3 Huguenin, G. R., and Papagiannis, M. D., [illegible] 28, 239 (1965).
4 Alexander, J. K., and Stone, R. G., *Astrophys. J.*, [illegible]
5 Mendillo, M., Hawkins, G. S., and Klobuchar, J. A., *Science*, 187, 343 (1975).
6 Mendillo, M., Hawkins, G. S., and Klobuchar, J. A., *J. geophys. Res.* (in the press).
7 Fehsenfeld, F. C., Schmeltekopf, A. L., and Ferguson, E. E., *J. chem. Phys.*, 46, 2802 (1967).

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Thermal regime of the Earth's interior

SEVERAL recent and apparently unrelated geophysical observations and deductions, when considered together, necessitate a fundamental re-examination of our ideas about the thermal state of the Earth's deep interior.

In summary they are:

(1) Precession must be discounted as a power source for the geomagnetic dynamo[1], leaving core convection as the only plausible basic mechanism. Therefore the temperature gradient in the outer core cannot be less than adiabatic.

(2) The thermodynamic Grüneisen parameter for the outer core is about 1.4 and is only slightly pressure dependent (R. D. Irvine and F.D.S., unpublished).

(3) The resulting adiabatic gradient is much steeper than the Fe melting point gradient[2] or the Fe–S eutectic gradient[3].

(4) Continents and ocean floors are both mobile, although the ocean floors are much more so[4], so that equality of the continental and oceanic heat fluxes which is observed[5] (in spite of gross differences between the radioactive constituents of the continental and oceanic crusts) cannot be explained in terms of an association of depleted regions of the mantle with continents.

(5) Hot spots, across which the tectonic plates move, are not themselves moving perceptibly with respect to one another and are presumed to mark features of a rigid lower mantle[4].

(6) A weak layer (asthenosphere) in the upper mantle allows virtually perfect isostatic balance of continent-sized blocks[6] but the lower mantle must have greater strength. Low degree harmonic features of the geoid are supported by stresses at depths of several hundred kilometres[7].

(7) Radiative transfer does not cause a dramatic increase in the thermal conductivity of mantle material at high temperatures[8,9].

As long as there remained a possibility that the geomagnetic dynamo is driven by precession[10], the fluid, outer core could be supposed to be stably stratified with a subadiabatic gradient. Although unconventional, the operation of a dynamo with the core in such a state could not be precluded[11] and the 'core paradox'[2,12] was avoided. This paradox arises because the adiabatic gradient in the core is much steeper than the melting point or eutectic temperature gradient, so that the conventional explanation of the solidity of the inner core as a pressure solidification of material from the outer core breaks down if the outer core is supposed adiabatic. But there now seems no possibility that precessional dissipation suffices for dynamo action. A 'rigid' core calculation indicated that the available power might be 3×10^{10} W (ref. 13), but Rochester[1] reported that proper inclusion of the internal core motions associated with core coupling to the mantle reduced the power estimate to less than 10^8 W, which certainly does not suffice to maintain the dynamo. Thus the temperature gradient of the outer core cannot be less than adiabatic.

So, how does one explain the apparently solid inner core? Elsasser and Isenberg[14] favoured an electronic phase transition in iron to the $3d^8$ state—collapse of all of the conduction or valence electrons into vacant 3d states. Probably because very little is known about such transitions (although one has been identified in caesium[15]) this suggestion has languished in obscurity; but now it emerges as the only real survivor. But it completely removes previously assumed constraints on the temperature of the core, based on the assumption that the inner core boundary represented the intersection of the temperature profile with the solidus of an iron-rich mix. Whether the inner core is really a crystalline solid must be doubted in view of its very high Poisson's ratio[16], which suggests some residual fluid character, but in any case extrapolation of ordinary melting point or eutectic data is irrelevant to an electronic phase transition.

