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R. S. Sternberg

Robert F. Butler
University of Portland, butler@up.edu

AN ARCHAEOMAGNETIC PALEOINTENSITY STUDY OF SOME HOHOKAM POTSHERDS FROM SNAKETOWN, ARIZONA

R.S. Sternberg and R.F. Butler

Department of Geosciences, University of Arizona, Tucson, Arizona 85721

Abstract. A paleointensity study on nine potsherds from the Hohokam Indian site of Snaketown, Arizona is described. The sherds range in age from A.D. 200-1400. Examination of different temperature subintervals from the Thellier-Thellier double heating experiment reveals that conventional statistical measures sometimes can unambiguously determine the best data subset for paleointensity calculations. However, it is often necessary to visually inspect the data and utilize physical insight in determining this data subset. Results suggest that the paleointensity was about 0.94 F_0 (F_0, present intensity ≈ 0.506 oe) at A.D. 200, 0.72 F_0 at A.D. 600, and 1.2 F_0 at A.D. 1400. The shape of our curve of paleointensity vs. age is congruent with a curve previously derived from other Snaketown artifacts, but our paleointensities are systematically lower by about 0.15 oe.

Introduction

The thermoremanent magnetization of aboriginal ceramic artifacts provides a tool by which the strength of the geomagnetic field during the last several thousand years can be studied. Paleointensity data are necessary for understanding the secular variation of the geomagnetic field and could be used for evaluating different secular variation models. These models bear on the nature of the geomagnetic dynamo. Secular variation data are also useful in studying relationships between the geomagnetic field and other parameters. It has been shown (Olsson, 1970) that a quasi-sinusoidal variation of the geomagnetic dipole moment with a period of about 10,000 years is responsible for long-term variations of atmospheric radiocarbon concentration. Relationships have been suggested between geomagnetism and climate, and better resolution of the paleomagnetic and paleoclimatic data would facilitate further study in this area (see, e.g., Chiu, 1974). Archaeomagnetic paleointensities would be useful in calibrating and corroborating paleointensities derived from the depositional remanent magnetization of sediments (Levi and Banerjee, 1976). Archaeomagnetic data also have several archaeological applications including the possibility for use as a dating technique (Aitken, 1974).

For this study, we have examined nine potsherds from the Hohokam Indian site of Snaketown, Arizona (111° 55' 3" W, 33° 11' 2" N). The sherds are from seven ceramic phases, each lasting about 200 years and covering a time span from A.D. 200-1400. The sherds have been dated through a combination of archaeological and physical dating methods (Haury, 1976).

Experimental Procedure and Data Reduction

Several samples were cored from each sherd and cast in plaster of paris. The disc-shaped samples have typical diameters of 12.3 mm, thicknesses of 6 mm, and weights of 1 g. A coordinate system was set up on each sherd so that all samples from a sherd are oriented relative to one another. This provides a means for checking the directional consistency of the magnetization for samples from the same sherd.

Our varient of the Thellier-Thellier paleointensity experiment is as follows: the sample is heated to a temperature, T, and held at that temperature for 15 minutes. After cooling in a laboratory field, F_{lab}, of known strength and direction, the magnetization of the sample is measured. The sample is then reheated to T and cooled in F_{lab}, but with the direction of the sample rotated by 180° with respect to the laboratory field. The magnetization is again measured. Vector addition and subtraction of the two measurements at the same T yield the natural remanent magnetization (NRM) remaining after heating to T and the partial thermoremanent magnetization (PTRM) gained by cooling from T in the presence of F_{lab}. All heatings are done in air. If NRM is plotted against PTRM (each point corresponding to a particular heating temperature), the result is a linear relationship if various conditions are met. The paleointensity F_e, is then given by: $F_e = -bF_{lab}$, where b is the (negative) slope of the data. The standard error $\sigma_{F_e} = \sigma_b F_{lab}$. We used $F_{lab} = 0.5$ oe throughout the experiment.

To verify the validity of our experimental procedures, two pottery samples and a red brick sample were heated to 720°C and cooled in a known laboratory field, thus producing a thermoremanent magnetization. The plots of "NRM" (actually laboratory TRM) vs. PTRM were linear for all three samples, and the directions of magnetization were stable. The experimentally determined field value of 0.74 ± 0.02 oe is not significantly different from the actual value of 0.77 oe used to produce the TRM.

