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Palaeomagnetic and rock magnetic reliability criteria in ophiolitic rocks: a case study from the Palaeozoic Ballantrae Ophiolite, Scotland

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ABSTRACT

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Palaeomagnetic studies of the Ballantrae Ophiolite, southwest Scotland, are assessed in the light of "palaeomagnetic" and "rock magnetic" reliability criteria. The remanence directions are palaeomagnetically *reliable* but have only limited *significance* for apparent polar wander path construction. "Rock magnetic" reliability criteria are found to be ambiguous when treated in isolation.

Original grain size effects are detected in magnetic properties across three sections through pillow lavas. We propose that a systematic study of spatial magnetic property variations within pillow lavas can provide an additional field test in the assessment of reliability.

Thermomagnetic analyses and microscopic observations suggest three remanence-carrying magnetic phases exist within the extrusive layer of the ophiolite. These are almost pure magnetite, hematite and titanomaghemite. Palaeomagnetic, rock magnetic and geological observations are best satisfied by a pre-obduction magnetisation age. This study links the remagnetisation of an intra-oceanic conglomerate to local hydrothermal circulation rather than to a pervasive remagnetisation event as previously suggested.

Introduction

The requirement for palaeomagnetic data of varying quality to be assessed during the construction of continental apparent polar wander paths (APWPs) has led to the establishment of palaeomagnetic reliability criteria (e.g. McElhinny, 1973; Briden and Duff, 1981; Van der Voo, 1988). These *palaeomagnetic* criteria, which are equally applicable to the palaeomagnetic study of ophiolites, can be informally summarised under the following headings:

(1) Thoroughness of field sampling strategy.

(2) Thoroughness of demagnetisation experiments.

(3) Lack of interpretational ambiguities e.g. the palaeopole does not resemble those from younger rocks, or polarity reversals are stratigraphically linked.

The magnetic study of ophiolite complexes was initiated after their recognition as the possible obducted remnants of oceanic lithosphere (Gass, 1968). Palaeomagnetic and rock magnetic research has subsequently been concentrated in the following fields: (a) determination of ophiolite magnetic properties for use in the geophysical modelling of marine magnetic anomalies; (b) palaeomagnetic recognition of intra-oceanic block rotations utilised in the study of ocean floor tectonics; and (c) palaeomagnetic studies aimed at the reconstruc-

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tion of ophiolite movement histories both prior to, and subsequent to their obduction.

Levi et al. (1978) compared the magnetic properties from the extrusive layers of five ophiolites with those of directly sampled ocean crust (Deep Sea Drilling Project samples). If a favourable comparison of the two was observed, this was considered to reflect a "reliable" ophiolite for use in marine magnetic anomaly modelling. Their minimum constraints for the magnetic properties of ophiolite extrusives are given below. These constraints can be considered as *rock magnetic* criteria in the assessment of ophiolite reliability:

(1) Initial Curie points (T_c) should be between 100° and 450°C (for $T_c > 200$°C, the saturation magnetisation versus temperature curves should be irreversible).

(2) J_{nrm} (intensity of initial magnetisation) > 100 mA/m.

(3) Q, Koenigsberger ratio (proportion of remanent to induced magnetisation) > 1.

(4) Alteration of the upper extrusives should not exceed zeolite facies metamorphism.

A plenary treatment of the palaeomagnetic reliability of ophiolites should therefore combine *palaeomagnetic* and *rock magnetic* reliability schemes. For example, a characteristic remanence which appears on "palaeomagnetic" grounds to be of primary origin, may in fact prove to be secondary when "rock magnetic" constraints are applied. Similarly, a remanence carried by an original mineralogy may yet prove secondary when appropriate palaeomagnetic field tests are performed.

In this study, we assess the palaeomagnetism of the Ordovician Ballantrae Ophiolite Complex of southwest Scotland in the context of both *palaeomagnetic* and rock *magnetic* reliability criteria. The tectonic implications of this palaeomagnetic work have been discussed elsewhere (Trench et al., 1988) and are only briefly reviewed herein.

