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Paleomagnetism and Rock Magnetism of the Pliocene Rhyolite at San Vincenzo, Tuscany, Italy *

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Abstract. A paleomagnetic study of the Pliocene rhyolite near San Vincenzo, Italy, indicates that the counter-clockwise rotation of the Italian peninsula was complete by 4.7 m.y.B.P. Studies of IRM acquisition, low-temperature behavior, and AF demagnetization of both NRM and IRM show that the remanence is carried by very fine-grained magnetite, in the single domain, and possibly pseudo-single domain size range. All 10 sites are reversely magnetized, with an over-all mean direction of $D=160°$, $I=-45°$, $\alpha_{95}=6.6°$. This direction is 21° away from the reverse of the present axial dipole field direction. Part of the discrepancy may originate from incomplete averaging of the magnetic field, but it is likely that tectonic rotations are also involved. There are significant deviations of both declination and inclination, so the result cannot be explained by rotation of the peninsula about a vertical axis. No sedimentary rocks are exposed, so we cannot make a standard bedding correction, but a correction can be made by assuming that cooling joints in the rhyolite formed with a near-vertical orientation. When their common intersection line is restored to vertical, the inverse of the present axial dipole direction falls within the circle of confidence of the corrected mean direction for the rhyolite.

Key words: Paleomagnetism – Rock magnetism – Italy – Pliocene rhyolite.

Introduction

Paleomagnetic studies of the Jurassic, Cretaceous, and Paleocene sedimentary rocks of the Umbrian sequence of northern Peninsular Italy have consistently shown remanent declinations in the northwest quadrant (Lowrie and Alvarez, 1974; 1975; 1977b; Channell and Tarling, 1975; Klootwijk and VandenBerg, 1975; VandenBerg et al., 1978; Roggenthen and Napoleone, 1977; Alvarez and Lowrie, 1978; Channell et al., 1978). These results have generally been inter-

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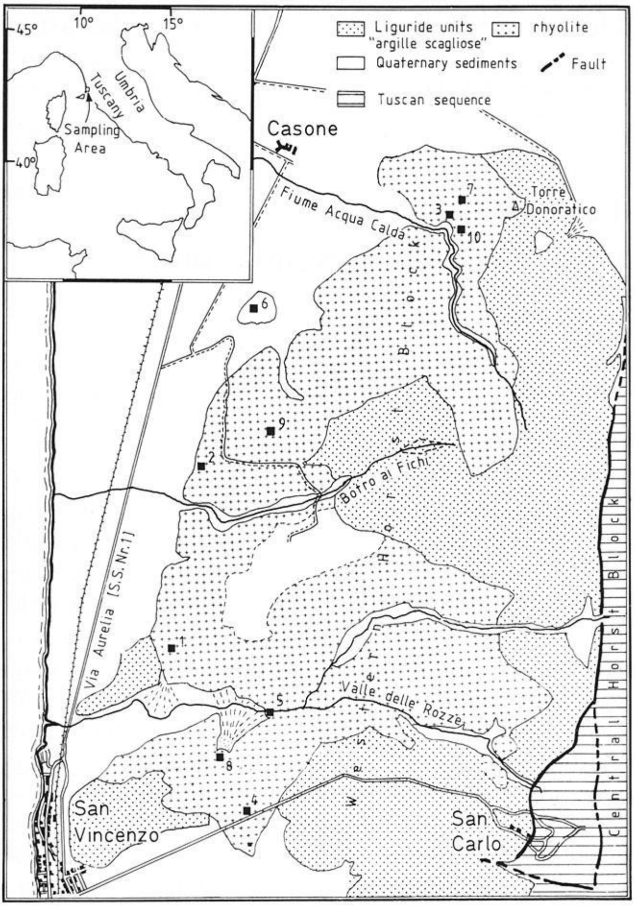

Fig. 1. Map showing the simplified geology of the sampling area (Giannini, 1955) and the locations of the ten paleomagnetic sites

preted as indicative of a counter-clockwise rotation of the Italian Peninsula. However, to accommodate the shortening represented by the folds of the Umbrian Apennines, the Umbrian sedimentary sequence is almost certainly detached at the level of the Triassic evaporites (Martinis and Pieri, 1964), so it is impossible to determine directly the rotational history of the basement, which does not outcrop. A further complication is the arcuate curvature of the Umbrian fold belt, which is reflected in the pattern of paleomagnetic declination (Channell, 1976; Channell et al., 1978).

Study of the thick stratigraphic sections at Gubbio and Moria (Lowrie and Alvarez, 1977a and b; Alvarez and Lowrie, 1978; VandenBerg et al., 1978) shows that some of the counter-clockwise rotation of the Umbrian sequence took place during the Cretaceous, but that further rotation must have occurred at some time after the Middle Eocene.

