examined for calcareous nannofossils. The distribution of the calcareous nannofossils is given in Figure 5. The preservation of the nannoflora is poor to moderate, as most chiasmoliths are missing their cross-bars and discoasters are overgrown and generally nondiagnostic at the species level. However, *Reticulofenestra umbilica* (common) and *Chiasmolithus solitus* (few) are present in all the four samples examined. The co-occurrence of these two species clearly places this core into Subzone CP14a (late middle Eocene). The presence of other age-diagnostic species, (i.e., *Chiasmolithus expansus, Chiasmolithus grandis,* and *Reticulofenestra bisecta*), are consistent with a late middle Eocene age. This age assignment agrees with that of Margolis and Kennett (1971).

DISCUSSION

Margolis and Kennett (1971) examined the species diversity of foraminifers in the *Eltanin* cores and found fairly low diversity. They concluded that cool Southern Ocean conditions prevailed throughout much of the Cenozoic. They also noted that low species diversity in the cores generally was associated with relatively high percentages of quartz, and conversely higher species diversity was associated with sediment that contained very little quartz. This relationship is not apparent, however, from the nannofossil data. Compared with nannofossil assemblages from the higher latitudes, such as the Falkland Plateau (Wise, 1983), Maud Rise (Pospichal and Wise, 1990; Wei and Wise, 1990), and the Kerguelen Plateau (Wei and Thierstein, 1991; Wei et al., this volume), the nannofossil assemblages in the *Eltanin* cores show warmer water characteristics, because they contain more discoasters (warm-water indicators) and fewer *Toweius* and *Reticulofenestra* (cool-water indicators). The low species diversity in the *Eltanin* cores is probably a result of poor preservation of the nannofossils.

As has been pointed out by Mercer (1973, 1978), the calving of outlet glaciers from mountain ice fields provide bergs too small to travel far from their source; only ice sheets can produce bergs large enough to survive a journey across

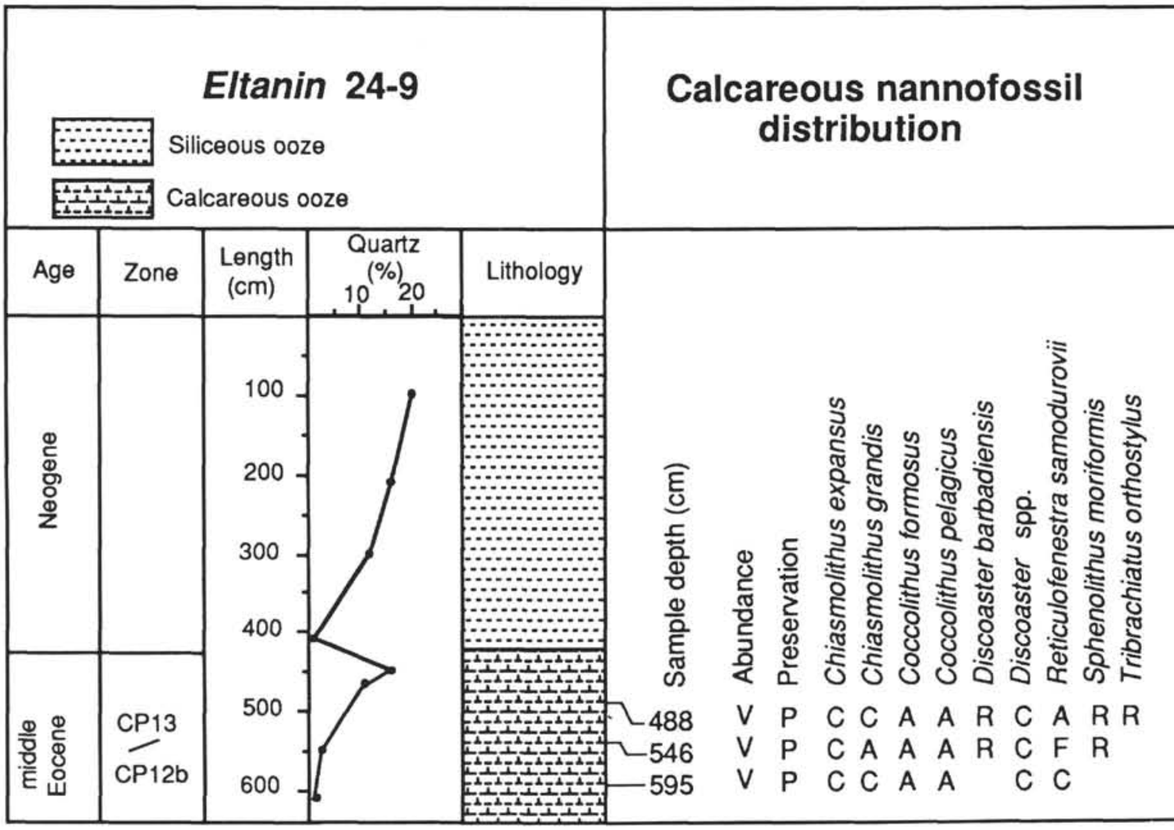


Figure 4. Lithology, percentage of quartz and calcareous nannofossil biostratigraphy of *Eltanin* piston Core 24-9. The lithology and percentage of quartz are taken from Margolis and Kennett (1971). The abundance of nannofossils is characterized by V = very abundant, A = abundant, C = common, F = few, and R = rare. For preservation, P = poor.