In this paper, we introduce new rock magnetic data, together with scanning electron microscope (SEM) and optical microscope observations, which bear on the palaeomagnetic record of the ophiolite. We also recount a variation in remanence characteristics observed within three pillow lavas sampled by Trench et al. (1988).

Geological background

The Ballantrae Ophiolite Complex (BOC; Stone and Rushton, 1983) forms an element of a widespread ophiolitic suite obducted at the southern Laurentian margin in Early Ordovician times (e.g. Dunning and Krogh, 1985). The BOC was first interpreted as an ophiolite by Church and Gayer (1973) and Dewey (1974). It comprises both ocean crust- and mantle- derived lithologies which became juxtaposed during obduction (Stone, 1984). These now crop out as two NE–SW trending tracts of ultramafics and spilites referred to as the northern and southern belts of the ophiolite. The BOC formed, and obducted, within Early Ordovician times, which is constrained using radiometric and palaeontological evidence (summarised by Stone and Smellie, 1988 p. 98). This paper focuses on results from the Slockenray Formation (Bluck, 1982), in the northern belt of the BOC (Fig. 1).

Sampling strategy and previous palaeomagnetic research

The Slockenray area of the BOC was considered to be of particular palaeomagnetic interest given several favourable factors described below:

– The low metamorphic grade, which is predominantly zeolite facies (Oliver et al., 1984; Smellie, 1984a).

– The low degree of deformation and alteration of the succession when compared to much of the ophiolite elsewhere. Indeed, Bailey and McCallien (1957) considered the possibility of a Tertiary age for the Slockenray sequence due to its minimal alteration.

– The excellent age control afforded by the presence of Early Arenigian graptolites (Rushton et al., 1986).

– The presence of an intra-formational conglomeratic facies (Bluck, 1982) allowing a palaeomagnetic conglomerate test (Trench et al., 1988).

– The availability of detailed, large-scale, plane table outcrop maps for the area (Bluck, 1982; Trench et al., 1988). This enabled a comparison of remanence directions between fault blocks within the sequence.

The determination of an accurate palaeohorizontal is of great importance in all palaeomagnetic studies. In this regard, lava sequences (oceanic layer 2) may prove problematical, most acutely where little inter-pillow sediment is present.

For the Slockenary Formation, two factors control the quality of palaeohorizontal determination. Firstly, the formation comprises approximately 70% inter-lava sediment, which allows the attitude of lavas to be determined with confidence from inter-bedded horizons. Secondly, large-scale foresets were identified in the sedimentary column which truncate the true bedding surfaces (Bluck, 1982). The latter were used for tectonic correction.

It was previously noted that the thoroughness of sampling proves an important factor in the assessment of palaeomagnetic reliability. All three previous palaeomagnetic studies of the BOC have been aimed at the reconstruction of its former position. The results of Nesbitt (1967), Piper (1978) and Trench et al. (1988) have therefore most easily been assessed using *palaeomagnetic* reliability criteria. Assignments of reliability from Briden and Duff (1981) and Stearns et al. (1989) for the Ballantrae studies are outlined in Table 1. The respective sampling areas of each study are shown in Fig. 1.

A combination of the palaeomagnetic results of Piper (1978) and Trench et al. (1988) furnishes constraints upon both the magnetic and tectonic history of the ophiolite. These can be summarised as follows;

(1) The Early Arenig Slockenray Formation carries a characteristic remanence (termed component M; Trench et al., 1988) which predates the folding of the succession into a steeply plunging synclinal structure: folding is thought to relate to the Early Ordovician obduction of the ophiolite (Stone, 1984).