This Cretaceous rotation has been variously interpreted. Lowrie and Alvarez (1975) associated it with the rotation of the Italian peninsula on a microplate independently of Europe and Africa. Channell and Tarling (1975) disregarded Umbrian data as irrelevant to the rotation of Italy on the grounds that the Umbrian sequence is allochthonous. VandenBerg et al. (1978) disputed this assumption and interpreted Umbrian data in terms of rotation of the Italian peninsula with the African plate.

The subsequent, post-Middle-Eocene rotation of Italy has been attributed by VandenBerg (1979) to continent-continent collision between the Adriatic promontory of the African plate (Channell and Horvath, 1976) and the Balkan-Rhodope-Turkey continental block. Unfortunately, in Umbria the soft shales that continue the section upward cannot be sampled readily for paleomagnetic investigation.

Suitable Oligocene and Miocene rocks have not been located, but in the hope of placing an upper limit on the time of rotation, we have studied the paleomagnetism of the Pliocene rhyolite of San Vincenzo. This locality on the west coast of Italy, about 40 km south of Pisa, is almost opposite the isle of Elba and about 100 km west of the outcrop area of the Umbrian sequence (Fig. 1).

Geologic Setting

Tectonics of Tuscany

Orogenic activity in the Northern Apennines began in the Late Oligocene. The locus of most intense compressional deformation has shifted progressively from west to east, and orogenic activity probably continues at the present time near the Adriatic coast. As the locus of deformation shifted, compression was followed in a particular area by extensional tectonics. As a result our area of interest, on the Tuscan coast, was undergoing normal faulting during the Late Miocene and Pliocene, while the Umbrian area was still being folded. This phenomenon has been discussed by Elter et al. (1975).

In coastal Tuscany, Upper Miocene and Pliocene sediments are found in small basins interrupted by horsts in which the older, deformed rocks are exposed. Two major groupings of deformed rocks are recognized throughout coastal Tuscany. The lower unit is the Tuscan sequence, which closely resembles the Umbrian sequence farther east. The Tuscan sequence, like the Umbrian sequence, was deposited on Italian continental crust during the Mesozoic and Early Tertiary; it is now folded, detached from the underlying basement, and has been thrust a moderate distance – perhaps a few tens of kilometers – to the east. Tectonically overlying the Tuscan sequence are ophiolites and deep-water oceanic sedimentary rocks ranging from kilometer-scale slabs down to chaotic melange units. These 'Liguride units' are generally interpreted as having originated in the Mesozoic-Early Tertiary Tethyan Ocean, to the west of the Italian continent. During the Apennine orogeny the Tethys was consumed, and the Liguride units were transported 100 km or more toward the east.

From late Pliocene time to the present, coastal Tuscany has been rising relative to sea level. As a result, the horst blocks, which formerly were islands, have gradually been incorporated into the Italian mainland. The process is clear from the present morphology. Islands not yet joined to the mainland form the Tuscan Archipelago (Elba, Giglio, Mentecristo, etc.). Blocks recently joined are linked to the mainland by tombolo sand spits (Monte Argentario) or low sandy areas (Piombino). Blocks incorporated still earlier form the characteristic patchy, irregular mountains of coastal Tuscany. One of these small mountain groups is the Monti di Campiglia; the rhyolite of San Vincenzo occupies its western side, facing the sea (Fig. 1).

Local Geology

Giannini (1955) has made a detailed study of the Monti di Campiglia. Structurally the mountain group is a compound horst. The central block, tapering northward, is the highest. It exposes mainly the lower part of the sequence (Jurassic): only in the southernmost part are remnants of the Liguride units preserved ('argille e calcari'). A small body of granite and several quartz porphyry dikes cut the Tuscan sequence limestones. In the eastern block, the upper part of the Tuscan sequence (Cretaceous-Lower Tertiary) is exposed, with a more extensive cover of Liguride allochthon.

The western block shows only a small exposure of Tuscan sequence, adjacent to its fault contact with the higher central block. This is covered by Liguride allochthon, which forms the surface of about half of the western block. In 1955 the Liguride ensemble had not yet been subdivided, and Giannini mapped all of this material as *argille scagliose* (scaly clays). In more recent map compilations (Bortolloti et al., 1970; Giannini et al., 1971) this material is assigned to two different units. In the western half of the western block, the Liguride units are covered by the Pliocene rhyolite of San Vincenzo.

The San Vincenzo Rhyolite

Two young magmatic provinces are recognized in the northern Apennines (Marinelli and Mittempergher, 1966; Marinelli, 1967; Bortolotti and Passerini, 1970). The Tuscan Province includes a number of small bodies of granitic intrusive rocks and acid volcanics, and has yielded age dates ranging from 7 to 0.43 m.y. (Barberi et al., 1971). The much larger volcanic districts of the Latian Province are dominantly of potassic trachyte composition and are almost entirely younger than 1 m.y. (Locardi et al., 1975; Alvarez, 1975).

