References and Notes

1. R. M. Goldstein and R. L. Carpenter, *Science* **139**, 910 (1963).
2. R. M. Goldstein and W. F. Gilmore, *Science* **141**, 1171 (1963).
3. D. O. Muhleman, *Astron. J.* **67**, 191 (1962).

12 August 1963

Stress Differences and the Reference Ellipsoid

In a recent communication, Hulley (*1*) has connected gravity anomalies with other geophysical phenomena including faults and the pole positions. Unfortunately, the latter suggestion is not substantiated mathematically; for areas of any extent and for realistic rheology the *polfluchtkraft* can even be in the direction opposite that shown in Hulley's diagram (*2*). In this work, Hulley made use of diagrams of the contours of the geoid supplied by Kaula (*3*). The geoid contours to which Hulley refers do not give a clear picture of the distribution of the stress differences. This is because the reference ellipsoid is an approximation to the average ellipsoid. Stress differences, however, arise from the difference between the actual form of the earth and the theoretical one for fluid equilibrium. The flattening which corresponds to fluid equilibrium is approximately 1/300 as was pointed out by Henriksen (*4*) and later discussed by O'Keefe (*5*) and Munk and MacDonald (*6*). If we plot the values of the gravity anomalies referred to an ellipsoid with a flattening of 1/300, we get the result as shown in Fig. 1, which is based on Kaula's work. In comparison with Hulley's paper, Fig. 1 indicates that there may be a relation between the tectonic activity and gravity anomalies; at least the strong positive anomalies in the East Indian area appear to correspond with the maximum tectonic activity.

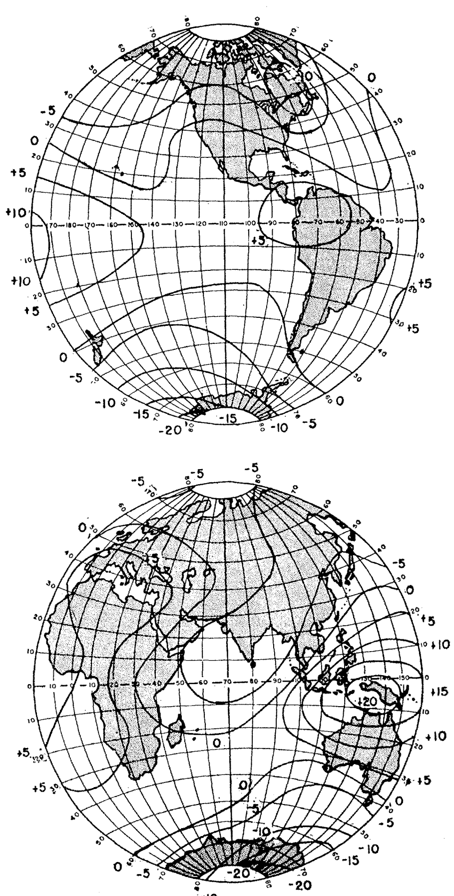

Fig. 1. Gravity anomalies, in milligals, derived from satellite perturbations and referred to an ellipsoid with a flattening of 1/299.8.

On the other hand, it should also be pointed out that there is a special explanation associated with the largest part of the discrepancy between the actual and equilibrium figures: the difference in oblateness can be considered as a lag of 10^7 years in adjustment to the slowing of the earth's rotation (*6*). So it is not entirely clear what the proper reference figure should be.

It is interesting to note that, regardless of the reference figure used, the shape of the geoid does not lend any particular support to the suggestion of Girdler (*7*) that the rift valleys and the mid-ocean ridges are the loci of up-currents in a convection system. It has been shown, by Licht (*8*) for example, that the top of a convection current should be in the area of positive gravity anomalies.

The positive anomaly areas near Central America, West Africa, and the East Indies are not associated with any ocean ridges. On the contrary, the ocean ridge system extending from the northwest Indian Ocean, around south of Australia, and up to the east Pacific is strongly correlated with a negative belt in the gravity field.

A similar negative correlation exists between heat flow and the gravity field, as shown by Lee and MacDonald (*9*), whose harmonic analysis of thermal measurements shows areas of maximum heat flow in central Asia and the eastern Pacific, and areas of minimum heat flow in the south Atlantic and western Pacific.

The various correlations shown are suggestive of what hypotheses to pursue, but they undoubtedly have a strong subjective element, and need both firmer mathematical models and more extensive data: in particular, more widespread gravimetry.