(2) Component M represents a chemical remanent magnetisation (CRM) or thermo-chemical remanent magnetisation (TCRM) overprint as it passes through an intra-formational basaltic conglomerate (Slockenray Conglomerate, Fig. 1). A purely thermal origin is unlikely as unblocking temperatures of 560°C occur within zeolite facies rocks. Trench et al. (1988) argue for a pervasive

TABLE 1

Published reliability assessments of palaeomagnetic studies from the Ballantrae ophiolite

Van der Voo reliability criteria (from Stearns et al., 1989)

	1	2	3	4	5	6	7	*Q*
Ballantrae intrusives combined (Piper, 1978)	(×) [a]	×	×				×	3 (4)
Slockenray Formation (Trench et al., 1988)	×	×	×	×			×	5

[a] This criterion is now satisfied (see text).

Briden and Duff reliability criteria (from Briden and Duff, 1981).

	Pole classification
Ballantrae gabbros (Piper, 1978)	B
Ballantrae serpentinites (Piper, 1978)	B

The Van der Voo (1988) *Q* factor comprises seven individual reliability criteria as follows: (1) rocks are well dated, to within half a period; (2) at least 25 individually oriented sample directions used, $\kappa > 10$ and $\alpha_{95} < 16°$; (3) demagnetisation is sufficiently performed and data published; (4) field tests constrain magnetisation age; (5) results are based on structural corrections and are not suspected as rotated; (6) reversals are present; and (7) palaeopole does not resemble any from younger rocks.

Briden and Duff's (1981) A–D pole classification is based on a flow diagram method of assigning reliability (see original reference). Criteria 1, 2, 3, 4 and 5 adopted by Van der Voo are effectively specified by Briden and Duff (with minor differences in acceptance levels). Greatest significance is attributed to the constraint of magnetisation age (4), for which a pole is given an "A" grading.

