Reprinted from Nature, Vol. 299, No. 5882, pp. 435-437, September 30 1982

Reappraisal of sea-ice distribution in Atlantic and Pacific sectors of the Southern Ocean at 18,000 yr BP

L. H. Burckle, D. Robinson & D. Cooke*

Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York 10964, USA
* Mineral Management Service, Metairie, Louisiana 70003, USA

The question of the distribution of winter and summer sea-ice cover during the last glacial maximum (18,000 yr BP) is important in climatic modelling of the glacial world. We have now compared satellite-derived 5-yr monthly averages of per cent summer sea-ice cover in the Atlantic and Pacific sectors of the Southern Ocean with the distribution of reported sedimentological indicators of sea-ice cover at the sediment–water interface. These data strongly suggest that sediment type cannot be used to identify summer sea-ice limits, neither in Recent sediments nor in sediments of the last glacial maximum. Rather, the data seem to indicate that sediment type can be used to identify spring sea-ice limits for these two time intervals.

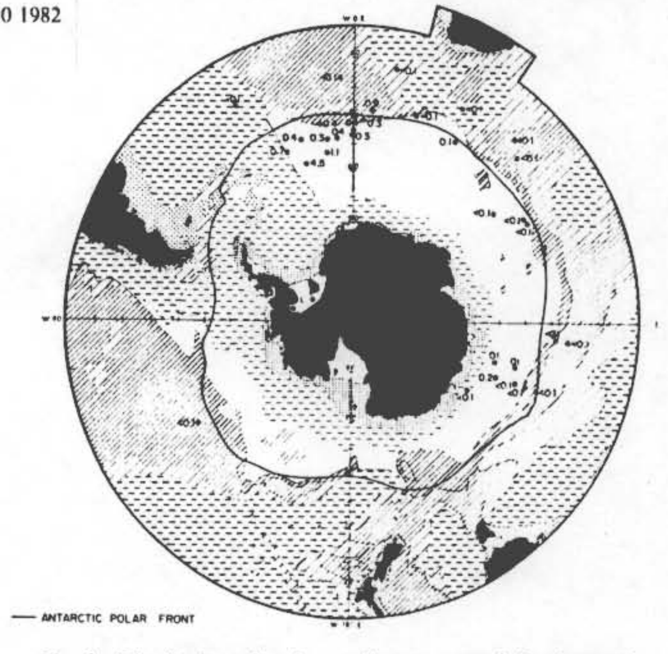

Fig. 1 Distribution of surface sediments around the Antarctic continent. Sediment types from the continent northward are as follows: parallel rows of dots = shelf and coastal deposits; light parallel dashed lines = siliceous silty clay and clayey silt; fine dots = siliceous ooze; oblique solid lines = calcareous ooze; oblique solid lines alternating with fine dots = siliceous and calcareous ooze. Numbers refer to percentage coarse fraction. Data are from ref. 9.

Attempts to define past sea-ice boundaries in the Antarctic began with Philippi[1] but considerable refinements have since been added[2–5]. Lozano and Hays[6] and Hays *et al.*[7] were the first to place the various sediment types within a good chronostratigraphical context. They were able to show that the thin veneer of diatomaceous ooze found in the Southern Ocean was Holocene in age while the underlying silty diatomaceous clay represented glacial conditions. These alternating sediment types were directly linked to the oxygen isotope record and permitted Hays *et al.*[7] and Cooke and Hays[8] to devise a chronostratigraphic framework for late Quaternary sediments of the Southern Ocean. These workers have argued that the boundary in surface sediment between the silty diatomaceous clay and the diatomaceous ooze represents the northern edge of summer sea ice (Fig. 1). Further, they argue that this boundary at 18,000 yr BP represents the northern limit of summer sea ice during a glacial maximum. As might be expected, this boundary is considerably north of the present-day boundary and closely approximates present-day winter sea-ice limits. The presence of diatoms within the silty diatomaceous clay is attributed both to leads which opened in the summer sea-ice, resulting in higher phytoplankton productivity, and to occasional summers when sea ice melted back almost to the continent (ref. 8 and J. D. Hays *et al.*, in preparation).

Here we suggest an alternative interpretation of the sediment analysis of Cooke[9] and Cooke and Hays[8]. We use modern-day sea-ice distribution and compare it with sediment distribution (and particularly sediment type boundaries) in surface sediments and at 18,000 yr BP[9]. Sea-ice cover, derived predominantly from satellite imagery[10], was averaged monthly over a 5-yr period (1973–77) for both the Atlantic and the Pacific sectors (Fig. 2). These data show a seasonally fluctuating ice front with a low percentage of sea-ice cover near the ice margin and higher values ranging between 75 and 100% some 6–7° behind the ice front. The lower values encountered in higher latitudes near the continent represent polynyas where high winds have pushed the sea ice away from the shore and maintained open water, sometimes well into the winter and in early spring.