several degrees of latitude. The early and middle Eocene ice-rafted quartz reported by Margolis and Kennett (1971) would suggest that very large glaciers, perhaps from a small ice sheet, extended to the coast of Antarctica during this time. Although this scenario has generally not been substantiated by other data, a few studies have argued for this possibility.

Birkenmajer et al. (1986) and Birkenmajer (1987) reported a K-Ar date of 49.4 ± 5 Ma (early-middle Eocene) from a basaltic lava flow capping marine, fossiliferous tillite in the area of the Krakow Icefield on King George Island of West Antarctica. The tillite contains reworked Cretaceous and Paleogene calcareous nannofossils, the youngest of which are early Eocene in age according to Birkenmajer et al. (1988). This would suggest that there was glacial activity on West Antarctica during the early-middle Eocene. Birkenmajer (1987) also found lower Oligocene and lower Miocene tillites and glaciomarine sediments, indicating multiglacial events thought to be Antarctic wide.

Matthews and Poore (1980) argued on the basis of their interpretation of oxygen isotopic data that an Antarctic ice sheet had formed by at least the earliest Oligocene and probably as early as the Cretaceous. This view was developed further in Poore and Matthews (1984) and Prentice and Matthews (1988). Rapid fluctuations of the sea-level curves of Haq et al. (1987) also suggest the existence of an ice sheet on Antarctica during Eocene time.

However, unequivocal Oligocene ice-rafted material has been found only in drill cores close to the Antarctic continent in the Ross Sea (upper Oligocene; Hayes, Frakes, et al., 1975), in the Weddell Sea (lower Oligocene; Barker, Kennett, et al., 1988, 1990), and in sediment from the Kerguelen Plateau in the southern Indian Ocean (lower Oligocene; Barron, Larsen, et al., 1989, 1991; Schlich, Wise, et al., 1989; Breza and Wise, this volume). Lower Oligocene diamictites were found in McMurdo Sound in the Ross Sea (Barrett, 1989) and in Prydz Bay, East Antarctica (Barron, Larsen, et al., 1989, 1991). Below the lower Oligocene diamictites in Prydz Bay are additional diamictites, which could be as old as the latest middle Eocene (41 Ma) according to the magnetostratigraphy of Barron et al. (1991). However, none of these intensive drilling activities around the Antarctic continent has produced evidence for early Eocene glaciation on Antarctica, and the middle Eocene glacial evidence is not unequivocal.

Because the four *Eltanin* cores have now been dated independently by calcareous nannofossils and confirmed as early Eocene and middle Eocene in age, what possible explanations can be suggested for the presumed Eocene ice-rafted quartz reported? There are three main possibilities.

The first possibility is that the quartz resulted from downcore contamination caused by flow-in during coring or subsequent core handling. Special efforts were made during this study to determine whether there had been intensive core mixing by examining the cores and by investigating the samples using the light microscope. The Neogene sections of the cores are all of darker color (gray brown) than the Eocene sections, which are nearly white. No patches of darker sediments (Neogene diatom ooze) within the white Eocene calcareous ooze are apparent. If extensive core mixing had occurred, Neogene diatoms should have been abundant in at least some of the slides made from the Eocene sections, but virtually no diatoms were found. It is, however, possible that