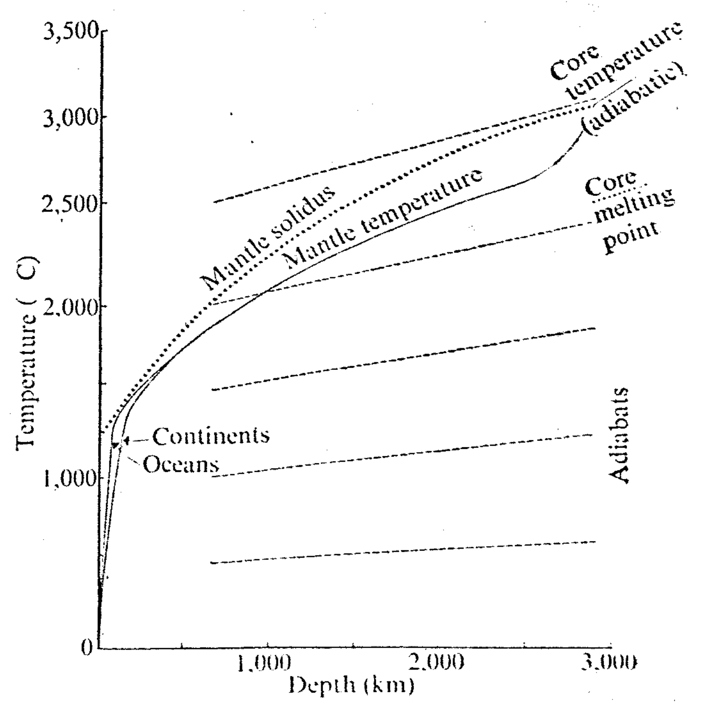
Rejection of core-constrained temperature profiles clears the way for an alternative approach, based on recent observations on the behaviour of the mantle. First, we note that the heat flux from the mantle into the continental crust is only about $0.015\ \mathrm{W\ m^{-2}}$, the balance of the total geothermal flux (about $0.06\ \mathrm{W\ m^{-2}}$) being provided by crustal radioactivity. But, since the continents ride about on rigid slabs of lithosphere, this mantle heat flux must be transmitted from depth by conduction. With an assumed conductivity of $2.5\ \mathrm{W\ m^{-1}\ deg^{-1}}$, the required lithospheric temperature gradient extrapolates to temperatures of partial melting ($\sim$ 1,300 °C) at a depth of 150 km or so. This determines the thickness of the continental lithosphere, the more fluid asthenosphere below it being in convection. On the other hand the oceanic lithosphere is not a permanent feature. After formation at an ocean ridge, any section of it survives only for about 10^8 years before plunging back into the mantle at a subduction zone. If we suppose that its duration is simply its thermal relaxation (cooling) time τ, then with a thermal diffusivity representative of igneous rock, $\eta=1.2\times10^{-6}\ \mathrm{m^2\ s^{-1}}$, we can estimate the thickness of the oceanic lithosphere as:

$$D=(\tau\eta)^{\frac{1}{2}}=70\ \mathrm{km}$$

It is important that these simple arguments give reasonable lithospheric thicknesses and are in accord with the equality of continental and oceanic heat fluxes without requiring any difference in radioactive heat generation in different parts of the mantle. Not only does it remove the implausible constraint to tectonic and convection theories that the continents must move about with their own sections of mantle, depleted in radioactivity, but also there is no reason for the temperature difference between the subcontinental and suboceanic mantle to extend into the lower mantle.

Increasing strength with depth below the asthenosphere (observation (6)) implies that the temperature profile diverges again from the solidus at greater depths. But we also require convective instability of the lower mantle to remove heat from the core, which, with a reasonably constrained conductivity[17], cannot be less than about 3×10^{12} W (ref. 18). In the case of the

Fig. 1 Suggested temperature profile of the mantle.



mantle, which has both a lower temperature and a smaller Grüneisen parameter, ~0.9 (R. D. Irvine and F.D.S., unpublished), than the core, we can hardly avoid the conclusion that the solidus gradient is steeper than the corresponding adiabat, at least in the upper levels, so that there is no difficulty in accepting an actual temperature gradient between the two. This is precisely the condition required for the establishment of narrow convective plumes[19]. The temperature gradient in the lower mantle gives increasing strength with depth, inhibiting convection in the whole mantle, but core heat must melt or nearly melt a thin layer at the base of the mantle. As this melted layer tends to rise, it finds that it is in a super-adiabatic gradient and so is convectively unstable, but that as it rises its melting point falls more rapidly than its actual temperature (decompression being approximately adiabatic) so that it becomes increasingly fluid and therefore heats a narrow channel to guide subsequent flow.