The volcanic rocks, dikes, and pluton of the Monti di Campiglia are part of the Tuscan Province. Their petrography and chemistry have been studied by Barberi et al. (1967), who conclude that all three types were derived from an anatectic magma of quartz-monzonite composition. Depending on the classification scheme, the volcanic rocks of San Vincenzo would be considered either rhyolites or quartz latites.

Borsi et al. (1967) have reported the following K/Ar age determinations on rocks from the Monti di Campiglia:

Granite: 5.7 ± 0.16 m.y. (orthoclase),
Granitic porphyry: 4.3 ± 0.13 m.y. (whole rock),
Pegmatitic vein: 5.0 ± 0.15 m.y. (phlogopite),
San Vincenzo volcanite: 4.7 ± 0.14 m.y. (biotite).

Barberi et al. (1967, p. 679) point out that the 4.3 m.y. date is from a metasomatically altered dike, and therefore gives only a minimum age for the time of emplacement.

The San Vincenzo rhyolite covers several square kilometers, but much of this is brushy country with poor outcrop; the best exposures are in quarries. Access is further hampered by the fact that the woods are on private farms and hunting estates, stocked with wild boar. We were able to find 10 sites with fairly large outcrops in quarries, road cuts and a stream bed (Fig. 1) and at these we drilled 100 samples for rock-magnetic and paleomagnetic evaluation. Because of the thick Quaternary sedimentary overburden and absence of underlying, overlying or interbedded sediments, local tectonic correction at each site was not possible.

The most striking feature of the rhyolite in outcrop is a layering or pseudostratification marked by planes of flattened vesicles with a centimeter-scale spacing (Fig. 2). We interpret the layering as marking planes of concentrated simple shear strain during flow of the rhyolite. This is supported by (1) the elongation of the vesicles, (2) the presence in some places (e.g., site CM-2) of microbrecciated zones along the planes of pseudostratification and (3) the occasional presence of inclusions which have clearly been rolled. In small outcrops the layering generally shows a uniform orientation, but its attitude varies considerably form place to place, and in some of the larger quarry faces it is strikingly folded (Fig. 2).

The layering may be marking the forms of lava domes, and the presence of such domes is reported by Micheluccini (1964) and by Barberi et al. (1967). However, on the basis of our admittedly brief examination, we would interpret the layering as levels of vesicles that formed on surfaces of high simple shear strain as the viscous lava flowed, and which were deformed as the flowing

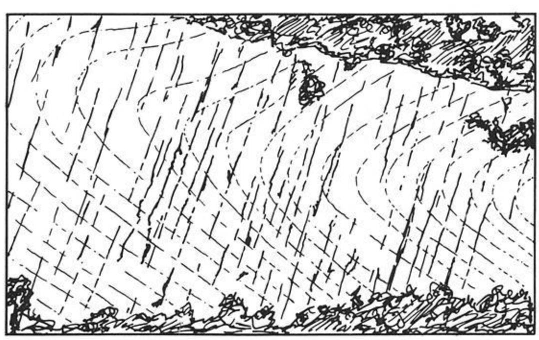

Fig. 2. Exposed face of the rhyolite in the abandoned quarry of site 3 (drawn from a photograph, looking NE). The curved flow lines are formed by layers of vesicles; the steep joint surfaces are tilted at a small angle from the vertical. The face is about 10–15 m high

continued. In this interpretation the large recumbent folds (Fig. 2) would have formed where the more viscous, frontal part of the flow was over-ridden and turned under as the rest of the flow continued to move.

Cooling joints are well developed in some exposures, for example locality CM-3, where they cut across the layering on both limbs of a recumbent fold of the layering (Fig. 2). As discussed below, the cooling joints are important, in that they provided us with a way of making a tectonic correction to our paleomagnetic data.

Rock Magnetism

Optical examination of polished sections was not particularly helpful in identifying the carrier of magnetization in the rock. The only opaque minerals observed were fairly coarse ilmenites and occasionally hematite. The nature and concentration of the visible grains can not account for the strong, stable magnetic character of the samples, which must be attributed therefore to very fine, sub-microscopic grain sizes. The magnetic mineralogy was investigated further by studying the acquisition of isothermal remanent magnetization (IRM), low-temperature behavior, and the alternating field (AF) demagnetization characteristics of IRM and the natural remanent magnetization (NRM).

IRM Acquisition

Selected samples from each site were progressively given IRM in ever-increasing magnetizing fields until saturation IRM was reached. The behavior of all the samples was remarkably similar, and their IRM acquisition curves fell between the two extreme cases illustrated in Fig. 3a. Between 85 and 95% of saturation IRM was reached in magnetizing fields of less than 0.2 T (2 kOe), and saturation was reached between 0.35 and 0.45 T (3.5 to 4.5 kOe). There were no indications of the presence of higher coercivity magnetic minerals. It therefore appears unlikely that the few coarse hematite grains observed optically can contribute appreciably to the natural remanence. The magnetic mineralogy is characterized by moderate coercivity carriers of remanence.