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References and Notes

1. J. L. C. Hulley, *Nature* **198**, 466 (1963).
2. A. H. Scheidegger, *Principles of Geodynamics* (Springer, Berlin, ed. 2, 1963), p. 170.
3. W. M. Kaula, *J. Geophys. Res.* **64**, 2401 (1959); **66**, 1799 (1961); **68**, 473 (1963).
4. S. W. Henriksen, *Ann. Intern. Geophys. Yr.* **12**, 197 (1960).
5. J. A. O'Keefe, *J. Geophys. Res.* **64**, 2389 (1959).
6. W. H. Munk and G. J. F. MacDonald, *ibid.* **65**, 2169 (1960).
7. R. W. Girdler, *Nature* **198**, 1037 (1963).
8. A. L. Licht, *J. Geophys. Res.* **65**, 349 (1960).
9. W. H. C. Lee and G. J. F. MacDonald, *ibid.*, in press.

9 September 1963

Geomagnetic Polarity Epochs: Sierra Nevada II

Abstract. *Ten new determinations on volcanic extrusions in the Sierra Nevada with potassium-argon ages of 3.1 million years or less indicate that the remanent magnetizations fall into two groups, a normal group in which the remanent magnetization is directed downward and to the north, and a reversed group magnetized up and to the south. Thermomagnetic experiments and mineralogic studies fail to provide an explanation of the opposing polarities in terms of mineralogic control, but rather suggest that the remanent magnetization reflects reversals of the main dipole field of the earth. All available radiometric ages are consistent with this field-reversal hypothesis and indicate that the present normal polarity epoch (N1) as well as the previous reversed epoch (R1) are 0.9 to 1.0 million years long, whereas the previous normal epoch (N2) was at least 25 percent longer.*

A recent paleomagnetic investigation (*1*) of six radiometrically dated igneous rocks of late Pliocene and Pleistocene age from the Sierra Nevada of California led to the conclusion that if the polarity epochs of the earth's magnetic field are equal or nearly equal in length, then they are either ½ or 1 million years long. Normal polarity epochs are defined in terms of the geomagnetic field-reversal hypothesis

(2) as intervals when the earth's field was directed toward the north and inclined below the horizontal in the northern hemisphere as it is at present; reversed polarity epochs are times when the field was directed toward the south and inclined above the horizontal in the northern hemisphere. Epochs of equal or nearly equal length were originally suggested by Khramov's (3) investigation of late Pliocene and Pleistocene sedimentary deposits near the Caspian Sea, in which sequences of strata with normal remanent magnetization alternate with sequences of reversely magnetized strata of equal or nearly equal thickness.

For geomagnetic polarity epochs of either ½ million or 1 million years' duration, the geomagnetic field would be normal for the past ½ million years and from 2 to 2½ million years, and reversed from 1½ to 2 million years ago. The magnetic polarities are different for the two time scales between ½ and 1½ and between 2½ and 3½ million years ago; however, none of the radiometrically dated rocks available for inclusion in the earlier report were unambiguously within these time ranges, so that it was not possible to decide whether the late Pliocene and Pleistocene polarity epochs were approximately ½ or 1 million years in length.

A subsequent study (4) of the normally magnetized basalt in Olduvai Gorge, Tanganyika, is consistent with both time scales, provided the date of the 2-million year polarity transition is reduced to 1.8 ± 0.1 million years, the exact figure depending on the age of the basalt. We here report paleomagnetic and radiometric data from ten additional extrusive rock units from the Sierra Nevada which clearly indicate that the more recent geomagnetic polarity epochs have been about 1 million years long but have not all been of exactly the same length. All of the results from the Sierra Nevada and Tanganyika investigations are shown in Fig. 1, along with three determinations from Europe reported previously by Rutten (5).

