A large deep freshwater lake beneath the ice of central East Antarctica

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IN 1974–75, an airborne radio-echo survey of ice depths over central East Antarctica led to the discovery of a sub-ice lake of unknown depth and composition, with an area of about 10,000 km^2 and lying beneath $\sim$4 km of ice[1]. In 1993, altimetric data from satellite measurements[2] provided independent evidence of the lake's areal extent, thus confirming it to be the largest known sub-ice lake by an order of magnitude. Here we analyse new altimetric and radio-echo data, along with existing seismic data[3], to show that the lake is deep (mean depth of 125 m or more) and fresh, and that it has an area that exceeds previous estimates by about 50%—dimensions comparable with those of Lake Ontario. We estimate that the residence time of the water in the lake is of the order of tens of thousands of years, and that the mean age of water in the lake, since deposition as surface ice, is about one million years. Regional ice-dynamics can be explained in terms of steady-state ice flow along and over the lake.

The satellite ERS-1, launched in 1991, was the first to be specifically programmed for altimetric surveys of large polar ice sheets. An analysis of the initial fast delivery data gave mean surface elevations every 6.7 km along the satellite track[2]. These defined areas of level ice, which roughly matched the lake boundary indicated by radio-echo sounding (RES), but differed by up to 30 km on the north-west of the lake and did not include Vostok, the Russian ice-drilling station (Fig. 1*b*).

The new lake boundary (Fig. 1*b*) is based on changes of surface slope derived from the waveform product[4] using the latest altimeter data. This agrees with the lake boundary shown by RES profiling (Fig. 1*a*) to within 5 km (maximum error in aircraft position from the inertial navigation[5]).

Owing mainly to the high noise level, seismic shooting studies of the ice cover of central Antarctica produced no evidence, before 1993, of the meltwater lenses beneath the ice sheet that had been suggested by Kapitsa[6] on morphological grounds. Although increasing shot depths[7,8] and higher frequency filtering[8] helped to reduce seismic noise levels, the only major reduction of noise on the Antarctic plateau had been achieved once by Kapitsa and Sorochtin[3] in 1964 by using a vertical seismometer spread from 2.5 to 49 m depth in a borehole at Vostok (Fig. 2). This clearly confirmed the ice depth as 3,700 m for the important ice-coring programme at Vostok, whereas a later echo was then interpreted as a sedimentary layer.

Confirmation of the ice depth and the presence of a water layer by RES profiling (Fig. 3*a*) led to reinterpretation of the seismic data shown in Fig. 2. The absence of any significant energy return between reflections from the ice–water interface and the lake floor on seismometers below 10 m depth, identification of the later echoes as P waves (Fig. 2), together with evidence from RES and altimetry, all confirm the presence of a deep water layer. Assuming 1,450 m s^{-1} as the P-wave velocity in water gives the water thickness as 510 m and the lake bed at 710 m below sea level.

Along RES flight lines, the depth of floating ice changes slowly. The lake bed is around 700 m below sea level near both ends of the lake. Assuming this applies over the whole lake bed, the mean water thickness beneath the south–north profile of flight 130 is 125 m and the total volume of water 1,800 km^3. The bedrock topography around the lake is similar to the topography around the rift valleys, especially of the deglaciated Lake Baikal, Siberia. Similar lake bathymetry would considerably increase the estimated volume of the sub-ice lake.