Low Temperature Studies

Nagata et al. (1964) have shown that it is possible to identify the magnetic minerals in a rock by observing the occurrence of low-temperature magnetic transitions during thermal cycling between liquid nitrogen temperature ($-196°$ C) and room temperature.

The sample was given an IRM in 1 Tesla at room temperature. Using a device described by Heiniger and Heller (1976) which allows continual monitoring of all three components of a remanence during thermal cycling, the magnetization was observed during cooling to $-196°$ C, and partial rewarming to

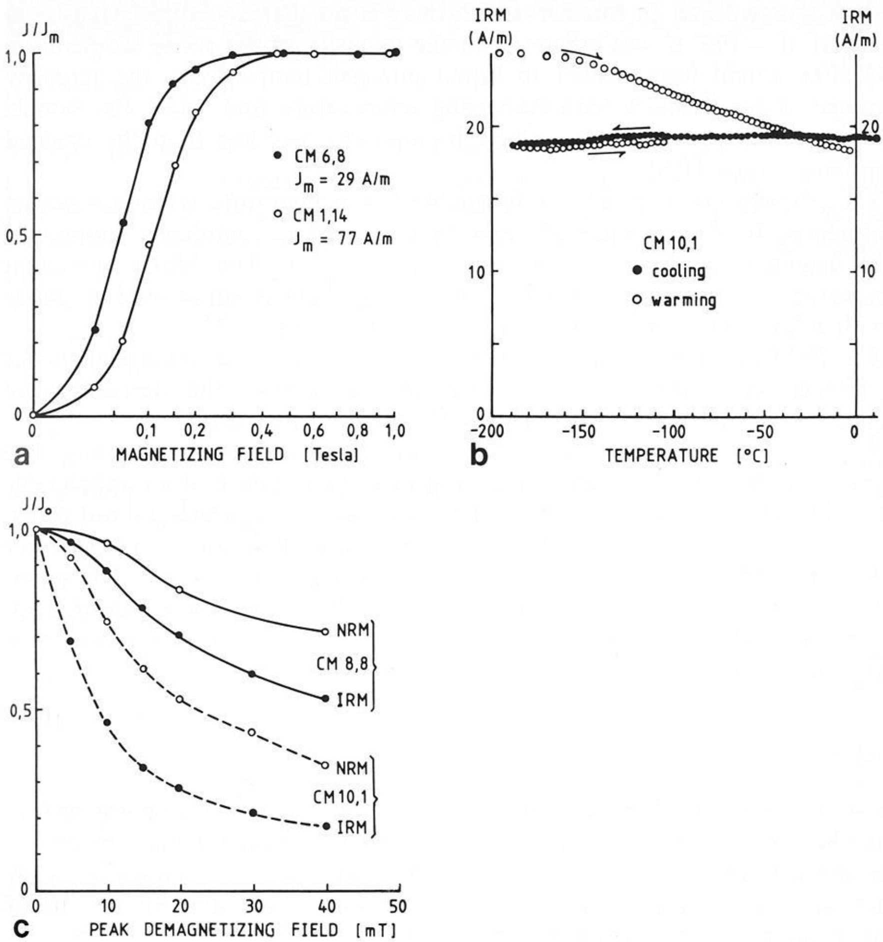

Fig. 3. **a** Acquisition of isothermal remanent magnetization (J), normalized in terms of the saturation *IRM* (J_m), for two representative samples. **b** Low temperature thermal cycling of *IRM*. *Solid points* represent intensities measured during cooling of room temperature *IRM*. *Open points* are values observed during rewarming of this *IRM*, and also during warming of an *IRM* given at low temperature (−196° C). **c** Normalized AF demagnetization curves for natural remanent magnetization (*NRM*) and isothermal remanent magnetization (*IRM*), showing the dominance of very fine grained carriers of remanence

about −105° C (Fig. 3b). Any component of the room temperature IRM carried by pure hematite would be lost at the Morin transition at about −20° C, and in coarse-grained magnetite a further change in magnetization should be observed at the magnetocrystalline anisotropy transition point (about −150° C for pure magnetite). Neither of these sharp transitions was observed; there was negligible change of remanence during cooling, or during rewarming to −105° C.

The sample was given a new IRM in an applied field of 1 T at liquid nitrogen temperature and the magnetization was monitored during re-warming to room temperature. In this curve also there is no sharply defined transition. The IRM at $-196°$ C was greater than the intensity of the room temperature IRM after it had been cooled to liquid nitrogen temperature, the intensity decreased almost linearly with increasing temperature and when the sample had warmed up to room temperature its intensity was less than the original room temperature IRM.