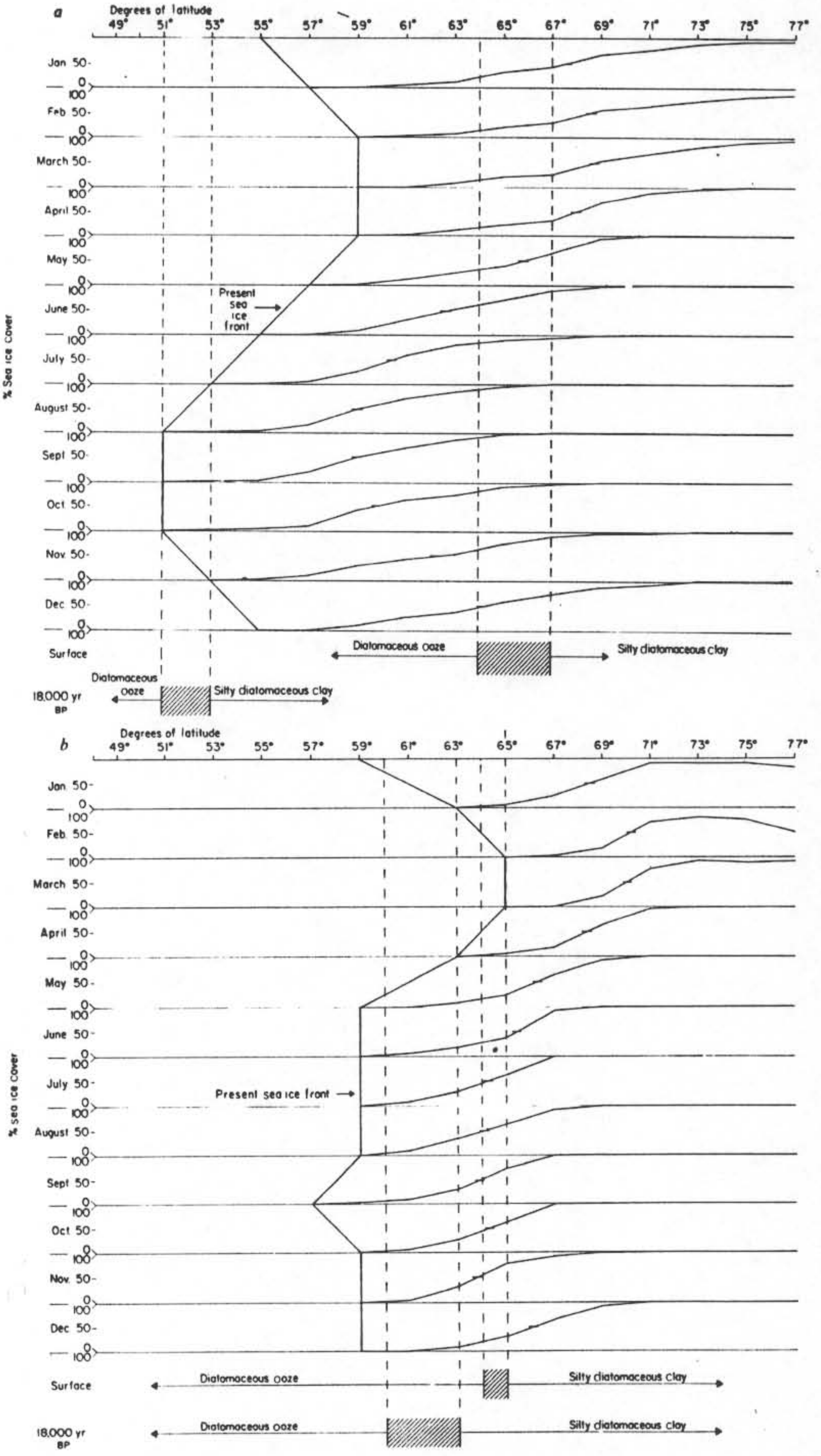

Fig. 2 Satellite-derived 5-yr monthly averages of per cent sea-ice cover for each 2° of latitude in the Pacific (*a*) and Atlantic (*b*) sectors the Southern Ocean. Data on sediment distribution in Recent and at 18,000 yr BP are taken from ref.[9].

Figure 2 also shows the present-day northern boundary between silty diatomaceous clay and diatom ooze[9]. This boundary is rather irregular but is between 64° and 65° in the Pacific sector and 64° and 67° in the Atlantic sector. If this boundary is projected up through the diagram, it is seen that in the Pacific the range of sea-ice cover is 0–0.1% during the months of February to March. In the Atlantic, on the other hand, the sea-ice concentration during those 2 months ranges over 15–30%. A sea-ice concentration of 0.1% seems insufficient to produce a silty diatomaceous clay on the sea floor. Equally doubtful is the notion that 15–30% sea-ice cover is enough to produce such a sediment type. It seems, therefore, that this sediment boundary cannot represent summer sea-ice limits.