For the paleointensity experiment, samples were generally heated from 100°C to 700°C at successive 50°C intervals. Most of our NRM-PTRM plots were linear over only part of the entire temperature range used. Possible causes of nonlinearity at low temperatures are viscous remanent magnetization, reheating of the pottery subsequent to manufacture, and lightning strikes. Non-linearity at high temperatures can be caused by an original firing temperature below the Curie point, possibly with a secondary chemical remanence; or by mineralogic change during the heating cycles of the experiment. It is not always immediately obvious which temperature interval should be used in calculating the paleointensity. Using two or more samples from the

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TABLE 1. Data for the Individual Samples

Sherd	Sample	NRM/g ($x 10^{-3}$)	T (°C)	n	F_e (oe)	σ	f	g	q
2	2	3.8	250-650	8	.60	.01	.867	.712	28.6
2	3	3.7	200-650	10	.60	.01	.892	.842	57.5
4	2	.92	0-600	11	.55	.01	.871	.836	43.1
4	4	.96	0-600	11	.50	.01	.841	.876	74.0
4	5	-	0-600	11	.52	.01	.855	.882	47.0
6	2	1.4	300-500	5	.33	.01	.531	.739	12.7
6	3	1.2	250-400	4	.40	.01	.332	.621	5.7
7	2	.53	300-650	7	.39	.02	.445	.783	7.3
7	3	.65	300-720	7	.43	.01	.622	.810	21.4
10	3	.65	350-600	5	.34	.004	.584	.498	22.3
10	4	.79	300-700	8	.37	.004	.734	.790	54.2
13	1	.93	350-550	4	.44	.02	.361	.656	6.6
13	2	.89	250-600	7	.34	.01	.662	.761	19.8
14	2	.90	250-500	5	.35	.02	.647	.711	9.1
14	3	1.7	300-450	4	.61	.02	.373	.658	8.3
17	3	1.7	300-500	5	.46	.01	.338	.725	8.8
17	4	2.3	300-600	6	.48	.01	.714	.770	29.1

NRM/g is the original NRM per unit mass (gauss-cm^3/g), T gives the temperature extremes of the data subset used for paleointensity determination, and n is the number of points used. F_e is the paleointensity and σ the standard error. The fraction of extrapolated NRM spanned by the data subset is f; g is the gap factor; and q is the overall quality factor (Coe et al., 1978).

same sherd is helpful in locating common temperature intervals where the data appear reliable and common inflection points which might indicate a transition from useful to extraneous data. Direction of magnetization with increasing temperature during the paleointensity experiment was also noted and used in selection of the appropriate data subset. The direction should be nearly constant over this temperature interval, indicating stable magnetization. The slope of the NRM-PTRM plot and hence the paleointensity is determined by least squares regression. Regression lines are often calculated under the assumption that all the error is present in the NRM measurements, while in reality, error exists in both the NRM and PTRM. York (1966, 1967) gives a general solution for the least squares problem that allows variable weighting of the ordinate and abscissa of each datum. Under the assumption that the standard deviation of the PTRM and NRM measurements are proportional to the strengths of the laboratory field and paleofield respectively, Coe et al. (1978) derive

weights of $W_{PTRM} = \frac{1}{\sigma^2_{PTRM}}$ and $W_{NRM} = \frac{1}{\sigma^2_{NRM}}$,

the same weights to apply to all points. These weights imply a least squares slope and standard

error of $b = \frac{\sigma_{NRM}}{\sigma_{PTRM}} [1 \pm (\frac{1 - r^2}{n})^{1/2}]$,

(Kermack and Haldane, 1950; York, 1966) where n is the number of points and r is the correlation coefficient between NRM and PTRM.

Next we need some criteria to determine which subset or temperature range of the data should be used for calculation of a paleointensity. For the 19 samples from our nine sherds, we considered 231 cases corresponding to different temperature intervals which appeared visually plausible for paleointensity determination. For each case, a paleointensity and standard error were calculated. Statistics calculated to aid in selection of the best case were: the relative error of the paleointensity (σ_{F_e}/F_e); the correlation coefficient between NRM and PTRM; and the Fisher z transformation (David, 1938) of the correlation coefficient ($z = .5 \ln \frac{1 + r}{1 - r}$) and the standard error of this z value ($\sigma_z = \frac{1}{\sqrt{n - 3}}$). The ratio z/σ_z is distributed approximately as the standard normal distribution and indicates the statistical significance of the correlation coefficient r. We are interested in cases which have small values for the relative error, a value of r close to -1.0 (linear NRM-PTRM plot), and large negative values of z/σ_z (high statistical significance). For comparison, we also calculated a quality factor (Coe et al., 1978) which depends on the relative error, the spacing between data points or gap factor, and the fraction of the total extrapolated NRM spanned by the chosen seg-

TABLE 2. Summary of Sherd Data

Sherd	Ceramic Phase	Epoch (A.D.)	$\overline{F}_e$ (oe)	σ	$\overline{F_e/F_0}$	σ	RDM (x 10^{25})	σ
2	Civano	1300-1450	.60	.002	1.18	.003	10.2	.03
4	Sacaton-Soho	1000-1200	.52	.01	1.02	.02	8.8	.2
6	Santa Cruz	700-900	.36	.03	.70	.07	6.1	.6
7	Santa Cruz	700-900	.42	.02	.83	.03	7.1	.3
	Epoch average	700-900	.41	.02	.80	.05	6.9	.4
10	Gila Butte	550-700	.37	.002	.73	.005	6.3	.04
13	Snaketown-Gila Butte	450-650	.37	.04	.72	.08	6.2	.7
14	Sweetwater-Snaketown	250-450	.48	.13	.94	.26	8.1	2.2
17	Estrella-Sweetwater	100-300	.48	.01	.94	.02	8.1	.2