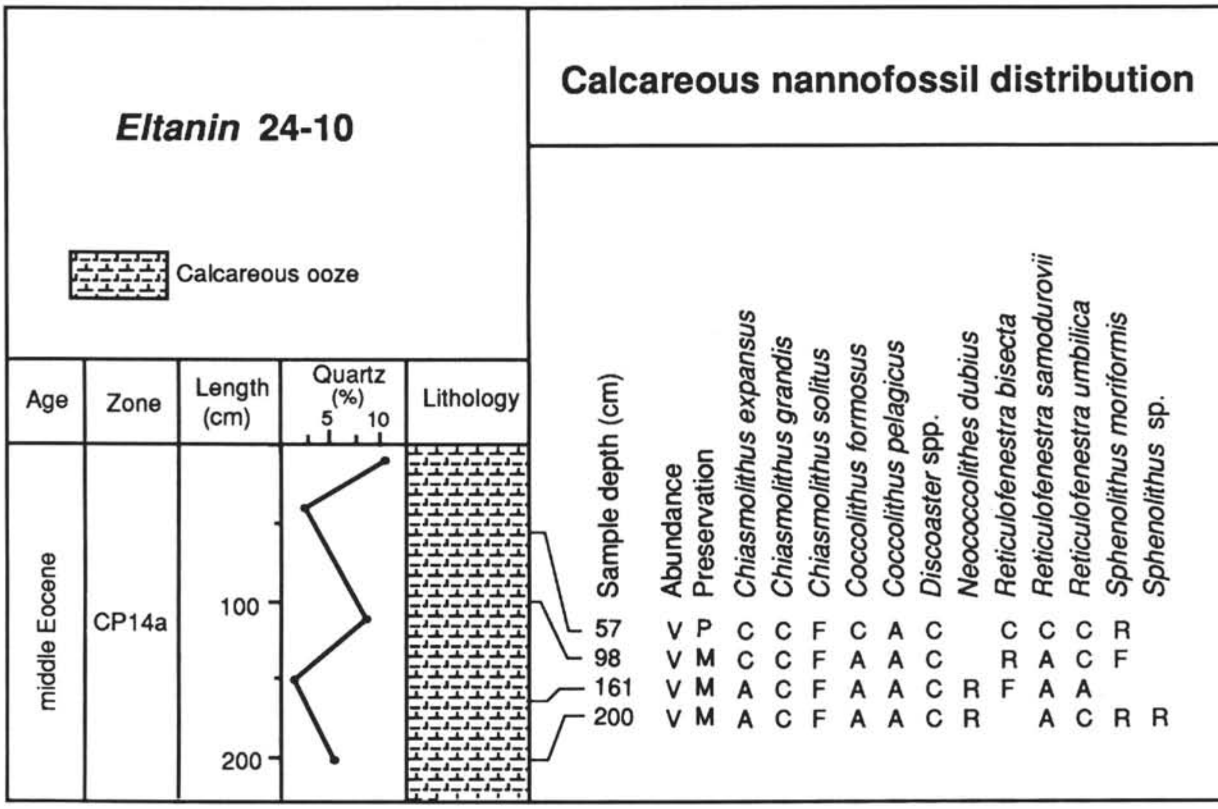


Figure 5. Lithology, percentage of quartz and calcareous nannofossil biostratigraphy of *Eltanin* piston Core 24-10. The lithology and percentage of quartz are taken from Margolis and Kennett (1971). The abundance of nannofossils is characterized by V = very abundant, A = abundant, C = common, F = few, and R = rare. For preservation, M = moderate and P = poor.

some Neogene sediment might have been smeared down core along the plastic liner and distributed along the side of the cores. Any such contaminants, however, are unlikely to constitute a significant proportion of the 8-cm^3-size samples taken by Margolis and Kennett (1971), which have yielded percentages of quartz grains in the Eocene sections comparable with those in the Neogene sections. In fact, the reported percentages of quartz grains are rather high at many levels in the Eocene sequences and are even higher than at some levels in the Neogene sequences. It is inconceivable that mixing could create this kind of distribution pattern among the quartz grains.

It also seems impossible that the mixing of Neogene into Paleogene sediment could occur in all the cores and at all levels that Margolis and Kennett (1971) sampled for quartz-grain analysis without being detected by Geitzenauer et al. (1968), Margolis and Kennett (1971), and this study. Moreover, in the case of Core 24-10, the whole core consists of Eocene calcareous ooze, and there is no overlying Neogene sediment to mix with it. That is, the ice-rafted grains cannot come from Neogene sediments but must be *in situ* within the Eocene. Furthermore, Geitzenauer et al. (1968) noted that in Core 13-4, quartz grains from the Neogene section show fresher surfaces with less evidence of dissolution than those grains in the Eocene section. This would further indicate that the quartz grains in the Eocene sediment were not brought down from the Neogene section by core mixing. Based on the above discussion, it is concluded that the Eocene quartz grains are quite unlikely to be downcore contaminants.

The second possibility is that those quartz grains are not ice-rafted. Margolis (1975) pointed out that surface features on sanidine and volcanic glass grains can closely resemble glacial quartz. Those Eocene grains reported by Margolis and Kennett (1971), however, have been more recently analyzed by energy dispersive X-ray and have been confirmed as quartz (S. Margolis, pers. comm., 1990). Transportation of the quartz grains by wind or turbidity currents is another possibility, but considering the remoteness of land or continental shelf from all these core locations (>1000 km), especially for Cores 24-8, 24-9, and 24-10, and the large grain size (mostly larger than 100 μm), this possibility is extremely small. If one supposes that these quartz grains are of volcanic origin, or were transported by wind, turbidity currents or other agents, why are quartz grains with similar characteristics not found in Eocene or older sediments elsewhere since the criteria used by Margolis and Kennett (1971) to identify ice-rafted quartz are still widely used today?