Although we believe that this approach is better than those previously attempted, there are some obvious weaknesses in it. Kukla and Gavin[10], for example, have recently shown that in the 1930s, summer ice conditions were heavier than at present. However, this translates into a change in the position of the sea-ice edge of the order of 1–2° of latitude. Further, as Kukla and Gavin[10] point out, there are periodic variations ranging from a few years to tens of years in duration in oceanic circulation and these would probably affect sea-ice distributions. Heap[11] also reported oscillations in ice extent in the Weddell Sea and Palmer Peninsula. Finally, Kukla[12] and Kukla and Gavin[10] report extensive spring sea ice in 1972 and 1973 and again in 1981, with declining spring sea-ice cover in the intervening years. These data suggest, therefore, that the use of sea-ice cover averaged over a 5-yr period or better may be adequate for comparison with surface sediment data.

If not summer sea-ice limits, what event does this sediment boundary record? Figure 2 shows that the biggest drop in ice cover overlying this sediment boundary occurs in the late spring (November–December). In the Pacific, the change is from 60–80% in November to ≤5% in January. In the Atlantic, the change is less dramatic, but the ice cover drops from as much as 90% in November to slightly more than 40% in January. The fact that this ice lingers on into late spring before retreating produces several results. Most important is the fact that the presence of the ice inhibits open-ocean diatom productivity. There are, of course, diatoms living and growing in the underside of the sea ice, but this production is considerably less than in the open ocean. The diatom growing season begins in November north of the Polar Front and moves southward as the spring and summer seasons progress. As has been shown[13,14], the southward progression of this productivity maximum is heavily dependent on the melt back of the ice. Thus, sea ice can act to inhibit diatom productivity.

A second important point is the origin and distribution of the silt and clay size particles associated with diatoms in the sediment. Spring peaks in aerosol concentration and condensation nuclei have been recorded in the Antarctic atmosphere[15,16], suggesting a Southern Hemisphere similarity with that of Murozumi *et al.*[17], who found terrestrial dust concentrations of Greenland spring snows to be three times larger than those of other seasons. It seems that whereas this dust falls on the open ocean and is dispersed, the sea ice acts as a collector and selectively dumps it in the late spring and summer. This point, combined with the fact that the sea ice may act as an inhibitor of diatom productivity until well into the growing season, seems to be a reasonable mechanism for producing a silty diatomaceous clay on the sea floor.

The lack of additional data both of ice cover and of surface and 18,000 yr BP sediment distribution prevent us from further speculation. However, we feel that the true summer ice limit during the last glacial maximum is more appropriately placed south of all data points. This means that this limit has not been defined for the last glaciation and is a suitable target for study by palaeoceanographers.

The manuscript was reviewed by G. Kukla and D. Lazarus. Research was supported by NSF grants DPP 80-08011 and ATM 80-01470. This is Lamont-Doherty Geological Observatory contribution no. 3361.

Received 6 May; accepted 22 July 1982.

1. Philippi, E. *Deutsche Sudpolar-Expedition, 1901–1903* Vol. 2, 415 (1912).
2. Schott, W. *Recent mar. Sediments* Vol. ? 396 (1939).
3. Hough, J. L. *J. Sedim. Petrol.* **26**, 301 (1956).
4. Lisitzin, A. P. *Deep Sea Res.* **7**, 88 (1960).
5. Goodell, H. G. *Antarctic Map Folio Ser.* No. 17 (1973).
6. Lozano, J. A. & Hays, J. D. *Investigation of Late Quaternary Paleoceanography and Paleoclimatology*, 303 (Geological Society of America, Washington, DC, 1976).
7. Hays, J. D. *et al. Investigfation of Late Quaternary Paleoceanography and Paleoclimatology*, 337 (Geological Society of America, Washington DC, 1976).
8. Cooke, D. W. & Hays, J. D. *Proc. 3rd Symp. Ant. Geol. Geophys.* (in the press).
9. Cooke, D. W. thesis, Columbia Univ. (1978).
10. Kukla, G. & Gavin, J. *Science* **214**, 497 (1981).
11. Heap, J. A. in *Antarctica*, 187 (Praeger, London, 1965).
12. Kukla, G. J. *Climatic Change*, 114 (Cambridge University Press, 1978).
13. Hart, T. J. *Discovery Rep.* **21**, 261 (1942).
14. Burckle, L. H. & Clarke, D. B. *Nova Hedwegia* **54**, 1977 (1977).
15. Hogan, A. S., Barnard, S. & Somson, J. *Geophysical Monitoring for Climatic Change*, No. 7, 89 (NOAA Environmental Research Laboratories, 1978).
16. Mosley-Thompson, E. *Inst. pol. Stud. Rep.* (1980).
17. Murozumi, M., Chow, T. & Patterson, C. *Geochim. cosmochim. Acta* **33**, 1247 (1969).