The temperature at which the magnetite low-temperature transition occurs is influenced by the presence of impurity ions, and is completely suppressed if the magnetite grains are in the single domain size. The Morin transition in hematite is not greatly sensitive to particle size, but is suppressed in grains finer than 0.01 μm, or when the titanium content exceeds 0.3%.

The IRM acquisition curves and the absence of a Morin transition in our low temperature experiments are interpreted as evidence that ferromagnetic hematite is not an important magnetic constituent of the rock.

The room temperature IRM is unaffected by low temperature cycling. The absence of a magnetite transition argues against the presence of an appreciable multidomain fraction, in sympathy with the optical studies which did not reveal any visible magnetite grains. The low temperature IRM was much stronger than room temperature IRM; the difference is probably due to ultra-fine grains which are single domain at low temperature, and which become superparamagnetic on warming up to room temperature, causing a loss of low-temperature IRM in the process.

Domain State Test

Lowrie and Fuller (1971) pointed out that the AF demagnetization characteristics of thermo-remanent magnetization (TRM) and IRM in multidomain magnetite were distinctively different from those in single domain magnetite. They suggested a simple test by means of which the domain state of the NRM carrier in an igneous rock (in which the NRM was a TRM) could be determined. This is, that single domain TRM is more resistant than IRM to AF demagnetization, and multi-domain TRM is less resistant than IRM. This test becomes less clear-cut when a mixture of single and multidomain grains are present, and it is not clear how effective it is for pseudo-single domain grains.

AF demagnetization curves of representative specimens of the San Vincenzo rhyolite show that the NRM is clearly more resistant than IRM to this form of demagnetization (Fig. 3c). The NRM is therefore dominated by single domain carriers of remanence.

Summary of Rock Magnetism

The rock magnetic studies show that the magnetic mineralogy is dominated by a single phase, consisting of magnetite in very fine, sub-microscopic grain sizes, probably in the very stable single domain range. It is also possible that some of these grains extend into the pseudo-single domain range, which might account for the slight losses of remanence during the low temperature cycling.

Paleomagnetism

AF Demagnetization

The average intensity of the NRM of the San Vincenzo rhyolite was 0.72 A/m (7.2×10^{-4} G). The remanences were uniformly stable, with quite high median destructive fields averaging almost 40 *mT* (400 Oe.). One specimen from each of the 100 samples was progressively demagnetized in alternating magnetic fields up to at least 80 *mT*. Vector diagrams (Fig. 4) were used to determine the inclination and declination of the characteristic remanent magnetization (ChRM).

All samples were reversely magnetized. A slight instability was remarked during the initial cleaning steps, accompanied in a few samples by a slight increase in intensity as a low coercivity normal component was removed. Beyond about 15 *mT* all samples possessed very good directional stability, with a single stable component of remanence defined by straight lines to the origin for both inclination and declination.

Thermal Demagnetization

A number of specimens were thermally demagnetized using either the conventional, progressive (stepwise) technique, or a continuous observation technique.

Stepwise thermal demagnetization (Fig. 5a) showed negligible change of NRM intensity or direction below 200° C. A single component of magnetization, leading linearly to the origin in both inclination and declination parts of the vector diagram, characterized the temperature range from 400° to 580° C. A small component remaining at 580° C was completely eliminated by heating to 600° C, and probably represents a small error in temperature calibration. The blocking temperature spectrum is clearly dominated by the high temperature component between 400° and 600° C (Fig. 5a).

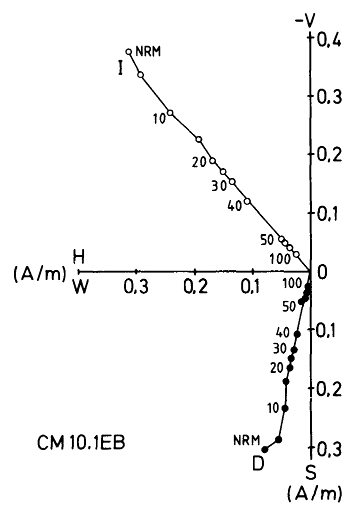

Fig. 4. Vector diagram illustrating the behavior of *NRM* during AF demagnetization. *Open points* are defined by the vertically upwards *(V)* and horizontal *(H)* components, and illustrate the variations of inclination. *Solid points* define south *(S)* and west *(W)* components and illustrate the variations of declination