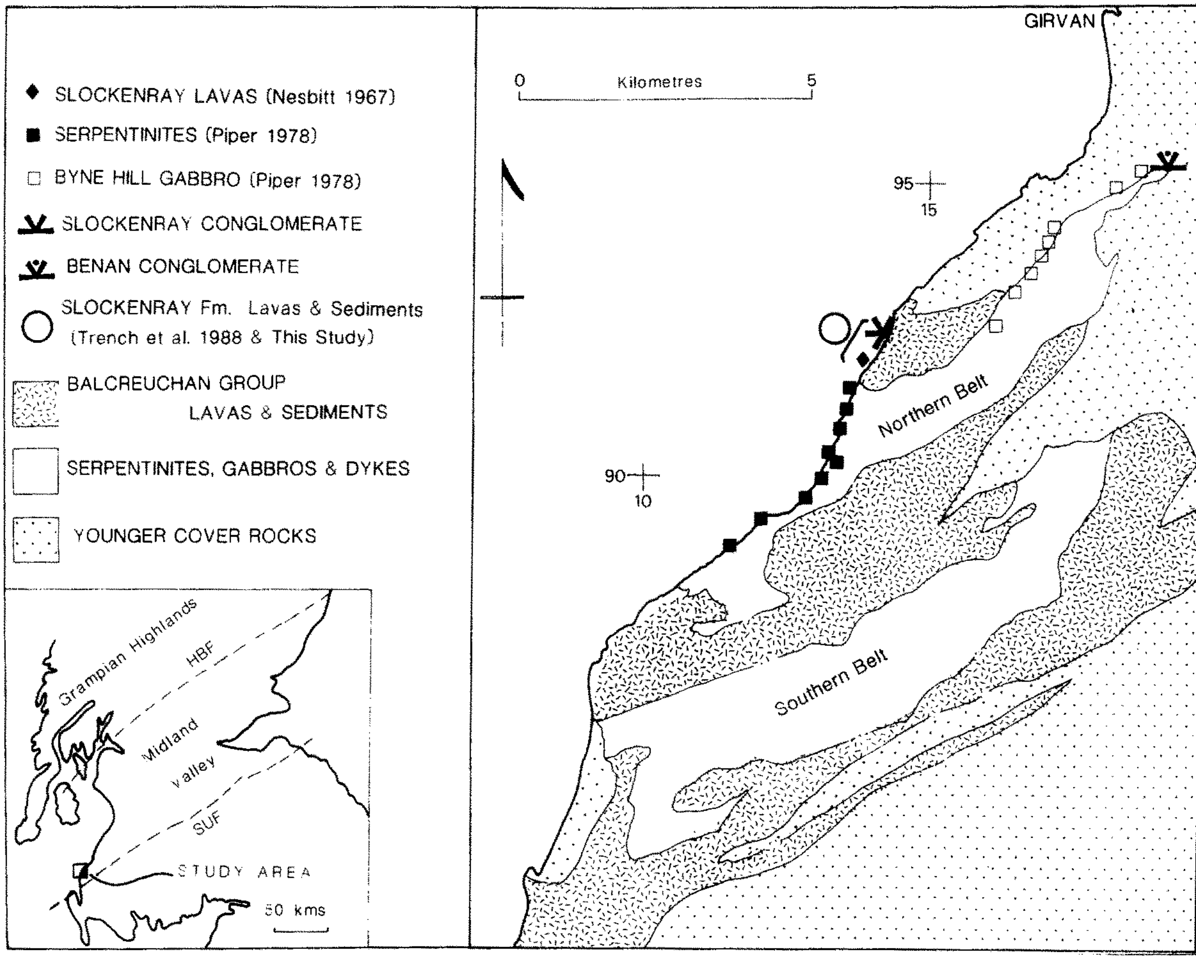


Fig. 1. Geographic location (inset) and geological sketch map of the Ballantrae Ophiolite complex showing positions of former palaeomagnetic studies. *HBF* = Highland Boundary Fault; *SUF* = Southern Upland fault.

overprint. New evidence presented here is more compatible with localised resetting of this conglomerate however.

(3) A positive conglomerate test (with respect to component M), from the overlying Benan Conglomerate (Fig. 1), defines a post-Early Arenig–pre Llandeilo remagnetisation window. Magnetisation of the Slockenray Formation therefore predates obduction (see (1) above).

(4) The Slockenray Formation (and by implication, the BOC), formed at moderate southerly latitudes (29 ± 4°S), suggesting a tectonic position within the Iapetus Ocean. In Ordovician times, the ocean is estimated to cover approximate latitudes of 25–35°S (Briden et al., 1984) or 25–60°S (Perroud et al., 1984) dependent upon contrasting interpretations of southern British palaeomagnetic data (see Van der Voo, 1988, and Torsvik et al., 1990, for reviews).

(5) The Slockenray Formation has experienced a clockwise rotation of approximately 80° when referred to contemporaneous British poles from the northern Iapetus margin (e.g. Watts and Briden, 1984; Watts, 1985).

New palaeomagnetic and rock magnetic observations

Variation of magnetic properties within pillow basalts

Here we report the variation in palaeomagnetic properties across three sample traverses within individual pillow lavas of the Slockenray Formation (sites PL1, PL2 and PL3, of Trench et al., 1988). A number of magnetic parameters were observed, including: (1) intensity and direction of natural remanent magnetisation (NRM); (2) low field volume susceptibility (K); (3) Koenigsberger ratio (Q); (4) stability of NRM to thermal or AF treatment (measured as a thermal stability index, J_{400}/J_{nrm}, or as the median destructive field, MDF); and (5) direction of thermal or AF cleaned magnetisation (component M).

Field and laboratory procedures are described by Trench et al. (1988). Susceptibility measurements were made in the field using a Microkappa Kappameter model KT-5 field rock susceptibility meter. The meter correlates well with laboratory susceptibility measurements over several orders of magnitude (Evans and Greenwood, 1988).