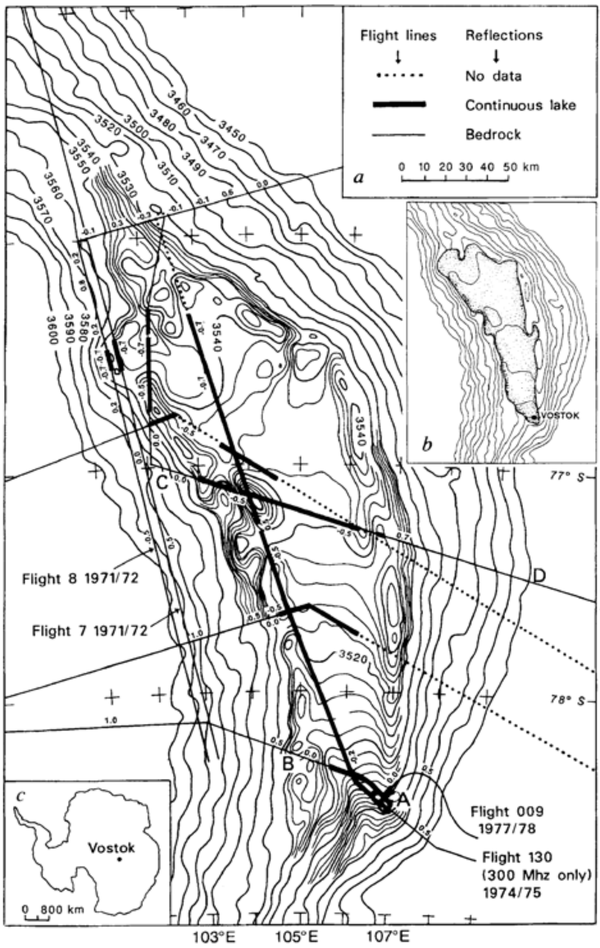

FIG. 1 Surface topography. ERS-1 waveform product (WAP) altimeter maps of the Vostok region. The improved altimetry comes from two 168-day repeat cycles of satellite orbits. The WAP analysis[4] gave 600,000 mean surface elevations over circles of 3 km diameter. Repeated and overlapping measurements gave an r.m.s. deviation of 20 cm and a maximum relative error in contours of 50 cm for the first time. This can now be extended over all inland ice north of 82° S. *a*, Contours over and near the lake boundary at 2-m intervals, with 10-m contours over steeper slopes. Figures along flightlines show bedrock elevations in kilometres above or below (−) sea level. Unpublished data now included come from flight 009 of 1977/78 and spot soundings every 2.4 km along the south-to-north flight 130 of 1974/75, which was not covered by continuous profiling (Fig. 3*a*). *b*, Shaded area shows the extent of the sub-ice lake, based on 2-m contouring but with 10-m contours over the entire area. This shows the low surface slope over the lake compared with the surrounding surface topography. *c*, Location of Vostok station.

With a surface accumulation of 2.7 g cm^2 yr^{-1}, an ice thickness of $\sim$3,700 m and a geothermal flux of $\sim$50 mW, steady-state theory[9,10] predicts basal melting of $\sim$1 mm yr^{-1}. Glacier dynamic theory indicates that the time taken for ice deposited on the surface to melt at the base is $\sim 10^6$ years. These estimates are in broad agreement with the age of the ice core at Vostok[11], shown by isotope profiling to a depth of 2,700 m.

Under steady-state conditions, a water input of 1 mm yr^{-1} from floating ice and a water layer 125 m thick gives a residence time of 1.25×10^5 years. Increasing the water input from basal melting of surrounding grounded ice and increasing the volume of the water in the same proportion leaves the residence time unchanged. With