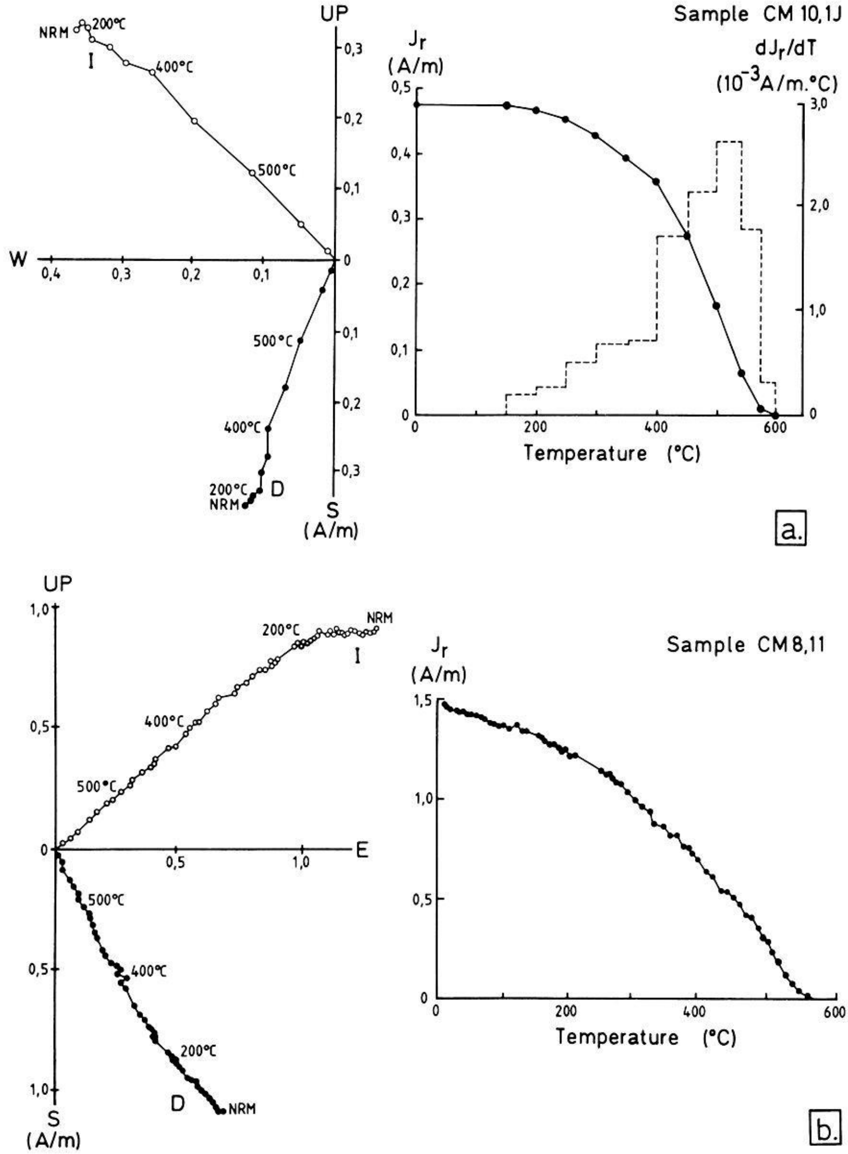

Fig. 5. **a** Step-wise thermal demagnetization of *NRM* (J_r). The dashed histogram is obtained by differentiating the intensity decay curve, and represents the blocking temperature spectrum. **b** Continuous thermal demagnetization of *NRM*. As in the stepwise example, the remanence above 400° C consists of a single stable component. Vector diagrams: as in Fig. 4

The continuous thermal demagnetization method (Heiniger and Heller, 1976) enables the determination of all three components of remanence at any temperature during heating up to the Curie temperature. An almost continuous record of the intensity and direction of the remanence is obtained. The method has the disadvantage that both decay of spontaneous magnetization and thermal unblocking of remanence are observed as a joint effect, but an advantage is that mineralogical changes can be recognized. The technique can be a powerful tool for unravelling the magnetic composition of complex rocks in which metastable magnetic minerals are present (see Heller, 1978).

The continuous demagnetization curves (Fig. 5b) are very similar to the progressive demagnetization curves, and indicate that no marked mineralogical change occurs during heating. The slight change of direction below 200° C is exaggerated by the accompanying reduction of spontaneous magnetization, there is again a slightly unsettled component between 200° and 400° C, and from 400° C until the magnetization disappears at about 560° C there is a single stable component of magnetization.

Characteristic Remanent Magnetization (ChRM)

The remanent magnetizations have a single, stable component after demagnetizing in fields above 20 mT or in temperatures above 400° C. The directions of this ChRM component were averaged using Fisher statistics for all sites. The site mean directions are given in Table 1 and Fig. 6a.

The data are extremely tightly clustered at each site, and the 10 site mean directions are also closely grouped. The overall mean direction for the formation has declination 160°, inclination −45°, and the circle of 95% confidence of this mean has radius 6.6°.

The negative polarity found in these rocks agrees well with that expected for their radiometric age of 4.7 m.y. On the geomagnetic reversal time scale (LaBrecque et al., 1977) this is appropriate to the reversed polarity zone near to the base of the Gilbert reversed chron. The direction expected for this age should not be very different from the reversed direction of the present axial dipole field, with declination 180° and inclination −62°. The observed formation mean direction is 21° away from the expected direction.