Investigation of geomagnetic polarity epochs, which involves analyzing the remanent magnetization acquired by rocks of known age at the time they were formed, is complicated by the phenomenon of mineralogically controlled self reversal, a process whereby certain ferromagnetic minerals may become reversely magnetized in a normal magnetic field (6). Because self reversal is specific to certain minerals such as pyrrhotite and ferrian ilmenite, and is sensitive to cooling rates, recognition of self reversals is aided by multiple sampling from each rock unit so as to include specimens which cooled at different rates and which contain different ferromagnetic minerals. Wherever possible this was done in the current study. A search was made for known self-reversing minerals by means of microscopic examination of polished sections, supplemented by thermomagnetic measurements for which a continuously recording thermomagnetic balance was used. In addition, the mode of decay of the natural remanent magnetization when heated by increments in a field-free space was determined for specimens from all of the units. Specimens from all of the rock units were also partially demagnetized in alternating magnetic fields, the remanent magnetization of each specimen being remeasured after demagnetization in peak alternating fields of 12.5, 25, 50, 100, 200, 400, 600, and 800 oersteds. The purpose of these experiments was to determine whether the original magnetization had been changed through the superposition of a later soft component of magnetization due to lightning strikes, or to long exposure after cooling to the weak magnetic field of the earth.

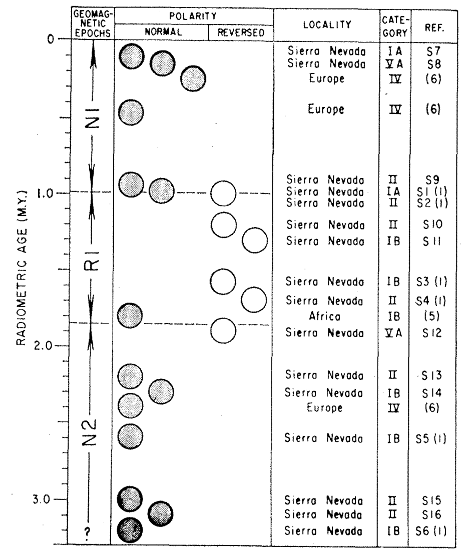

Fig. 1. Time scale for geomagnetic polarity epochs. Categories IA and IB designate determinations where there is evidence that self reversal has not occurred. For category II laboratory experiments indicate that self reversal is unlikely. For categories IV, VA, and VB, there either is evidence relevant to the possibility of self reversal or it is ambiguous. (Further details in text.)

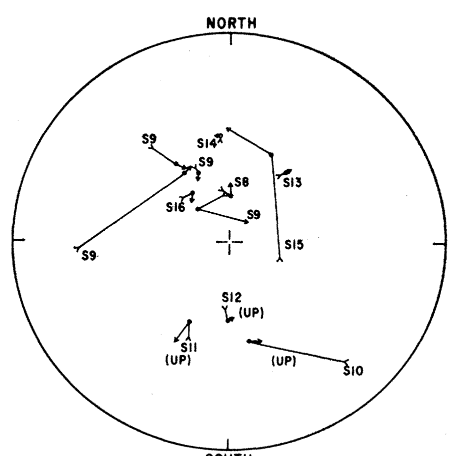

Fig. 2. Changes in direction of magnetization of typical specimens plotted on Schmidt projection. Original direction of magnetization (tail of arrow), direction after demagnetization in 400-oersted peak field (dot), and direction after demagnetization in 800-oersted field (head of arrow), are shown. All points are on lower hemisphere except for S *10*, *S 11*, and *S 12*. (Three specimens from *S 9* are shown.)

For some rocks it could be established with reasonable certainty that self reversal had not occurred. For others, however, the experimental results were difficult to interpret unambiguously, because of the incomplete state of our knowledge of the ways in which self reversal may occur and the ways in which it can be detected. To provide a basis for assigning varying degrees of reliability to paleomagnetic studies of polarity, we have divided our results into categories.

IA. The direction of remanent magnetization in a xenolithic inclusion or contact zone baked by an igneous rock is the same as that in the igneous rock body. Self reversal is highly unlikely.

IB. Hematite and magnetite or titanomagnetite coexist in the same rock and the direction of magnetization remains the same before and after heating to 600°C, this temperature being sufficient to destroy the remanent magnetization of the magnetite or titanomagnetite. Concordance between the directions of magnetization of these two ferromagnetic minerals makes self-reversal very unlikely.

II. The following criteria are satisfied (1): Partial demagnetization experiments in alternating fields have removed any soft components of magnetization acquired after the rock was formed. Typical results from such experiments are shown in Fig. 2 (2).

When heated and cooled in the present normal geomagnetic field, the rock becomes normally magnetized (*3*). When heated by increments in zero magnetic field, the curve of magnetic intensity plotted against temperature maintains a negative slope and decreases monotonically to zero, as shown in Fig. 3, indicating that only one magnetic constituent is probably present. Self reversal is unlikely, but the evidence is less strong than for categories IA and IB.

