snow impacting on the surface are also effective bubble producers (*13*).

The bubble-jet-drop mechanism of water-to-air transport of bacteria and virus deserves more study. Many species of bacteria as well as other organisms should be used, both aerobic and anaerobic, sporal and nonsporal. The depth of water through which the bubbles rise should be varied because bacteria can become attached to bubbles and be carried to the surface. Our initial work in this direction with *S. marcescens* indicates that the concentration factor *C* for the top jet drop increases by about five times as the depth of bubble origin increases from 1 to 30 cm.

A high concentration of many species of organisms would be expected at air-water interfaces, for it is here that they can find the organic material they need for survival (*14*). Not only might viable material be concentrated in natural surface films, but one would also expect to find high concentrations of DDT and radioactivity. One of the main sources of these films is plankton and fish, both of which are known to concentrate DDT and radioactivity (*15*). If this material is present near the surface, then it, too, like *S. marcescens*, may become concentrated in airborne drops.

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Thermochemical Remanent Magnetization and Thermal Remanent Magnetization: Comparison in a Basalt

Abstract. *Recent studies have shown that the remanent magnetization carried by an extrusive igneous rock may not be entirely thermal remanent magnetization (TRM). Some may be thermochemical remanent magnetization (TCRM) acquired by the rock at temperatures at least as low as 300°C during oxidation of the contained titanomagnetite grains. Results from a study of a set of basaltic samples from one locality indicate that the intensity of TCRM acquired by a sample in a known magnetic field is equal to that of TRM subsequently produced in the same sample in the same field. On the assumption that the samples we studied are not magnetically unique, we tentatively conclude that paleointensity studies are valid in spite of the presence of TCRM, as long as the rock acquired the magnetization during the initial cooling.*

Verhoogen has pointed out (*1*) that the stable phases of the natural iron-titanium oxides in air at low temperatures are hematite and rutile. The fact that we commonly observe other phases, such as ilmenite and members of the ulvöspinel-magnetite solid-solution series, in basalts suggests that either the basalts cooled quickly or there was insufficient oxygen (such as air or oxygenated water) present to cause the reactions to go to completion.

Certain quickly quenched basalts often have virtually unaltered titanomagnetites, with Curie temperatures of 200°C or less (*2, 3*). When these basalts (or powdered preparations made from them) are heated in air at a temperature higher than the initial Curie temperature, certain changes occur, notably an increase in Curie temperature and saturation magnetization J_s (*3, 4, 4a*). The changes in magnetic character occur simultaneously with the alteration of the original titanomagnetite to ilmenite and a ferrimagnetic mineral (generally titanomagnetite) of lower titanium content than the original material. Further heating in air will cause the iron-titanium oxides to oxidize to hematite and rutile; and at temperatures at or above 600°C pseudobrookite can be formed (*4a, 5*).

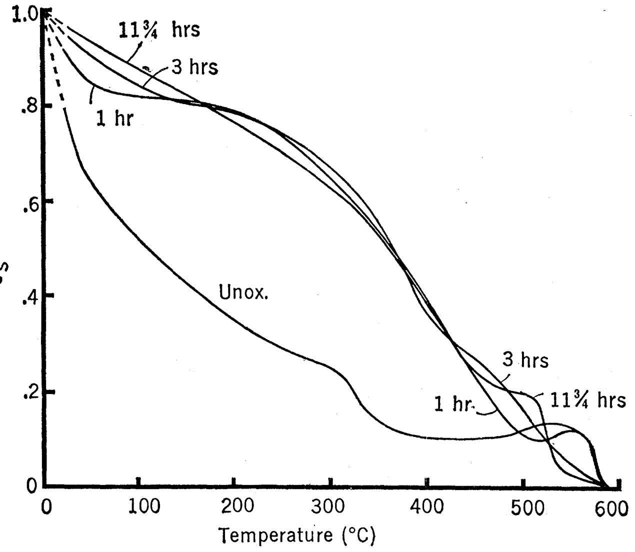
During the heating, in the presence of oxygen, of a basalt that contains homogeneous titanomagnetite, a stable remanence may be induced at temperatures far lower than the final Curie temperature (or range of Curie temperatures) of the magnetic material. Remanence acquired in virtue of this process has been referred to as thermochemical remanent magnetization (TCRM) (*6*).

In 1967 Carmichael and Nicholls pointed out (*7*) that the titanomagnetite in most naturally occurring basalts has undergone some oxidational modification and that the natural remanent magnetization (NRM) was probably not entirely thermal remanent magnetization (TRM). A more recent study of basalts in some Hawaiian lava lakes (*8*) reconfirmed the latter conclusion and indicated that the NRM of the lava-lake basalts was probably a TCRM.