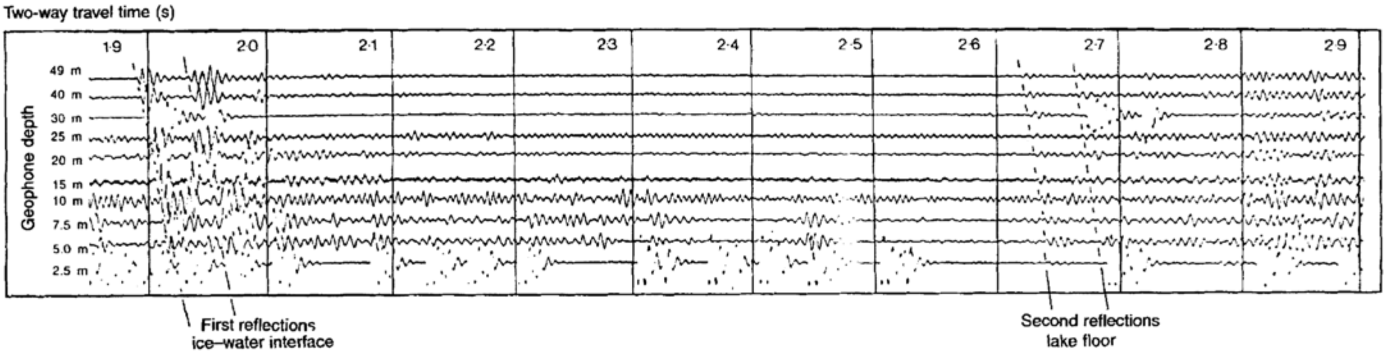

FIG. 2 Section of 24-channel seismic record from Vostok relevant to the sub-ice water layer. This records movement over a vertical line of seismometers from 49 m to 2.5 m depth in a borehole 180 m from the explosion. It covers a period from 1.85 to 2.9 seconds after the explosion of 5 kg of TNT at 39 m depth and 180 m from the vertical seismometer line. A conventional horizontal spread of twelve seismometers at 20-m intervals recorded the same echoes (not shown) at $\sim$1.92 and $\sim$2.65 s against a much higher background noise. The echo from the base of the floating ice reaches the deepest (49 m) seismometer first at $\sim$1.91 s. It then travels up the seismometer line to the surface where it is reflected down to pass the 49-m seismometer $\sim$50 ms later. This has a mean velocity of $\sim$2,200 m s^{-1}, typical of compression (P) waves in the top 50 m of firn in this region[12]. About 45 ms after the first arrival, a second wave train of similar intensity and duration follows as a result of the initial surface reflection of the explosion at 39 m depth. There is no significant return of energy between $\sim$2.00 and $\sim$2.63 s, when a weaker wave train passes up and down the seismometer line at the same velocity. This confirms that they are compressive (P) waves, the only waves that travel through water, and not transverse (PS) waves, which are sometimes recorded from shots on ice shelves.

both likely, our best estimate of the residence time of the lake water is of the order of 50,000 years.

Other features of the water mass can be deduced from the relationship between surface elevation and thickness of floating ice (Fig. 4). The close scatter of points around the freshwater line (Fig. 4) indicates a hydraulic pressure equivalent to a head of water close to 3,140 m above sea level over the whole lake. It also confirms the presence of relatively fresh water in the lake. Any difference between the mean slope of plotted points and the freshwater line could be due to residual effects from boundary stresses, to errors of ice thickness and/or to the effect of limited salinity on water density within a range of 0.00 and 0.05‰.

The decrease of ice thickness of $\sim$500 m from north to south of the lake means that the ice–water contact temperature will be 0.4 °C lower at the north than at the southern end. This may be sufficient to drive slow circulation of sub-ice water and affect basal heat flow distribution over the lake, but not the total heat flux.

Ice cores from Vostok contains micro-organisms[12,13] carried by air to Antarctica on dust particles from low latitudes. Melting of basal ice will release these microorganisms into lake water around one million years after deposition on the ice surface. Sediments on the lake bed may be several million years older, an exceptional environment that should provide useful information to the biological community and to geologists.

The lake occupies the lower part of a basin or rift (Figs 1, 3*b*). Unlike grounded ice, basal drag is negligible over an ice–water