III. Accurate measurements have been made of the direction and intensity of the remanent magnetization of multiple specimens from the same rock unit, and demagnetization in alternating fields has been used to remove soft components which may mask the original magnetization. Measurements in this category are certain to have identified correctly the polarity of the hard component of magnetization in the rock, but offer no evidence relevant to self reversal. (None of the determinations in the present study are in this category.)

IV. The only measurements available are polarity determinations made with a compass, a portable magnetometer, or with some other semiquantitative instrument. Measurements in this category are somewhat less certain in identifying the polarity of the stable remanent magnetization of rocks, and offer no evidence relevant to self reversal unless baked zones or inclusions are also studied.

VA. Criteria (1) and (2) of category II are satisfied; however the curve of intensity of magnetization plotted against the temperature in zero field first decreases with increasing temperature, then increases, then decreases to zero as the temperature rises, as shown in the example (Fig. 3). This suggests (but does not prove) that two magnetic constituents with different Curie temperatures are interacting, and this in turn may reflect a tendency toward self reversal. The interpretation of these experimental results is ambiguous.

VB. Identical with VA, except that on heating to higher fields the polarity reverses. Self reversal is more likely than for VA, but again it is uncertain whether the original remanent magnetization of the rock is self reversed.

VI. Rock is reproducibly self reversing. Self reversal is likely, and the direction of the remanent magnetization may be interpreted as indicating a geomagnetic polarity epoch of opposite polarity. (None of the determinations in the present study are in this category.)

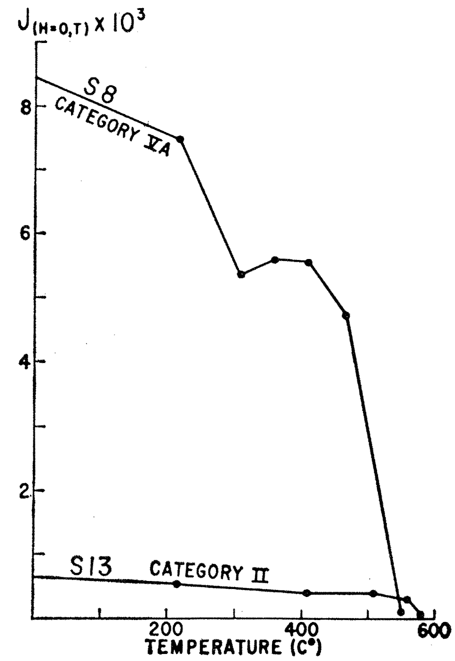

Fig. 3. Mode of thermal decay of natural remanent magnetization (NRM). $J_{(H=0,\ T)}$ is the NRM remaining after heating to temperature T and cooling in zero magnetic field. Specimen from *S 13* illustrates category II, that from *S 8* is typical of the rocks in category VA. The differences in magnetic intensity, which are unrelated to categories, are typical of flow-to-flow differences.

The radiometric ages of the rocks in this study were determined by the potassium-argon method. The isotope dilution method and a Reynolds-type mass spectrometer were used in the argon analyses; potassium values were determined with a Perkin-Elmer flame photometer, lithium being used as an internal standard. Decay constants used for the calculation of the ages are $\lambda_e = 0.585 \times 10^{-10}$ yr^{-1} and $\lambda_\beta = 4.72 \times 10^{-10}$ yr^{-1}.

In the following paragraphs, which summarize the results of the paleomagnetic and radiometric data, the locality listed for each rock unit, together with the township and range designation, serve to identify it in the references cited, where more complete geologic information is given.

S 7. Basalt in Sawmill Canyon (*7, 8*). *Age*: (a) 0.09 ± 0.09 million years. Whole rock, wt. = 19.01 g, %K = 1.37, $\%Ar_{40}^{at}$ = 99.35, $Ar_{40}^{r}/K_{40} = 0.053 \times 10^{-4}$; (b) 0.06 ± 0.05 million years (argon rerun from same sample). Wt. = 13.33 g, $\%Ar_{40}^{at}$ = 99.43, $Ar_{40}^{r}/K_{40} = 0.033 \times 10^{-4}$. *Location*: Mt. Pinchot 1:62,500 quadrangle (1953), in canyon of Sawmill Creek in sec. 7 and 18, T12S, R34E, and sec. 12 and 13, T12S, R33E. *Paleomagnetic sampling*: 8 samples of basalt spanning 30 meters, 2 samples of granite baked by basalt, 2 unbaked granite samples. *Petrography*: dense holocrystalline porphyritic basalt containing phenocrysts of olivine and clinopyroxene. Only opaque mineral is titanomagnetite with a bimodal size distribution; abundant euhedral grains 2 to 10 μ in diameter, and less abundant irregular skeletal grains from 20 to 40 μ in diameter. No intergrowths observed. T_c (Curie temperature): 220°C. *Remanent magnetization*: normal. Baked granite is magnetized in same direction as basalt, unbaked granite is not. Category IA.