These considerations are summarised by the illustrative temperature profile in Fig. 1. This curve can be extended into the outer core for any assumed depth of convection by taking the adiabatic gradient (for which the assumption of a constant Grüneisen ratio of 1.4 suffices), but the required heat sources may be distributed through the outer core only, so that eventually we must reach a depth at which the heat flux is insufficient to support the adiabatic gradient. Below this the heat flow is diffusive only, but detailed calculations[20] show that the depth of transition from convective to diffusive regimes is very sensitive to the total heat flux; a 10% increment above the flux required just to maintain diffusion down the adiabatic gradient causes convection over most of the outer core. Thus it is reasonable to assume an adiabatic gradient through the whole outer core. This gives a temperature ratio between the inner core boundary and core–mantle boundary, $T_{IC}/T_{C-M}=1.33$ and, with the values in Fig. 1, an inner core temperature of 4,100 °C.

It may be noted that the necessity for 3×10^{12}–5×10^{12} W of radiogenic heat in the core seems unavoidable. Thus the geophysical requirement for potassium in the Fe–S outer core seems even stronger than the geochemical evidence[21–23] for it.

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Received February 20; accepted March 25, 1975.

1 Rochester, M. G., *EOS (Trans. Am. geophys. Un.)*, 56, 1108 (1974).
2 Higgins, G., and Kennedy, G. C., *J. geophys. Res.*, 76, 1870 (1971).
3 Tolland, H. G., *Phys. Earth planet. Interiors*, 8, 282 (1974).
4 Jordan, T. H., *EOS (Trans Am. geophys. Un.)*, 56, 1102 (1974).
5 Lee, W. H. K., *Phys. Earth planet. Interiors*, 2, 332 (1970).
6 Stacey, F. D., *Physics of the Earth* (Wiley, New York, 1969).
7 Higbie, J., and Stacey, F. D., *Phys. Earth planet. Interiors*, 4, 145 (1970).
8 Fujisawa, H., Fujii, N., Mizutani, H., Kanamori, H., and Akimoto, S., *J. Geophys. Res.*, 73, 4727 (1968).
9 Schatz, J. F., and Simmons, G., *J. geophys. Res.*, 77, 6966 (1972).
10 Malkus, W. V. R., *J. Geophys. Res.*, 68, 2871 (1963); *Science*, 160, 259 (1968).
11 Bullard, E., and Gubbins, D., *Nature*, 232, 548 (1971).
12 Kennedy, G. C., and Higgins, G. H., *J. geophys. Res.*, 78, 900 (1973).
13 Stacey, F. D., *Geophys. J. R. astr. Soc.*, 33, 47 (1973).
14 Elsasser, W. M., and Isenberg, I., *Phys. Rev.*, 76, 469 (1949).
15 Sternheimer, R., *Phys. Rev.*, 78, 235 (1950).
16 Dziewonski, A. M., Hales, A. L., and Lapwood, E. R., *Phys. Earthplanet. Interiors* (in the press).
17 Gardiner, R. B., and Stacey, F. D., *Phys. Earth planet. Interiors*, 4, 406 (1971).
18 Stacey, F. D., *Geophys. Surveys*, 1, 99 (1972).
19 Morgan, W. J., *Nature*, 230, 42 (1971).
20 Metchnik, V. I., Gladwin, M. T., and Stacey, F. D., *J. Geomag. Geoelect. (Kyoto)* 26, 405 (1974).
21 Goles, G. G., in *Handbook of Geochemistry*, 1, (edit. by Wedepohl, K. H.), 116 (Springer, Berlin, 1969).
22 Lewis, J. S., *Earth planet. Sci. Lett.*, 11, 130 (1971).
23 Hall, H. T., and Murthy, V. R., *Earth planet. Sci. Lett.*, 11, 239 (1971).