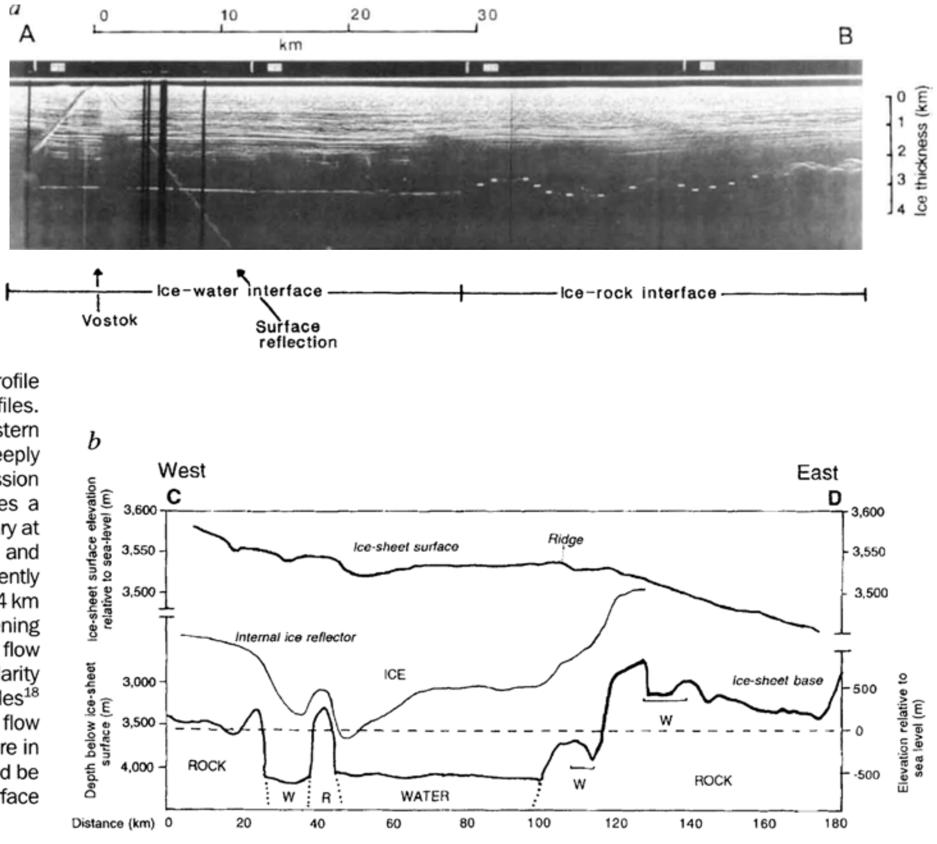

FIG. 3 Two-dimensional vertical profiles. *a*, Continuous RES profile from flight 009, 1977–78, along line A–B in Fig. 1*a*. Strong ice–water reflections beneath Vostok and for $\sim$30 km give way to weaker echoes from ice–bedrock further west. White dots have been added to show the location of weak bedrock echoes where they are not clearly visible. The near-horizontal layering in the upper 2 km is due to higher-conductivity layers of ice resulting from widespread deposition of volcanic material over the ice sheet[17]. Slanting echoes show reflections from surface buildings as the aircraft passed over Vostok. *b*, Cross-section along line C–D in Fig. 1*a*, showing deepest continuous RES layer. The vertical scale of the surface profile is increased to 100 times that of the deep profiles. Surface valleys around 30 and 50 km near the western boundary are due to downslope motion over steeply falling bedrock, opposed by horizontal compression that continues over the lake. After the ice crosses a rock ridge approximately parallel to the lake boundary at 40 km, a deeper valley is formed. The surface and reflecting layer then rise from 50 to 70 km, apparently above faster, southward-moving ice. From 70 to 94 km the surface is almost level, with slight ice thickening probable, until the lake boundary is crossed. A flow component to the east is then indicated by the similarity of the profile in *b* to that across the Doake Ice Rumples[18] (77.7° S, 66.6° W) to the Ronne Ice Shelf. Upslope flow appears to be initiated and driven by vertical pressure in both cases. The alternative to this flow pattern would be an ice divide, with static ice below an almost flat surface for at least 10 km to either side.

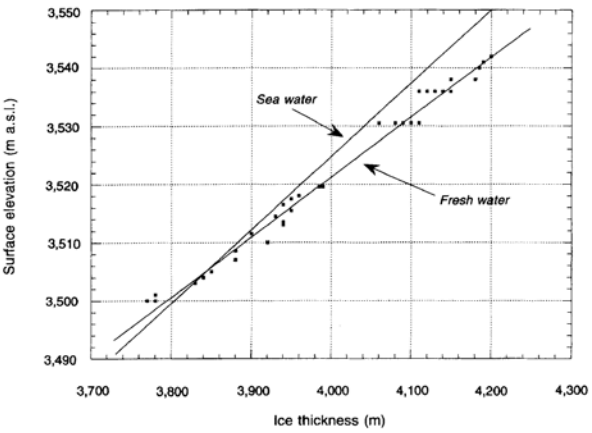

FIG. 4 Thickness (*h*) of floating ice by RES plotted against surface elevation (*E*) at locations clear of boundary effects along and across the lake on flight 130. Elevation of corresponding points is interpolated from satellite altimetry (Fig. 1). Also shown are computed slopes for ice in the same thickness range given by $dE/dh = (1 - \rho_{\text{ice}}/\rho_{\text{water}})$, where ρ denotes density. The lines shown apply to pressures from 30 to 40 MPa (ref. 19), salinity of sea water[20] of 35‰, and a mean density of ice of 0.913 g cm^{-3} from the borehole at VS (J.-R. Petit, personal communication). Resultant slopes are 0.105 for fresh water and 0.127 for sea water. Relative errors in ice thickness measurements should not exceed 30 m.