S 8. Basalt near Sonora Pass (*9*). *Age*: 0.150 ± 0.025 million years. Whole rock, wt. = 14.76 g, %K = 1.47, $\%Ar_{40}^{at}$ = 96, $Ar_{40}^{r}/K_{40} = 0.090 \times 10^{-4}$. *Location*: Dardanelles Cone 1:62,500 quadrangle (1956), south ½ of sec. 29 and southeast ¼ of sec. 30, T6N, R20E. *Paleomagnetic sampling*: 8 samples spanning 400 m laterally. *Petrography*: relatively coarsegrained, feldspathic diktytaxitic porphyritic andesite with abundant olivine phenocrysts. Only opaque mineral is titanomagnetite occurring both as coarsely skeletal to completely developed crystals 10 to 70 μ in diameter, and as rims 1 to 2 μ thick on the margins of mafic grains. T_c: 400°C and 550°C. *Remanent magnetization*: normal. Category VA.

S 9. Basalt at Devil's Postpile (*10*). *Age*: 0.94 ± 0.16 million years. Plagioclase, wt. = 8.76 g, %K = 0.22, $\%Ar_{40}^{at}$ = 95, $Ar_{40}^{r}/K_{40} = 0.547 \times 10^{-4}$. *Location*: Devil's Postpile 1:62,500 quadrangle (1953); ¼ mi east of Devil's Postpile National Monument and 3,100 ft S, 35 E from bench mark 7559. *Paleomagnetic sampling*: 8 samples spanning 140 m. *Petrography*: dense porphyritic basalt with abundant plagioclase phenocrysts. Titanomagnetite occurs as coarsely skeletal, nearly equant grains 20 to 70 μ in diameter and as thin rims around mafic silicate grains, both habits being equally abundant; intergrowths are very uncommon, but some of the rim-forming titanomagnetite is partially oxidized to hematite. Ilmenite occurs as elongate skeletal crystals 15 to 45 μ long. T_c: 440°C and 500°C. *Remanent magnetization*: normal. Category II.

S 10. Olivine latite near Bald Mountain (*11*). *Age*: 1.2 ± 0.1 million years. Whole rock, wt. = 9.89 g, %K = 1.61, $\%Ar_{40}^{at}$ = 51, $Ar_{40}^{r}/K_{40} = 0.727 \times 10^{-4}$. *Location*: Truckee 1:62,500 quadrangle (1955), south ½ of sec. 28, T17N, R16E. *Paleomagnetic sampling*: 6 samples spanning 40 m laterally. *Petrography*: dense porphyritic holocrystalline latite with abundant olivine phenocrysts. Titanomagnetite grains range in size from 20 to 120 μ, the smaller grains being euhedral and commonly oxidized, larger grains being coarsely skeletal and not martitized. Ilmenite intergrowths are common except where titanomagnetite grains are inclusions in olivine. T_c: 520°C. *Remanent magnetization*: reversed. Category II.

S 11. Olivine latite near Hirshdale (11). Age: 1.3 ± 0.1 million years. Whole rock, wt. = 8.80 g, %K = 2.53, $\%Ar_{40}^{at}$ = 29, Ar_{40}^{r}/K_{40} = 0.733 × 10^{-4}. *Location*: Truckee 1:62,500 quadrangle (1955), eastern ½ of sec. 32, T18N, R17E. *Paleomagnetic sampling*: 7 samples spanning 70 m. *Petrography*: vesicular porphyritic holocrystalline latite with abundant olivine phenocrysts. Titanomagnetite occurs as euhedral and subhedral grains 5 to 120 μ in diameter. Hematite is ubiquitously associated with the titanomagnetite both as coarse lamellae along the planes and as extensive martite replacement. Ilmenite occurs as blades 25 to 75 μ long; only a few grains show very fine lamellae of hematite. T_c: 540°C and 675°C. *Remanent magnetization*: reversed. Category IV.