We report here some preliminary results concerning the nature of TCRM. Our conclusions bear significantly on the validity of some paleomagnetic studies which have been carried out under the assumption that the magnetization of naturally occurring samples is a simple TRM.

A set of cores (2.5 cm in diameter) was drilled from a highly vesicular basalt collected in northern New Mexico. Analysis with a Curie balance indicated that the magnetic opaque minerals possessed an initial Curie temperature of 340°C. To facilitate uniform oxidation throughout the specimens and to increase the rapidity of heating and cooling, the cores were sliced into disks 0.4 to 2.0 cm thick.

In order to obtain the rate at which the samples altered, two equal-sized disks, each 0.4 cm thick, were heated, one at 300°C and the other at 400°C, in air for various periods up to 2 days. The heating was interrupted at different stages, and the acquired TCRM was measured. Care was taken to replace the samples in exactly the same position in the furnace after each measurement. Heating was done in the vertical component of the earth's field; Helmholtz coils were used to cancel the horizontal components. The samples were then cooled in field-free space to room temperature and removed from the furnace; their magnetization was measured on a Marathon spinner magnetometer with a sensitivity of 10^{-7} electromagnetic unit/cm^3 (*9*). During heating at 400°C, there was a rapid increase in magnetization, which leveled off in 20 to 30 hours. The rate of increase at 300°C was much slower than it was



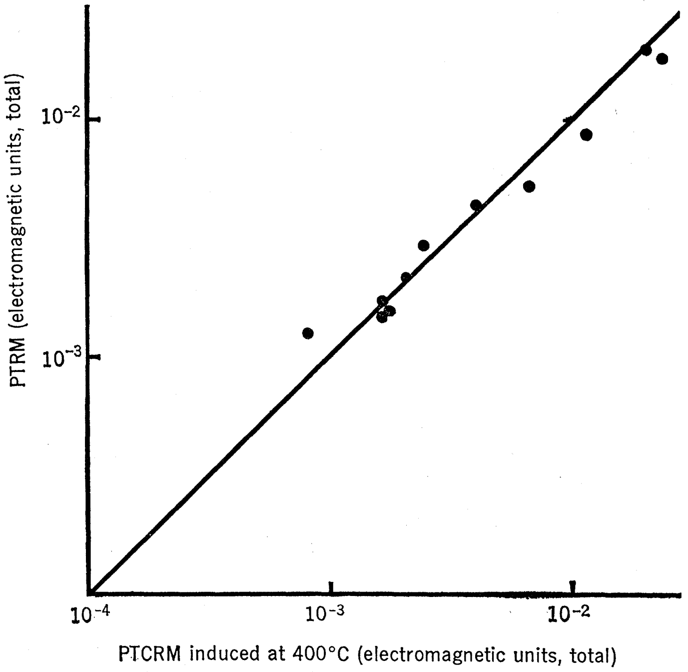


Fig. 1 (above). Curves of saturation magnetization (J_s) plotted against temperature for an unoxidized sample and for samples held at 400°C for 1, 3, and 11¾ hours before measurement. All curves are normalized to an initial value for J_s of 1.0. The return (cooling) curves, which lie significantly above the heating curves, are not shown, as they depend on the amount of time the samples were held at the higher temperature and on the maximum temperature reached. Fig. 2 (right). Partial TRM as a function of partial TCRM for 11 variously sized samples between 400°C and the final Curie temperature. The line represents a theoretical 1:1 correlation between PTRM and PTCRM.

at 400°C but was not imperceptible.

Figure 1 shows the change in saturation magnetization with time for the sample heated at 400°C. (All are heating curves; the cooling paths are not shown.) The opaque minerals in the unoxidized sample possess a range of low Curie temperatures with one being prominent at about 340°C. A final Curie temperature of about 580°C indicates that some pure magnetite was produced during heating in the Curie balance. After TCRM inducement for 1 hour, the Curie temperature increased and the curve became more inflated. After 11¾ hours, oxidation was still not complete (however, the return path more nearly retraces the heating curve). As shown, the greatest oxidational changes occurred during the early stages of heating.