Determination of the Tectonic Correction

The absence of stratified rocks in association with the rhyolite prevented us from making a standard bedding correction, but the presence of cooling joints in the rhyolite suggested that we might correct for post-eruption tilting by restoring the joint-bounded columns to a vertical orientation. We recognize that not all cooling joints are vertical when formed, but the consistency of orientation of the columns throughout the rhyolite mass makes an original vertical orientation quite likely.

The available number, degree of exposure and quality of the joint surfaces were variable. At each of the four quarry sites we measured the orientations of 20 cooling joint surfaces (Fig. 7a). A great circle (the π-plane) was fitted to the set of poles at each site using the method given by Ramsay (1967). The pole of the π-plane (D_p, I_p) and the angular standard deviation of the measured poles from the best-fitting π-plane (δ) were computed. The degree

.**Table 1.**

Site number	Number of samples	k	Declination	Inclination	α_{95}
1	14	369	154	−48	2.1
2	9	948	165	−42	1.7
3	8	1383	171	−55	1.5
4	7	424	159	−40	2.9
5	5[a]	680	137	−43	2.9
6	8	1022	156	−35	1.7
7	10	102	173	−49	4.8
8	11	240	156	−41	3.0
9	10	609	148	−42	2.0
10	15	678	184	−45	1.5
Site mean	10	54	160	−45	6.6

[a] Three samples rejected from unsafe (displaced) outcrop in stream bed

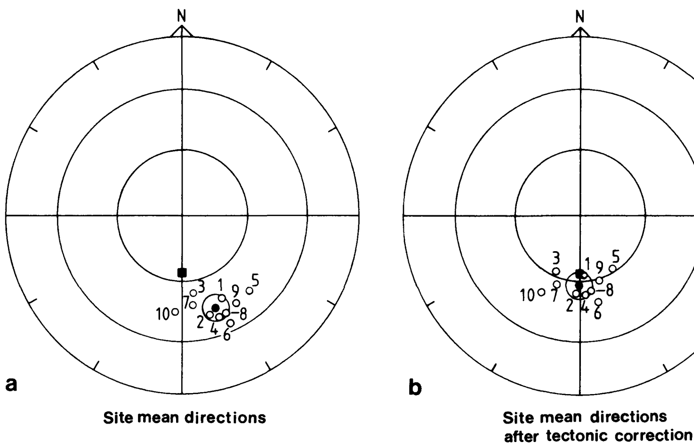

Fig. 6. a Stereographic projection of the site mean directions of the characteristic remanence (ChRM). *Solid square:* direction of reversed axial dipole field. *Small circle:* circle of 95% confidence of the overall mean direction *(solid circle).* **b** The directions of **a** after making a tectonic correction by rotating the pole of the π-plane (Fig. 7b) to the vertical

of fit was good at all sites (δ less than 10°), and the poles of the π-planes were very similar. Differences between the poles could be due to local differences in tilt from site to site. However, we were able to determine a π-plane pole at only four of the ten sites, and this did not allow us to make individual tectonic corrections at each site.

Instead, we note that the differences between the π-plane poles are small, and combine the 80 measured poles to columnar joint surfaces at all four sites to obtain a single π-plane (Fig. 7b). The pole of this plane has azimuth 300° and inclination of 70°. The tectonic correction for all paleomagnetic sites consists of restoring this pole, which corresponds to the mean intersection line of the