The variations of these parameters across each pillow lava are shown in Figs. 2, 3 and 4. The magnetic properties of the pillow lavas are compared with other Slockenray Formation lithologies in Table 2.

We draw attention to systematic variations in parameters which are symmetrical across the pillow lava traverses. This is evident for NRM intensity, Koenigsberger ratio, thermal stability index and NRM inclination in Figs. 2 and 3, and volume susceptibility in Fig. 4.

Systematic changes of this type have previously been described within "in situ," oceanic pillow basalts (e.g. Watkins et al., 1970; Marshall and Cox, 1971; Soroka and Beske-Diehl, 1984), where they are principally attributed to variations in quenching-rate. A study of Precambrian pillows failed to show any systematic variations (Piper, 1976). If the trends in the Ballantrae pillows are explicable in similar terms to above, they would represent the first such observations for a Palaeozoic ophiolite. A brief description of the variations is given below:

(1) NRM intensity decreases toward the pillow centre and peaks close to the pillow margin (Figs. 2 and 3), which may reflect a greater NRM retentivity within the "quenched" smaller grain sizes at the margin (Watkins et al., 1970). An intensity peak below the outer crust for pillows of equivalent dimension to these is reported by Marshall and Cox (1971) and is again related to a grain size effect. (Note that the NRM intensities in the latter study are approximately 1000 times greater than those reported here.)

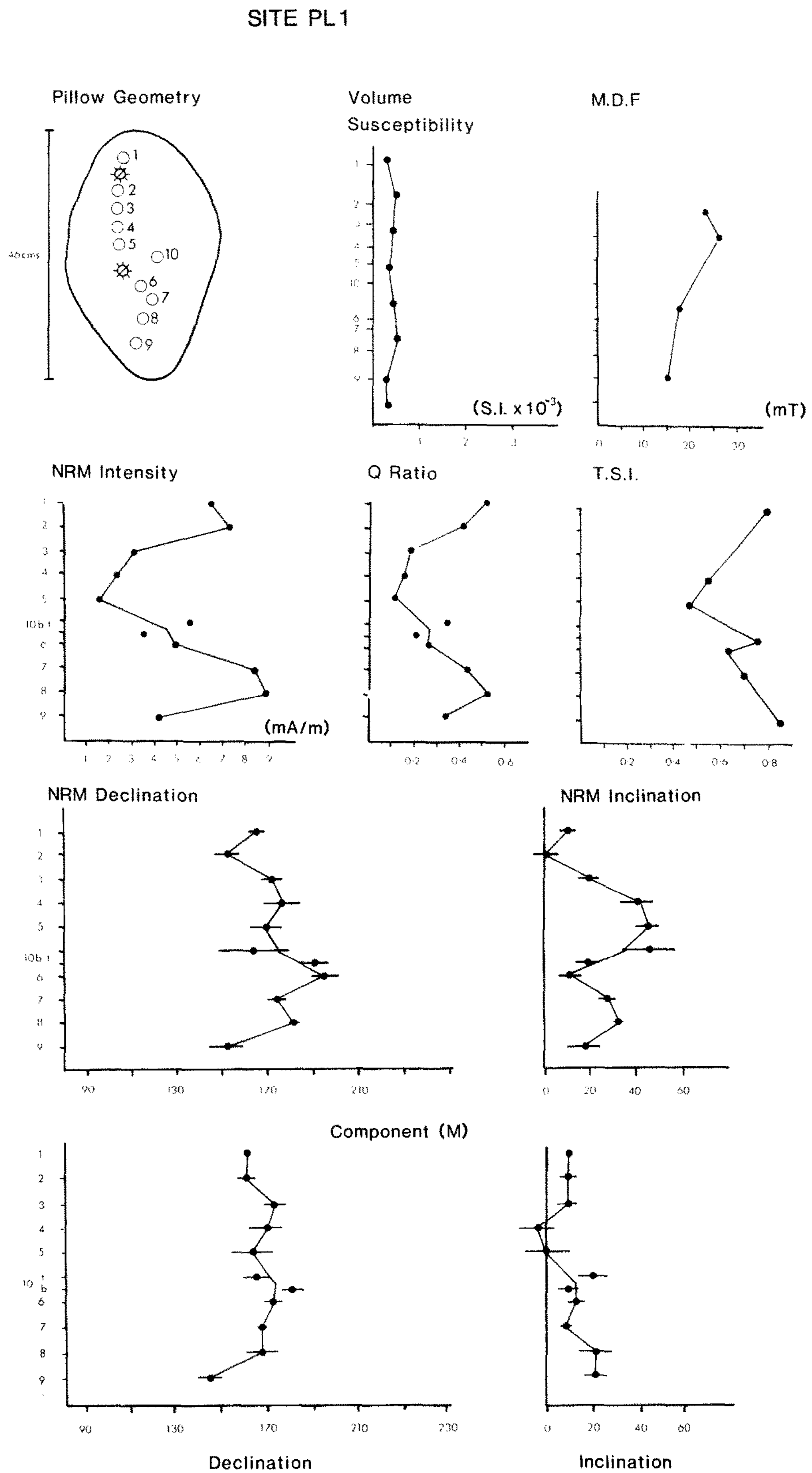
(2) Koenigsberger ratios are generally low throughout; being less than unity for 24 of the 26 samples. However, the ratios decrease significantly toward the pillow centre in a similar fashion to the NRM intensities (Figs. 2 and 3). The trend is interpreted as an increasing grain size and accompanying change in domain state from pillow margin to centre (Cox and Doell, 1962; Stacey, 1967). Site PL3 does not conform to this simple relationship although Koenigsberger ratios remain generally low and only two exceed unity (Fig. 4). Low ratios result from order of magnitude in-