The third possibility is that these Eocene quartz grains are indeed ice-rafted. If so, the presence of ice-rafted quartz in these subantarctic *Eltanin* cores but not in cores from other sectors of the Southern Ocean would suggest that the *Eltanin* core locations were on iceberg tracks during the Eocene. With Drake Passage probably closed during the Eocene, water circulation patterns might be sufficiently different to allow icebergs to move into the area. Extensive early-middle Eocene ice on West Antarctica (Birkenmajer et al., 1986; Birkenmajer, 1987) may be a good candidate for supplying icebergs to the *Eltanin* core region. Unfortunately, there are currently no other Eocene cores from this area to test the iceberg track hypothesis further.

The detailed mechanism of Paleogene ice rafting is not yet well understood; for example the Oligocene ice-rafted quartz was found at a few locations but not in other areas, including some very near the Antarctic continent. It should also be stressed that the relatively warm Eocene epoch in general, as suggested by oxygen isotope and fossil data, does not necessarily exclude the possibility of significant ice on Antarctica. Frakes and Francis (1988) have even found Lower Cretaceous ice-rafted boulders in central Australia that indicate glacial activity during one of the warmest periods in the Phanerozoic. Warm climate would facilitate the transport of moisture inland and result in higher accumulation rate of snow. Modeling experiments suggest that the interior of the Antarctic continent remains quite cold regardless of the Southern Ocean surface temperature (Barron et al., 1981). Modeling of the Antarctic ice sheet (Oerlemans, 1982) also indicates that for an annual mean sea-level temperature of 0°C (about 20°C above the present), an ice sheet could cover most of Antarctica (both East and West) with ice thickness exceeding 4 km in some places; for an annual mean sea-level temperature of 5°C, one third of East Antarctica would still be covered with ice. After reviewing the Phanerozoic record of ice-rafted deposits, Frakes and Francis (1988) concluded that ice-rafted sediments exist at high paleolatitudes for every period of the Phanerozoic Era except the Triassic, and that the possibility of an ice-free Earth having ever existed appears small.

During the past quarter century, accumulating evidence has pushed the earliest Cenozoic glaciation history of Antarctica further and further back, from 2.4 Ma (Opdyke et al., 1966) to 4 Ma (Hays and Opdyke, 1967), to the late Oligocene (Hayes, Frakes, et al., 1975), and to the earliest Oligocene (Barrett, 1989; Barron, Larsen, et al., 1989, 1991; Schlich, Wise, et al., 1989; Breza and Wise, this volume). It is, therefore, not appropriate to assume that a complete Paleogene record of Antarctic glaciation has been recovered and well interpreted now. Further drilling and coring in the Southern Ocean, including the areas where the *Eltanin* cores were taken, should help unravel the Cenozoic history of Antarctic glaciation.

CONCLUSIONS

This study has updated the age for *Eltanin* Core 13-4 and established nannofossil biostratigraphy for *Eltanin* Cores 24-8, 24-9, and 24-10. More precise dates were obtained for the Eocene ice-rafted quartz identified by Margolis and Kennett (1971). Semiquantitative data indicate that the nannofossil assemblages in the cores contain more discoasters and fewer *Reticulofenestra* and *Toweius*, and show warmer water characteristics than those in the higher latitudes (such as Falkland Plateau, Maud Rise, and Kerguelen Plateau).

This study also indicates that ice-rafted quartz in the Eocene sediment is unlikely to be downcore contaminants and suggests that the quartz grains are most likely the result of Eocene ice-rafting from Antarctica when the Drake Passage was closed and water circulation patterns were different from those of today. Evidence of a early-middle Eocene glacial event(s) on West Antarctica (Birkenmajer et al., 1986; Birkenmajer, 1987), results of climatic modeling (Barron et al., 1981) and ice sheet modeling (Oerlemans, 1982), interpretations of sea-level curves of Haq et al. (1987), and oxygen isotope data (Matthews and Poore, 1980) all suggest significant ice volume on Antarctica during intervals of the early and middle Eocene, when glaciers probably reached coastal regions and produced large icebergs capable of traveling several thousand kilometers.

This study further suggests that the *Eltanin* locations were probably on iceberg tracks during the early and middle Eocene. The lowest Oligocene IRD recovered at ODP Site 748 and other sites, therefore, represents an episode of extensive glaciation on Antarctica in the earliest Oligocene, but should probably not be taken as the first glacial event during the Cenozoic.