In order to compare TCRM with TRM, 11 disks of differing thicknesses were heated in the earth's vertical field at 400°C for lengths of time that varied between 15 and 48 hours, after which they were cooled in field-free space and their TCRM's were measured. Because the field was nullified during cooling to room temperature, this remanence is more accurately termed a partial TCRM or PTCRM. The same samples were then placed in a furnace at 580°C to 600°C for several minutes (the large ones for as long as 15 minutes, the smaller ones for 5 to 7 minutes), then quickly cooled in the same constant field to 400°C. Subsequently, the field was nullified, the samples were cooled to room temperature, and the magnetization (PTCRM between 400°C and the Curie temperature) was measured. In Fig. 2 the graph of PTCRM as a function of partial TRM (PTRM) for each disk is shown. The 45° line represents what would be a perfect 1 : 1 correlation between PTCRM and PTRM.

Departure of points from the line may be related to inadequacies in our experimental methods. Points significantly above the line probably represent samples in which oxidation during TRCM inducement was incomplete. Additional oxidation to magnetite at the elevated temperatures during subsequent development of PTRM would then increase the total saturation moment. Points significantly below the line appear to be associated with samples in which oxidation to magnetite during TCRM acquisition was essentially complete and some magnetite oxidized to hematite during subsequent heating (for development of PTRM), thereby reducing the saturation moment considerably.

In another experiment a powdered sample was heated in the Curie balance at 400°C for 44 hours and the saturation magnetization was determined as a function of heating time. As shown in Fig. 3, there are two portions of the curve: a very rapid increase in J_s during

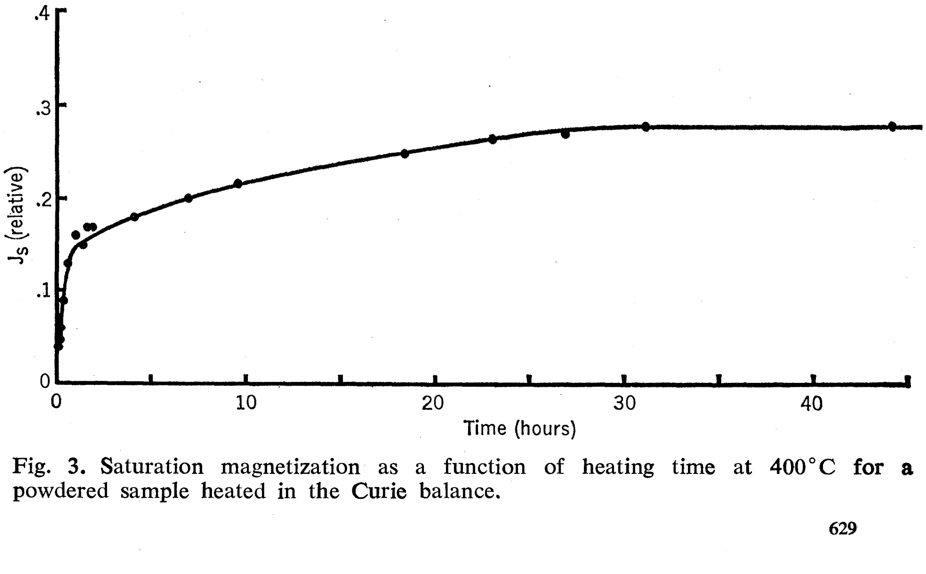


Fig. 3. Saturation magnetization as a function of heating time at 400°C for a powdered sample heated in the Curie balance.

the first hour, followed by a more gradual increase in saturation magnetization. The transition is more abrupt than might be expected on the basis of a simple exponential decrease; we do not fully understand the physical reasons for this sudden change in rate. As noted earlier, TCRM at 400°C also increased rapidly, then tailed off abruptly. Day *et al.* (*10*) have postulated two stages in the oxidation of a titanomagnetite: (i) the conversion of octahedral Fe^{2+} to Fe^{3+}, followed by (ii) the migration of Ti^{4+} from the spinel lattice to the rhombohedral ilmeno-hematite phases, which results in a more iron-rich magnetite phase. Perhaps these two successive steps are being reflected in our heating experiments.

One of the primary assumptions inherent in any paleointensity study of a rock is that the TRM developed in the laboratory is being compared with a TRM acquired in nature at the time the rock cooled through the Curie temperature (*11*). However, if the original remanent magnetization was induced at temperatures below the final Curie temperature by the oxidation process outlined here, then serious questions are raised as to the validity of Thellier's method.

Our primary conclusion is that for the basalt we studied the remanence is proportional to the field intensity, whether it is simple TRM or whether it is wholly or partially TCRM. On the assumption that the New Mexico basalt is not magnetically unique (and there seems little reason to suspect that it is), we further tentatively conclude that the presence of TCRM does not invalidate paleointensity studies.