TABLE 2

Summary of Slockenray Formation magnetic properties

Lithology (site)	NRM intensity range (mA/m)	Susceptibility range (10^{-3} SI units)	Mean ratio (Q) [a]
Basalt (TSR1)	306 –2210	0.36– 7.71	5.9
Basalt (TSR2)	135 –3194	7.81–17.6	4.2
Basalt (TSR3)	111 – 410	0.90–27.5	1.1
Black shale (TSR5)	24.2 – 162	–	–
Chert (TSR6)	0.72– 25.2	0.10– 0.29	0.51
Lithic arenite (TSR7)	77.7 – 496	0.84– 9.54	1.9
Lithic arenite (TSR8)	7.4 – 44.5	–	–
Pillow lava (PL1 and PL2)	0.8 – 8.9	0.29– 0.49	0.28
Pillow lava (PL3)	11.2 – 355	1.65– 9.99	0.29
Red lava top (BOL1) and reddened tuff (BOL2, 3 and 4)	130 –1670	0.48– 2.71	6.7

[a] Koenigsberger ratios are calculated as geometric means of intensity and susceptibility distributions.


SITE PL1
Pillow Geometry
46cms
1
2
3
4
5
10
6
7
8
9
Volume
Susceptibility
(S.I. x 10^{-3})
1 2 3
M.D.F
(mT)
0 10 20 30
NRM Intensity
(mA/m)
1 2 3 4 5 6 7 8 9
10b t
Q Ratio
0·2 0·4 0·6
T.S.I.
0·2 0·4 0·6 0·8
NRM Declination
90 130 170 210
NRM Inclination
0 20 40 60
Component (M)
90 130 170 210 230
0 20 40 60
Declination
Inclination

creases in both intensity and susceptibility for sites PL1 and PL2.

(3) The NRM is more stable to thermal demagnetisation at the pillow margins than for the pillow interior (Figs. 2 and 3). There is no comparable trend in Fig. 4. Stability variation of this type is consistent with a change in grain size/domain state across the pillows. The trend suggests a concentration of the multidomain, less thermally stable particles, towards the pillow centre. Soroka and Beske-Diehl (1984) have described a similar effect for alternating field demagnetisation across a single pillow.