TAXONOMIC NOTES

Toweius magnicrassus (Bukry) Romein, 1979
(Plate 1, Figs. 1, 2, and 13)

Coccolithus magnicrassus Bukry, 1971, p. 309, pl. 2, figs. 1–8.
non *Toweius magnicrassus* (Bukry) Romein, 1979, p. 216, pl. 4, figs. 2 and 3.
Toweius? *magnicrassus* (Bukry) Romein, Perch-Nielsen, 1985, p. 504, figs. 85.24 and 85.25.

Remarks. Plate 1, Figures 1 and 2, are probably the first published SEM photomicrographs of *Toweius magnicrassus*. This species resembles *Reticulofenestra hillae* Bukry and Percival (1971) but clearly belongs to the genus *Toweius*, because it has two cycles of elements around the central opening (Plate 1, Figs. 1 and 2) whereas species of *Reticulofenestra* have only one cycle around the central opening. In cross-polarized light, the collar around the small central opening is bright and quite distinct (Plate 1, Fig. 13). Bukry (1971) originally gave the size range of the species as 16–20 μm. The specimens illustrated in Plate 1, Figures 1 and 2, are about 18 and 13 μm, respectively. It appears that the species commonly ranges from 11 to 20 μm. The two specimens figured by Romein (1979, pl. 4, figs. 2 and 3) have a size of about 7 μm and are *Toweius callosus* Perch-Nielsen (1971) rather than *Toweius magnicrassus* (see also Perch-Nielsen, 1985, p. 505).

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APPENDIX

Calcareous Nannofossil Species Considered in This Study

Chiasmolithus eograndis Perch-Nielsen, 1971. Plate 2, Fig. 15
Chiasmolithus expansus (Bramlette and Sullivan) Gartner, 1970. Plate 3, Figs. 8 and 14
Chiasmolithus grandis (Bramlette and Riedel) Radomski, 1968. Plate 2, Figs. 13 and 14
Chiasmolithus solitus (Bramlette and Sullivan) Locker, 1968. Plate 1, Fig. 5; Plate 2, Fig. 8; Plate 3, Figs. 4 and 5
Coccolithus formosus (Kamptner) Wise, 1973. Plate 1, Fig. 9; Plate 2, Fig. 2; Plate 3, Fig. 12
Coccolithus pelagicus (Wallich) Schiller, 1930. Plate 3, Fig. 6
Discoaster barbadiensis Tan, 1927. Plate 1, Fig. 7; Plate 2, Fig. 18
Discoaster kuepperi Stradner, 1959. Plate 1, Figs. 3, 4, 10, and 11
Discoaster lodoensis Bramlette and Riedel, 1954. Plate 1, Fig. 6; Plate 2, Figs. 16 and 17
Discoaster saipanensis Bramlette and Riedel, 1954. Plate 2, Fig. 9
Discoaster sublodoensis Bramlette and Sullivan, 1961. Plate 1, Fig. 12; Plate 2, Fig. 12
Isthmolithus recurvus Deflandre, 1954. Plate 2, Fig. 11
Markalius inversus (Deflandre) Bramlette and Martini, 1964. Plate 1, Fig. 8; Plate 2, Fig. 1
Neococcolithus dubius (Deflandre) Black, 1967. Plate 2, Fig. 7
Reticulofenestra bisecta (Hay, Mohler and Wade) Roth, 1970. Plate 3, Fig. 11
Reticulofenestra daviesii (Haq) Haq, 1971. Plate 3, Figs, 1 and 2
Reticulofenestra samodurovii (Hay, Mohler and Wade) Roth, 1970. Plate 3, Fig. 9, 10, and 17
Reticulofenestra umbilica (Levin) Martini and Ritzkowski, 1968. Plate 3, Fig. 18
Sphenolithus moriformis (Brönnimann and Stradner) Bramlette and Wilcoxon, 1967. Plate 3, Fig. 3
Toweius magnicrassus (Bukry) Romein, 1979. Plate 1, Figs. 1, 2, and 13
Tribrachiatus orthostylus Shamrai, 1963. Plate 2, Figs. 3–6