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Convection in the Antarctic Ice Sheet Leading to a Surge of the Ice Sheet and Possibly to a New Ice Age

Abstract. *The Antarctic surge theory of Pleistocene glaciation is reexamined in the context of thermal convection theory applied to the Antarctic ice sheet. The ice sheet surges when a water layer at the base of the ice sheet reaches the edge of the ice sheet over broad fronts and has a thickness sufficient to drown the projections from the bed that most strongly hinder basal ice flow. Frictional heat from convection flow promotes basal melting, and, as the ice sheet grows to the continental shelf of Antarctica, a surge of the ice sheet appears likely.*

Wilson (*1*) has suggested that Pleistocene ice ages could be triggered by a surge of the Antarctic ice sheet. He makes the following basic assumptions: (i) the periodicity of the surge is controlled by variations in solar input to the Antarctic ice sheet due to precession of the earth's rotational axis; (ii) a surge begins when the entire ice sheet is underlain by a basal water layer; (iii) the surge creates an ice shelf extending to the Antarctic convergence, causing an increase in the earth's albedo sufficient to preserve the winter snow cover in high latitudes.

Weertman (*2*) gave a quantitative treatment of Wilson's assumption (ii) and concluded that, before a surge could occur, the basal water layer must exceed a "controlling obstacle size" related to bed roughness and the basal shear stress. Once a surge began, however, it could discharge sufficient ice into the Antarctic Ocean to satisfy Wilson's assumption (iii), but as icebergs rather than as a stable ice shelf.

When our estimates of the controlling obstacle size are compared with our estimates of the basal water layer thickness due to gravity sliding and thermal convection, it appears that thermal convection is necessary to make the Antarctic ice sheet surge. The possibility of convection was realized (*3*) when the CRREL core hole drilled to bedrock at Byrd Station, Antarctica (80°1′S,119°31′W), indicated a density inversion 1000 m below the surface (*4*). This density inversion will be confirmed when the compressibility coefficient is definitely established.

An estimate of the controlling obstacle size can be made from the basal sliding velocity U_x'', which is the measured downstream velocity at the surface of the ice sheet, U_x, minus the downstream velocity due to differential motion within the ice, U_x'. On the assumption that laminar flow dominates, U_x' can be estimated from the flow law of ice (*5*):

$$\dot{\gamma} = A(f,T,P)\ \tau^n \qquad (1)$$

where $\dot{\gamma}$ is the effective shear strain rate, τ is the effective shear stress, n is a viscoplastic parameter, and $A(f,T,P)$ is a function of ice fabric f, temperature T, and hydrostatic pressure P. For laminar flow

$$\tau = \tau_{xz} = \rho\ g\ (h - z) \tan \theta \qquad (2)$$

where ρ is ice density, g is the acceleration of gravity, θ is the surface slope, x is horizontal and positive downstream, z is vertical with $z = h$ at the surface, and the origin of the orthogonal x,y,z coordinates is placed on the bottom of the ice sheet.

Since $A(f,T,P) = A(z)$, we can combine Eqs. 1 and 2 to obtain

$$U_x' = \int_0^h \dot{\gamma}_{xz}\, dz = \int_0^h A(z)\ \tau_{xz}{}^n\, dz = (\rho\ g \tan \theta)^n \int_0^h A(z)\ (h - z)^n\, dz \qquad (3)$$

where h is the total ice thickness at distance x from the ice divide, and

$$A(z) = A(f,T,P) = B(f) \exp{(-Q/RT)} = B(f) \exp{(-C\ T_M/T)} \qquad (4)$$

where Q is the activation energy of self-diffusion, T_M is the pressure melting point, (T/T_M) is the homologous temperature, and C is a constant.

Figure 1 shows the quantitative z dependence of T, (T/T_M), and ρ, and the qualitative z dependence of f at Byrd Station. For a graphical solution of Eq. 3 from Fig. 1:

$$U_x' = \Sigma_h\ \Delta U_x' = \Sigma_z\ \overline{A}\ (\rho\ g \tan \theta)^n\ (h - \bar{z})^n\ \Delta z \qquad (5)$$

where

$$\Delta U_x' = (U_x')_{z_2} - (U_x')_{z_1}$$
$$\overline{A} = \tfrac{1}{2}\ (A_{z_2} + A_{z_1})$$
$$\bar{z} = \tfrac{1}{2}(z_2 + z_1)$$
$$\Delta z = z_2 - z_1$$