S 12. Olivine latite near Tahoe City (11). Age: 1.9 ± 0.1 million years. Whole rock, wt. = 9.68 g, %K = 1.88 $\%Ar_{40}^{at}$ = 14, Ar_{40}^{r}/K_{40} = 1.10 × 10^{-4}. *Location*: Tahoe 1:62,500 quadrangle (1955), north ½ of sec. 12, T15N, R16E. *Paleomagnetic sampling*: 7 samples spanning 105 m. *Petrography*: slightly vesicular porphyritic pilotaxitic latite containing olivine phenocrysts and some interstitial glass. Titanomagnetite grains are 20 to 100 μ in diameter and occur in forms varying from euhedral to highly skeletal. Ilmenite is abundant as skeletal elongate crystals 20 to 200 μ long. No intergrowths occur either in the titanomagnetite or the ilmenite. T_c: 390°C and 540°C. *Remanent magnetization*: reversed. Category VA.

S 13. Andesite at Watson's Creek (8, 12). Age: 2.2 ± 0.1 million years. Whole rock, wt. = 10.69 g, %K = 1.69, $\%Ar_{40}^{at}$ = 14, Ar_{40}^{r}/K_{40} = 1.27 × 10^{-4}. *Location*: Tahoe 1:62,500 quadrangle (1955), southeast ¼ of sec. 21, T16N, R17E. *Paleomagnetic sampling*: 6 samples spanning 11 m. *Petrography*: dense fine-grained holocrystalline feldspathic andesite with sparse hypersthene phenocrysts. Euhedral magnetite or titanomagnetite grains are 5 to 50 μ in diameter; a very few grains show minor marginal oxidation to hematite, and a few grains have very coarse ilmenite intergrowths. Ilmenite also occurs as elongate grains 10 to 40 μ long, a very few of which contain hematite lamellae. T_c: 570°C. *Remanent magnetization*: normal. Category II.

S 14. Andesite at Alder Hill (11). Age: 2.3 ± 0.1 million years. Whole rock, wt. = 9.26 g, %K = 2.11, $\%Ar_{40}^{at}$ = 25, Ar_{40}^{r}/K_{40} = 1.34 × 10^{-4}. *Location*: Truckee 1:62,500 quadrangle (1955), south ½ of sec. 3, T17N, R16E, and sec. 35, T18N, R16E. *Paleomagnetic sampling*: 4 samples spanning 700 m laterally. *Petrography*: very feldspathic vesicular andesite with hypersthene phenocrysts. Magnetite occurs as equant grains 10 to 50 μ in diameter which have been extensively replaced by hematite, especially near vesicles; the replacement hematite occurs as lamellae of varying coarseness, as rims of irregular thickness, and as irregular patches. Ilmenite occurs as elongate grains 15 to 45 μ long which commonly have hematite intergrowths, especially near vesicles. T_c: 570°C and 675° C. *Remanent magnetization*: normal. Category IB.

S 15. Quartz latite at Two Teats Mountain (8, 13). Age: 3.0 ± 0.1 million years. Plagioclase, wt. = 7.08 g, %K = 0.77, $\%Ar_{40}^{at}$ = 69, Ar_{40}^{r}/K_{40} = 1.76 × 10^{-4}. *Location*: Devil's Postpile 1:62,500 quadrangle (1953), on Two Teats Mountain. Stratigraphically overlies *S 16*. *Paleomagnetic sampling*: 4 samples spanning 300 m. *Petrography*: vesicular hornblende biotite andesite with phenocrysts of andesine. Magnetite occurs as equant grains 5 to 320 μ long. Several grains have platey inclusions of pyrrhotite, and one grain has thin intergrowths of hematite. T_c: 565°C. *Remanent magnetization*: normal. Category II.