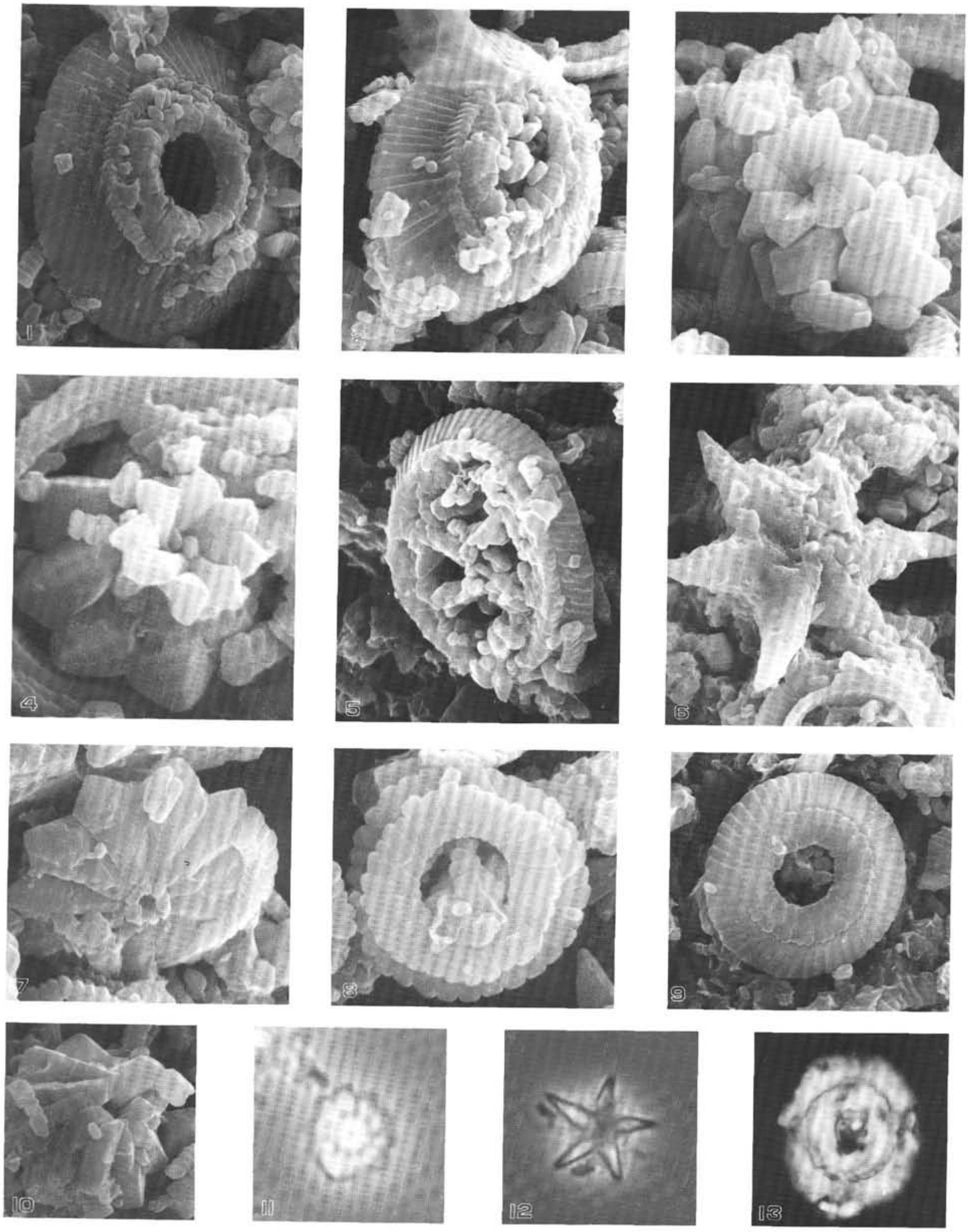

Plate 1. Calcareous nannofossils from *Eltanin* piston Core 13-4. **1–9, 13.** Sample 1666; (1, 2) *Toweius magnicrassus*, ×4000; (3, 4) *Discoaster kuepperi*, ×7000; (5) *Chiasmolithus solitus*, ×4000; (6) *Discoaster lodoensis*, ×3300; (7) *Discoaster barbadiensis*, ×5500; (8) *Markalius inversus*, ×5500; (9) *Coccolithus formosus*, ×3700; (13) *Toweius magnicrassus*, ×2000. **10–12.** Sample 1470; (10) *Discoaster kuepperi*, ×4000; (11) *Discoaster kuepperi*, ×2200; (12) *Discoaster sublodoensis*, ×2200.