Ceramic phases and corresponding epochs follow the chronology of Haury (1976). $\overline{F}_e$ is the weighted mean paleointensity for the sherd and σ is the standard error of the mean. $\overline{F_e/F_0}$ is the weighted mean ratio of paleointensity to present field strength (.506 oe). RDM is the reduced dipole moment in 10^{25} gauss-cm^3 for the present geomagnetic latitude of 41.3°N.

ment of the NRM-PTRM plot. Of the 19 samples, there were seven for which all three statistical measures were optimized for the same case (i.e., same temperature interval) and an unambiguous selection of the best case was possible. Six of these seven cases also would have been chosen based on the quality factor. For the seventh sample, the quality factor for the case optimizing the statistical parameters was just slightly less than the maximum value of the quality factor for that sample. For the other 12 samples, selection of the best cases was more difficult, and visual inspection of the NRM-PTRM diagrams and stereographic plots of the directions for all samples from the same sherd became key factors in case selection. Sherd color appears to be related to quality of paleointensity data. A uniformly reddish-brown, well-oxidized sherd is unlikely to undergo chemical change during the heating cycles of the paleointensity experiment and generally yields reliable paleointensity data.

Results and Discussion

Results for the cases selected for each sample are shown in Table 1. A weighted mean and standard error of this mean for each sherd are calculated with the weight of each sample equal to C^2/σ^2. Although it is usually assumed that $C = 1$, an independent estimate of its value can be made (van der Waerden, 1969) such that the equations for the weighted mean and its standard error are

$$\overline{X} = \frac{\Sigma\ (X/\sigma^2)}{\Sigma\ (1/\sigma^2)}$$

$$\sigma_{\overline{X}} = \sqrt{\frac{\Sigma\ [(X-\overline{X})/\sigma]^2}{(n-1)\ \Sigma\ (1/\sigma)^2}}$$

The equation for $\sigma_{\overline{X}}$ is a weighted dispersion about the weighted mean, and dispersion of the data is as significant as the individual σ values. This suggests the importance of measuring multiple samples from the same sherd. These equations reduce to the familiar forms for equally accurate data if all the σ values are indeed equal. Sherd averages are given in Table 2. The value used for the present day field at Snaketown is F_0 = .506 oe as interpolated from the U.S. total intensity map, epoch 1975.0 (Fabiano et al., 1976). Sherd averages are plotted (circles) in Figure 1. Error bars for sherd ages encompass the entire ceramic period which the sherd represents. For the period A.D. 700-900, the solid line representing the trend of our data is drawn through the weighted mean of the two sherds from this period. Data of Bucha et al. (1970) from Snaketown are also plotted (triangles). For these data, no errors were given for individual paleointensity determinations. Where multiple sample measurements were taken from a sherd, we have calculated unweighted sherd averages and standard errors. The dashed trend line for Bucha's data passes through the weighted average of the two sherds at A.D. 1000, and the unweighted average of the sherds at A.D. 1.

The shape of our secular variation curve of field intensity is congruent with that of Bucha et al., but our paleointensities are systematically lower by about 0.15 oe. Artifacts for both studies were dated using the same chronology, so that even though there may be errors in determining the absolute ages of the ceramic phases, the given dates for these two studies should be comparable. A possible reason for the systematic offset would be incorrect measurement of F_{lab}. This would linearly affect all paleointensity results. We measured F_{lab} (produced by a solenoid) with a digital magneto-

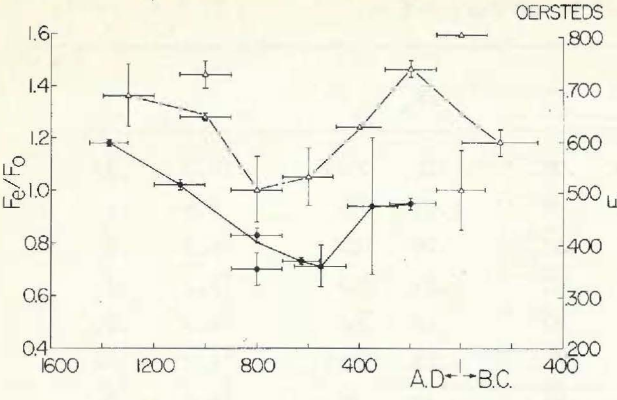

Fig. 1. Paleointensity (F_e) and ratio to present intensity (F_e/F_0) vs. age for Snaketown data. Triangles from Bucha et al. (1970), and circles from this work.

meter which was calibrated against the proton precession magnetometer at the U.S. Geological Survey Magnetic Observatory in Tucson and against test coils. We have been unable to find other errors in our experimental procedure or data reduction, so at this time we cannot isolate the source of discrepancy between the two curves. Although the data of Lee (1975) show no marked field minimum ca. A.D. 700, his two data for this time period may be from another area of the Southwest where there was no pronounced minimum. Also, Lee does not indicate how his dates were obtained. Presumably, our low paleointensities represent a non-dipole geomagnetic feature of restricted spatial extent. We hope to verify this in future studies on other southwestern U.S. artifacts.

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