interface. Away from boundary effects, ice is compressed vertically and expands horizontally in directions offering least resistance. This produces a low surface slope normal to the surface contours as on ice shelves[14,15]. At Vostok[16], flow is normal to local contours. In general, contours show flow along the length of the lake, starting eastwards at 76.5° S and swinging southwards by 77.0° S. The near-parallel tilt of the deep reflecting layer to the valley walls (Fig. 3*b*) is also typical of flow in the direction of a long valley[17].

The above flow requires a high influx of ice over the northwest boundary. Here steep surface slopes, thick grounded ice with basal melting and sliding, and ice-sheet contours to the west, all indicate a large inflow at around 76.5° S.

Along almost all the eastern boundary, the ice surface rises ~8 m in 10 km before falling at the regional slope to the east, still over rising bedrock (Fig. 3*b*). This is similar to profiles across the Ronne Ice Shelf rumples[18]. In both cases, the initial rise extends over three times the ice thickness (*h*). Although magnitudes of ice thickness, slopes and velocities differ considerably, the same basic mechanism driven by vertical compression seems to apply.

The loss of ice over the eastern boundary will decrease velocity towards the south. To maintain vertical compression at around a constant level in order to drive flow to the south and east requires an increased surface slope in proportion to the slowing of velocity. Narrowing of the lake southwards, as with ice shelves, also requires increased surface slope.

Estimates of the easterly ice movement onto the lake, based on steady-state discharge of accumulation from the ice divide (dome B) 240 km to the west, give a mean inflow at ~3 m yr^{-1}. The only direct measurement of ice velocity in the region was made at Vostok[16] by star sights in 1964 and 1972. this gave 3.7 ± 0.7 m yr^{-1} at $142 \pm 10°$ true, approximately normal to contours within 20 km of the station. This gives an eastward component of 2.3 ± 0.5 m yr^{-1}. When the loss of ice across the eastern boundary is considered, the flow at Vostok equals that expected for steady-state flow to well within the error of estimates and measurements.

Further geophysical studies of the lake are being planned. Satellite global positioning system measurements of surface velocity in three dimensions will show how closely steady-state flow applies from changes of surface elevation with time, while measured horizontal motion will test predictions of flow from ice dynamics. Seismic sounding and RES survey will give a much better knowledge of the lake depth and extent and determine the nature of any lake bed sediments. Finally, sampling of lake water and sediments should be undertaken, subject to Environmental Impact Assessment under the Antarctic Treaty, when this can be achieved without polluting the lake water. ☐

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Geochemistry of mantle–core differentiation at high pressure

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THE apparent excess of siderophile (iron-loving) elements in the Earth's mantle has been a long-standing enigma in the geochemistry of mantle–core differentiation[1,2]. Although current models have proved successful in explaining some aspects of this problem[3–7], important questions remain. In particular, the mantle's near-chondritic ratio of nickel to cobalt (close to that expected for the material from which the Earth formed) is hard to explain, given the markedly different ambient-pressure partitioning behaviour of these elements between iron-alloy and silicate melts[3–8]. Here we report experimental results which show that both elements become less siderophile with pressure, but the effect is much more pronounced for Ni, so that the partition coefficients of the two elements become essentially equivalent at an extrapolated pressure of ~28 GPa. The absolute and relative abundances of Ni and Co in the mantle are therefore consistent with alloy–silicate chemical equilibrium at high pressure, indicating that core formation may have taken place in a magma ocean with a depth of 750–1,100 km. We also find that, unlike Ni and Co, sulphur becomes more siderophile with pressure. Sulphur's increased affinity for iron with depth could make it the dominant light element in the Earth's core.

Early studies on core–mantle formation based on low-pressure partitioning data suggested an apparent disequilibrium between the mantle and the core[1,2,5,6,8]. Moderately and highly siderophile elements are far in excess in the mantle with respect to what would be expected from these data. Moreover some elements with very different low-pressure partition coefficients have near-chondritic