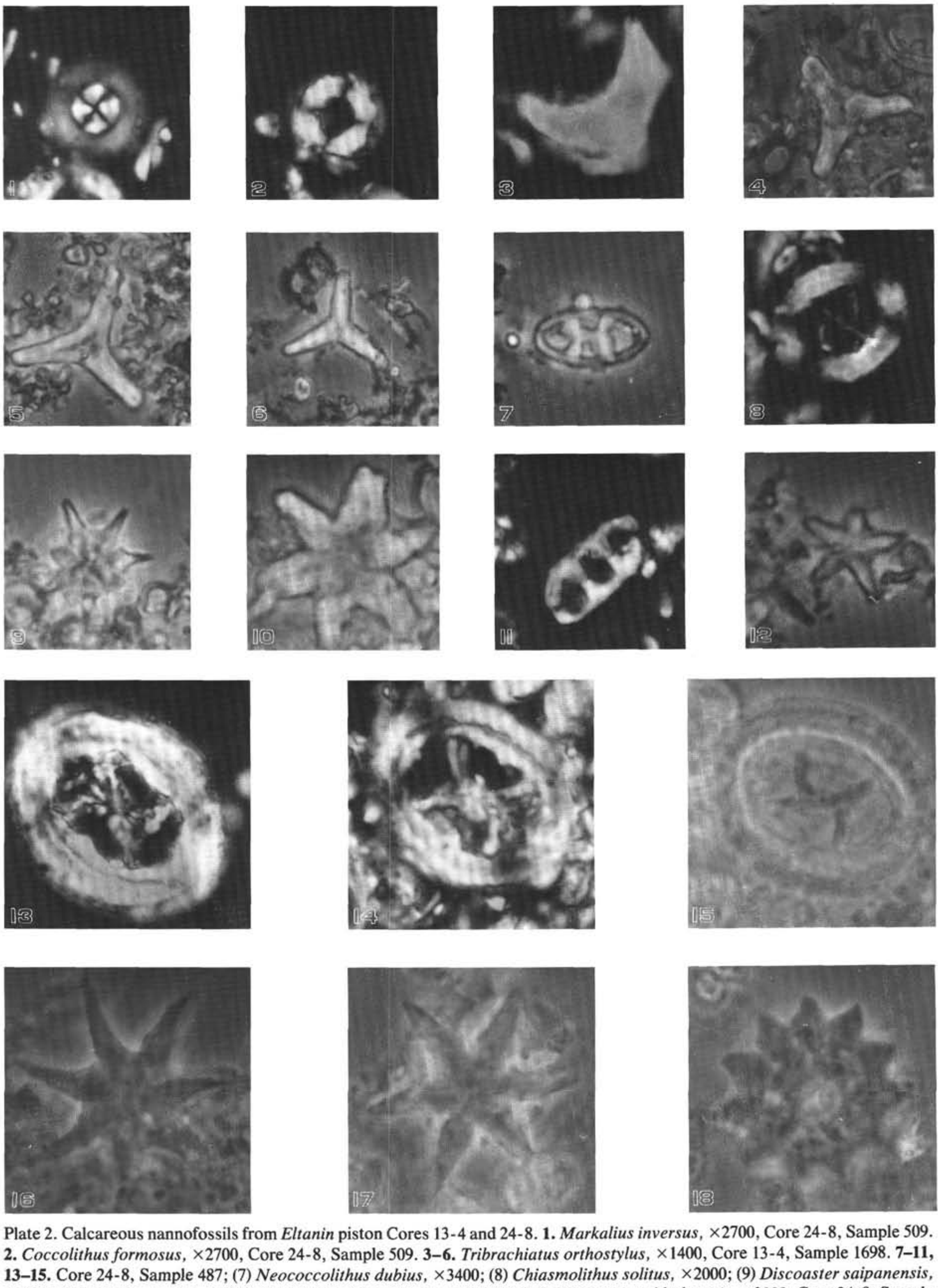

Plate 2. Calcareous nannofossils from *Eltanin* piston Cores 13-4 and 24-8. **1.** *Markalius inversus*, ×2700, Core 24-8, Sample 509. **2.** *Coccolithus formosus*, ×2700, Core 24-8, Sample 509. **3–6.** *Tribrachiatus orthostylus*, ×1400, Core 13-4, Sample 1698. **7–11, 13–15.** Core 24-8, Sample 487; (7) *Neococcolithus dubius*, ×3400; (8) *Chiasmolithus solitus*, ×2000; (9) *Discoaster saipanensis*, ×2000; (10) *Discoaster* sp., ×2000; (11) *Isthmolithus recurvus*, ×2700; (12) *Discoaster sublodoensis*, ×2000, Core 24-8, Sample 509; (13, 14) *Chiasmolithus grandis*, ×2400; (15) *Chiasmolithus eograndis*, ×2400. **16, 17.** *Discoaster lodoensis*, ×2000, Core 13-4, Sample 1666. **18.** *Discoaster barbadiensis*, ×2000, Core 13-4, Sample 1666.