Wire strainmeters on ice

WE have installed a geophysical wire strainmeter for brief periods on the Roslin glacier, Stauning Alps, East Greenland and on the sea ice at Bylot Sound, Thule, North-west Greenland. The technique promises to yield information concerning the mechanism of deformation and flow in ice.

Previous studies of ice flow have involved intermittent measurements of surface markers[1], but although such methods

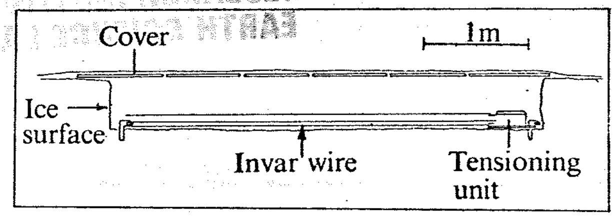

Fig. 1 The strainmeter installed on the glacier.

illustrate the presence and rate of cumulative motion over large distances they do little to reveal the internal mechanism of flow. Few measurements have been made of the detailed strain behaviour of ice continuously at a point *in situ*[2,3]. For example, it is not known whether glacier flow is episodic and rapid with long periods of quiescence, or whether continuous creep occurs.

The wire strainmeter used in the experiments was designed for studies of earth strain (ref. 4 and G. C. P. King and R. G. Bilham, to be published) and has a measuring range of 10^{-4} strain (Fig. 1). The strainmeter uses 0.5-mm Invar wire held in tension by a lever and weight system. Rotation of the tensioning lever is detected electronically, amplified, filtered and recorded on a chart recorder. Using it as a 10-m strainmeter results in a maximum strain resolution of 10^{-10} in an environment where the temperature is stable. The temperature coefficient of the instrument is approximately 5×10^{-7} per °C. The instrument weighed only 10 kg and was easily transported to the sites.

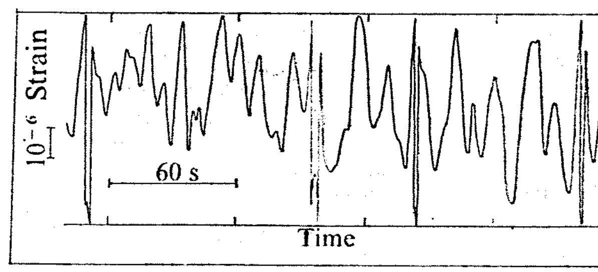

Fig. 2 Continuous record of strain changes measured on the sea ice in Bylot Sound, North-west Greenland in June 1974. The abrupt changes are d.c. shifts automatically introduced to keep the record on the chart.

The instrument was connected to the ice using two ice screws—a quick and simple means of obtaining the attachment stability required. Temperature stability (0 ± 0.5 °C) was obtained by building a cover when the strainmeter was on the sea ice and digging a 0.5-m trench and covering it when the strainmeter was installed on the glacier.

In June 1974 the instrument was installed on some first-year ice in Bylot sound (76°27′ N, 69°30′ W). The ice there was 0.86 m thick 71.7 m from the ice edge and 140 m from the shore. Some of the data obtained are shown in Fig. 2. Figure 3 is a frequency transform of approximately 1 h of data. The high-frequency part of the spectrum is suppressed by electronic filtering (−3 dB at 6-s periods). The spectrum is similar to that obtained from sea waves[5] with a dominant peak at approximately 16 s. There is a pronounced minimum between periods of 1 and 2 min, and the waves of longer period (20 min) present probably arise from seiches in the bay[6].

In July and August the instrument was taken to the Roslin glacier (71°51′N, 24°59′W). The glacier is a typical sub-polar–valley glacier 33 km long, 2 km wide and 300 m deep at the site. The surface velocity at the site, determined from four years of stake measurements, is about 9.5 m yr^{-1}. The instrument was installed centrally, parallel to glacier motion, slightly below