(4) The "in-situ" NRMs for sites PL1 and PL2 are steeper at the pillow centres than margins (Figs. 2 and 3), and smear towards a present day field direction at the site. This behaviour may reflect an increased acquisition of viscous remanence (VRM) at the pillow centres (see Soroka and Beske-Diehl, 1984) and again suggests a prevalence of multidomain grains in the central portion of the pillows. The VRM trend is removed upon demagnetisation (component M).

(5) The volume susceptibility at site PL3 increases significantly from the margin to the pillow centre (Fig. 4), and suggests an increased magnetic grain size at the centre. No detectable trend is recorded for sites PL1 and PL2 (Figs. 2 and 3).

The variations described above no doubt oversimplify the parameters which control the magnetisation state of a pillow lava (see Marshall and Cox, 1971; Ryall and Ade-Hall, 1975; Smith and Banerjee, 1985a; Beske-Diehl, 1990). Importantly, significant inter-pillow (cf. PL1/PL2 and PL3) as well as intra-pillow variations exist. For our present purposes however, we contend that the trends are accountable either in terms of quenching-related grain size changes or as the result of differential low-temperature alteration across the pillows. Both processes favour magnetisation of the pillows prior to obduction. This need not imply the preservation of a "primary" remanence, however, but requires that an original grain size distribution of magnetic mineralogy is preserved.

Scanning electron microscopy (SEM)

SEM observations and energy dispersive analyses (EDA) were carried out on representative polished thin sections. The opaque phases of a massive lava (type 1 of Bluck, 1982), pillowed lava (adjacent to site PL2) and a reddened lava top were studied. Observations made for each of these samples are described below:

Massive lava

Iron–titanium-bearing grains exceeding 50 μm were commonly observed. Exsolution lamellae resulting from high-temperature oxidation cover less than 50% of the grains (Fig. 5a, b) and are indicative of oxidation class II (Ade-Hall et al., 1968). EDA spectra confirm Fe–Ti partitioning to be present between lamellae (Fig. 5c, d). Lamellae compositions approach ilmenite (c) and magnetite (d) end members but spectra peaks are only semi-quantitative. Intra-grain cracking is observed (Fig. 5a, b) and is related to low-temperature alteration producing shrinkage. Crack formation is at stage 1–2 of Johnson and Hall (1978).

Reddened lava top

High-temperature oxidation exsolution lamellae occur within partially corroded grains of up to 50 μm. The lamellae generally cover a higher proportion of the exposed grain than for the massive lava and approach oxidation class III (Ade-Hall et al., 1968). Fe–Ti partitioning is again confirmed upon collection of EDA spectra. Compositions are

Fig. 2. Magnetic property variations within pillow lava site PL1. Individual graphs are as follows (from left to right, top to bottom)—*Pillow geometry*: field sketch of sample distribution across pillow lava. Sample numbers are indicated; open circles show sample positions, "crossed" circles show positions of broken cores which were not recovered. *Volume susceptibility* (in 10^{-3} SI units; note scale change for PL3). *M.D.F.*: median destructive field with AF demagnetisation. *NRM intensity* (in mA/m; note scale change for PL3). In this figure, variation is plotted through the mean of two specimens from core 10. *Q ratio*: Koenigsberger ratio. *T.S.I.*: thermal stability index, J_{400}/J_{nrm}. *NRM declination and inclination*: "in-situ" NRM declination and inclination with 95% confidence. *Component (M)*: cleaned component (M) declination and inclination with error estimates from LINEFIND analysis (Kent et al., 1983).