S 16. Basalt at San Joaquin Mountain (8, 13). Age: 3.1± 0.1 million years. Whole rock, wt. = 9.81 g, %K = 2.62, $\%Ar_{40}^{at}$ = 5, Ar_{40}^{r}/K_{40} = 1.84 × 10^{-4}. *Location*: Devil's Postpile 1:62,500 quadrangle (1953), near San Joaquin Mountain. Stratigraphically underlies *S 15*. The sample used for age analysis was separated from the paleomagnetic samples by approx. 5 mi, but all samples were collected from a horizontal sequence of basalt flows which is exposed continuously for a distance of about 6 mi and reaches a thickness of approximately 1,500 ft. The sample used for age analysis most nearly corresponds to the lowest of the paleomagnetic samples. *Paleomagnetic sampling*: 22 samples from 3 basalt flows spanning a stratigraphic thickness of 530 m. *Petrography*: specimen 2C577 from upper flow is vesicular porphyritic hypersthene andesite. Titanomagnetite grains 5 to 30 μ in diameter are extensively oxidized to hematite; commonly contain large ilmenite blebs. Ilmenite also occurs as grains 15 to 40 μ long. Specimen 2C595 from bottom flow is a dense, fine-grained trachyandesite with olivine and augite phenocrysts. Titanomagnetite occurs as subhedral crystals 5 to 60 μ in diameter; no alteration was observed, and the only intergrowths are rare, large granules of ilmenite and rare plates exolved along (111) in magnetite. T_c: 2C577, 550°C and 2C593, 330°C but changing during heating. *Remanent magnetization*: all specimens are normal. Category II for the upper flow and VB for the bottom flow.

Although many different types of volcanic rocks were included in this and our earlier investigation, we have been unable to discover any petrographic, mineralogic, or magnetic property which correlates with the polarity of the remanent magnetization. The only property of the rocks which appears to be systematically related to their magnetic polarity is their age, as determined by the potassium-argon method. Because of the global nature of the geomagnetic field, data from all continents must fit a common time scale if geomagnetic field reversal is the true explanation of reversely magnetized rocks. The data presently available from three continents fit the time scale shown in Fig. 1 within the precision of the dates.

Additional polarity epochs with lengths of the order of 10^5 years or less might be inserted at several points where gaps exist in the present radiometric age data, although the investigation of sediments by Khramov (*3*) suggests that such extremely short polarity epochs either do not exist or else are confined to times when no sedimentation occurred, which is unlikely. Keeping in mind the need for additional radiometric and paleomagnetic results to fill in the gaps in the present time scale, the following conclusions are the simplest ones which fit all of the data presently available: (i) The internal consistency of the paleomagnetic results and the radiometric ages confirms the hypothesis of geomagnetic fields reversal. (ii) The three most recent polarity epochs are probably not all of equal length (Fig. 1); the epoch in which we live, N1, and the previous reversed epoch, R1, are both 0.9 to 1.0 million years long, but N2 may be longer by 25 percent or more. (iii) Among the ferromagnetic minerals included in these studies, self reversals are rare or absent (*14*).

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References and Notes

1. A. Cox, R. R. Doell, G. B. Dalrymple, *Nature*, **198**, 1049 (1963).
2. M. Matuyama, *Proc. Imp. Acad. Japan*, **5**, 203 (1929); J Hospers, *Koninkl. Nederlandse Akad. Wetenschap. Proc. Ser. B* **57**, 112 (1954).
3. A. N. Khramov, *Akad. Nauk SSSR Doklady Izv. Ser. Geol.* **112**, 849 (1957); English Transl. Consultants Bureau, 1958.
4. C. S. Grommé and R. L. Hay, *Nature*, in press.
5. M. G. Rutten, *Geol. Mijnbouw* **10**, 373 (1959).
6. S. Uyeda, *Japan. J. Geophys.* **2**, 1 (1958).
7. A. Knopf, *U.S. Geol. Surv. Profess. Papers* (1918), No. 110, p. 130.
8. G. B. Dalrymple, thesis, Univ. of California (1963), p. 107.
9. D. B. Slemmons, thesis, Univ. of California (1953), p. 201.
10. F. E. Matthes, *U.S. Geol. Surv. Profess. Papers* (1960), No. 329, p. 62.
11. P. W. Birkland, thesis, Stanford Univ. (1961), p. 126; ———, Guidebook, annual field trip of the Geological Society of Sacramento, and the Sacramento section of the California Association of Engineering Geologists (1962), p. 59.
12. W. Lindgren, *U.S. Geol. Surv. Geol. Atlas* (1897), Folio 39.
13. H. D. Erwin, *Calif. J. Mines Geol.* (1934), Rept. 30, p. 7.
14. Publication authorized by director, U.S. Geological Survey.

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