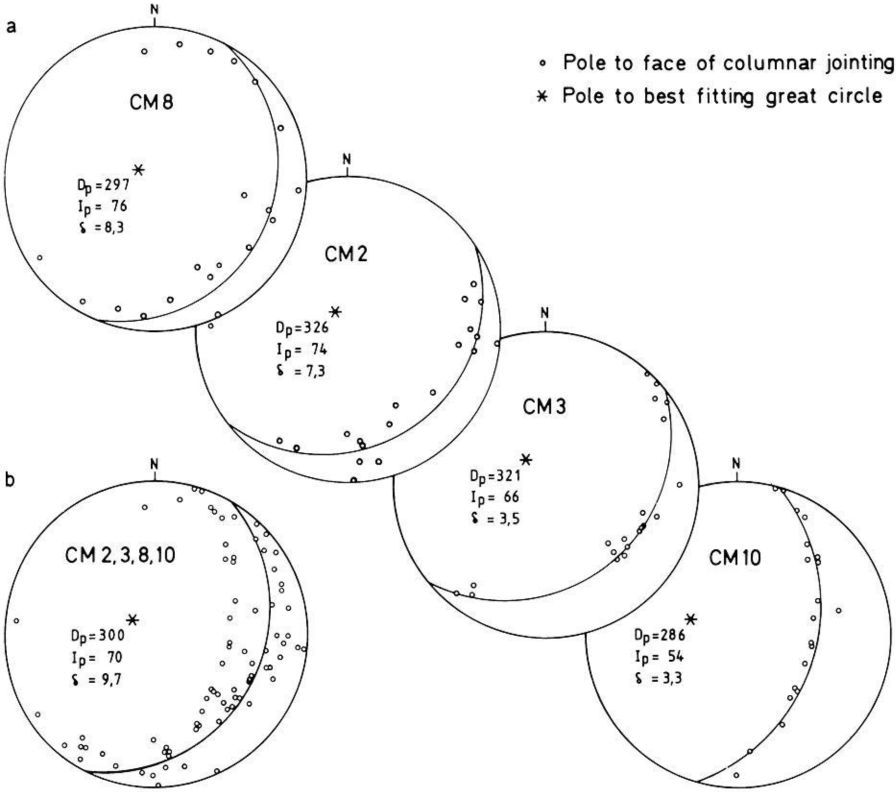

Fig. 7. a At sites 2, 3, 8, and 10 in active or abandoned quarries, the poles to 20 measured joint surfaces form a girdle. The best fitting great circle (π-plane) has its pole at (D_p, I_p); the goodness of fit is expressed by the average angular deviation from the π-plane (δ). **b** The combined data (80 poles) from all 4 sites define a π-plane whose pole has azimuth 300° and inclination 70° to the horizontal

joints, to the vertical. When this is done, the mean paleomagnetic direction has declination 179°, inclination −58°, and the reversed axial dipole direction falls within the circle of confidence of the corrected mean (Fig. 6b).

VGP Positions and Dispersion

The normal paleomagnetic pole position calculated from the uncorrected formation mean direction lies at 67°N 242°E and the principal axes of the oval of 95% confidence are $\delta p = 5.2°$, $\delta m = 8.3°$. The corresponding pole position for the tectonically corrected formation mean is 84°N 240°E ($\delta p = 7.2°$, $\delta m = 9.7°$).

If the virtual geomagnetic pole positions (VGP) computed from the uncorrected mean direction of each site are averaged they give a pole position at 67°N 241°E, and the circle of 95% confidence has radius $\alpha_{95} = 7.7°$. The corresponding pole position for tectonically corrected data is 84°N 246°E ($\alpha_{95}=7.7°$). The angular dispersion of this distribution (given by $\alpha=81/\sqrt{k}$, where k is the best estimate of Fisher's precision parameter) was found to be 12.8°.

Models of the secular variation of the geomagnetic field (Cox, 1970) recognize two contributions to the angular dispersion of VGP. The first of these is wobble of the geomagnetic dipole about its mean (axial) position; this effect has an amplitude of about 11° and is independent of the latitude of the observation. The second contribution derives from secular variation of the non-dipole field and is latitude-dependent (Creer, 1962). For the latitude of our sampling sites (43.1°N) the non-dipole contribution to the angular dispersion of VGP ought to amount to about 10.8°. The observed dispersion in our study is slightly too high to be accounted for by the nondipole field alone. If dipole wobble is incorporated with non-dipole effects secular variation models *C* and *D* (Cox, 1970) predict a dispersion of 16.0° to 16.4°, considerably larger than that observed.

Discussion

The San Vincenzo rhyolite has a single, stable component of natural remanent magnetization. Magnetic mineralogy studies indicate that the NRM is probably carried by single domain or pseudo single domain magnetite. The remanence is probably original; there are no rock magnetic indications that it has been altered since formation of the lava.

Although the polarity of NRM corresponds to expectation for the appropriate position in the Gilbert reversed chron at the base of the Pliocene, the uncorrected mean direction has an inclination that is lower than expected, and a declination that is rotated by 20° in a counterclockwise sense.

The observed VGP dispersion is indicative of incomplete averaging of secular variation effects. This thick lava would not cool rapidly except at its surface; it evidently took long enough passing through the broad range of blocking temperatures (Fig. 5a) to record a variety of total field directions. The observed VGP dispersion is large enough to account for averaging of the non-dipole field but not of the total field. It is probable that at least part of the 21° discrepancy between the observed formation mean direction and the reversed axial dipole field direction can be accounted for in this way. However, part must also derive from other causes, probably tectonic in origin.

Examination of the local tectonic situation indicated that some compensation for possible rotation about a horizontal axis would be necessary. The only geological indicator that could be used was an internal structural feature of the lava, the attitude of cooling joint surfaces. By measuring the poles to these joint planes, and determining the pole to their π-planes, an overall tectonic correction for the region was obtained, assuming that the joints were vertical when formed.

On application of the tectonic correction the mean direction was found to be not significantly different from that of the expected reversed field. Therefore, any rotation of the northern part of the Italian peninsula, as proposed by VandenBerg (1979) for the Late Tertiary, was completed by the time of formation of the San Vincenzo rhyolite, 4.7 m.y. ago, in the earliest Pliocene.

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