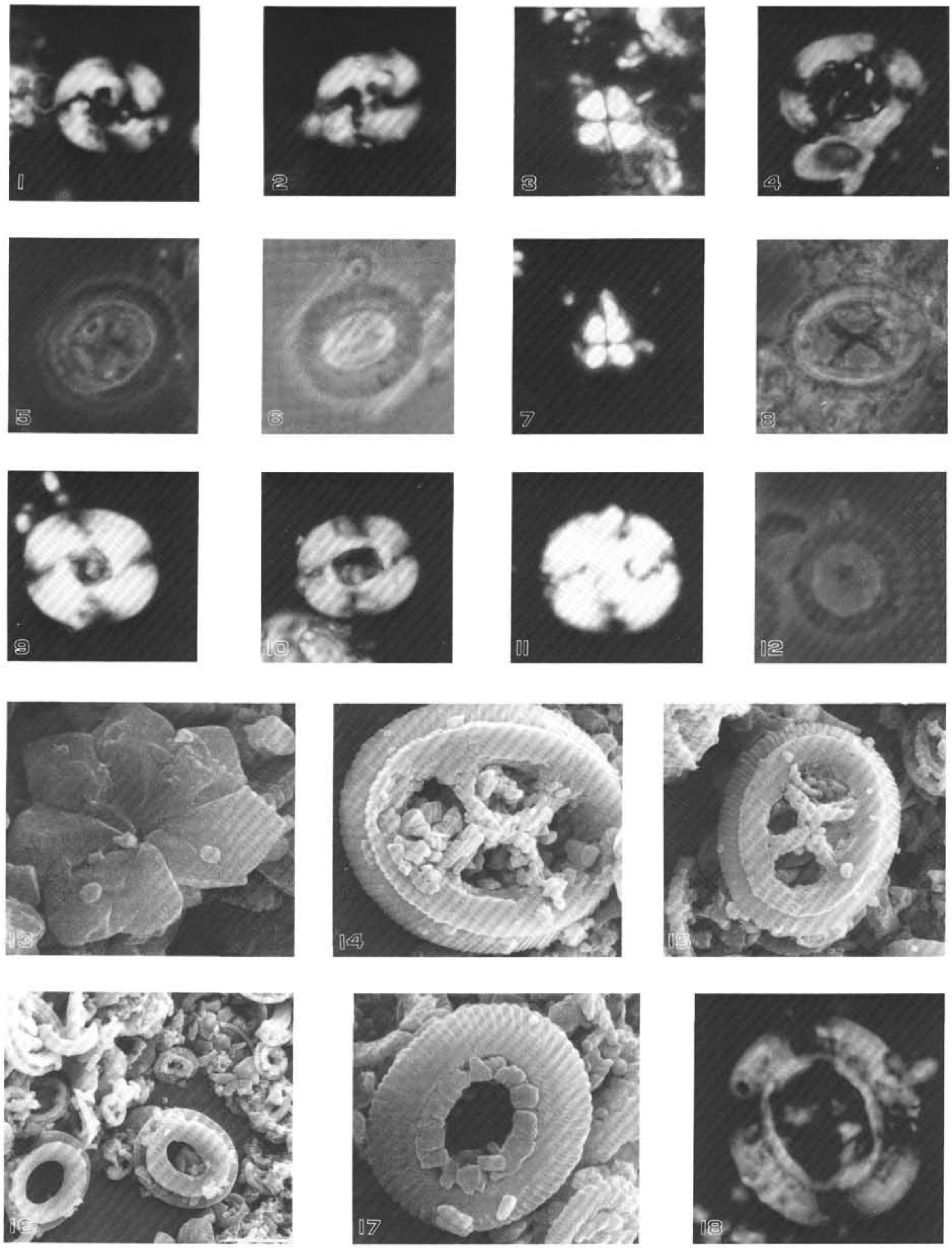

Plate 3. Calcareous nannofossils from *Eltanin* piston Cores 24-8 and 24-10. **1–6.** Core 24-8, Sample 487; (1, 2) *Reticulofenestra daviesii*, ×2800; (3) *Sphenolithus moriformis*, ×2800; (4, 5) *Chiasmolithus solitus*, ×2000; (6) *Coccolithus pelagicus*, ×2000. **7–12, 14–18.** Core 24-10, Sample 200; (7) *Sphenolithus* sp., ×2800; (8) *Chiasmolithus expansus*, ×1800; (9, 10) *Reticulofenestra samodurovii*, ×2400; (11) *Reticulofenestra bisecta*, ×2400; (12) *Coccolithus formosus*, ×2000; (14) *Chiasmolithus expansus*, ×2700; (15) *Chiasmolithus* cf. *expansus*, ×2200; (16) general appearance of the nannofossil assemblage, ×1100; (17) *Reticulofenestra samodurovii*, ×4500; (18) *Reticulofenestra umbilica*, ×2200. **13.** *Discoaster* sp., ×6500, Core 13-4, Sample 1543.