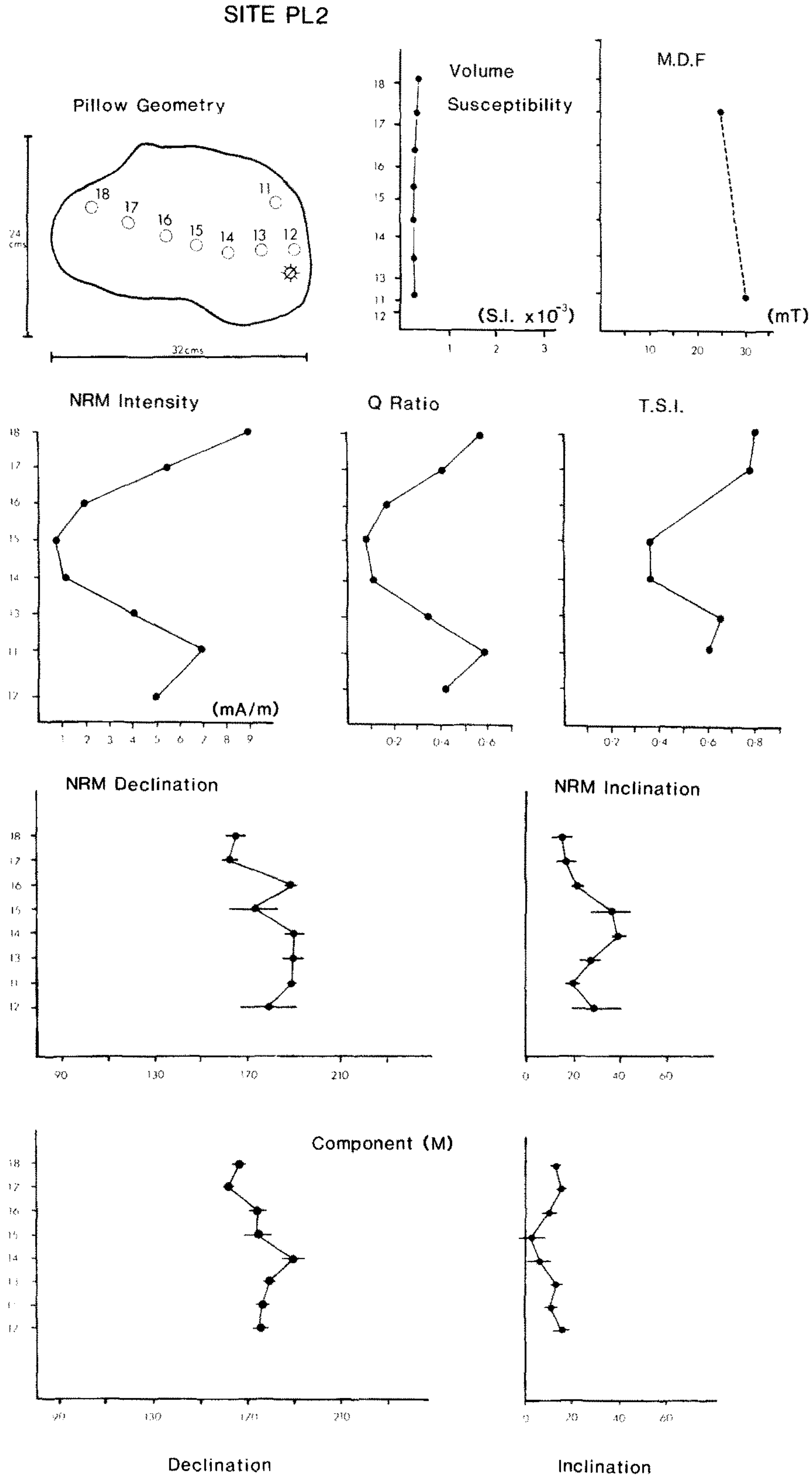


Fig. 3. Magnetic property variations within pillow lava site PL2. See Fig. 2 for explanation of individual graphs.

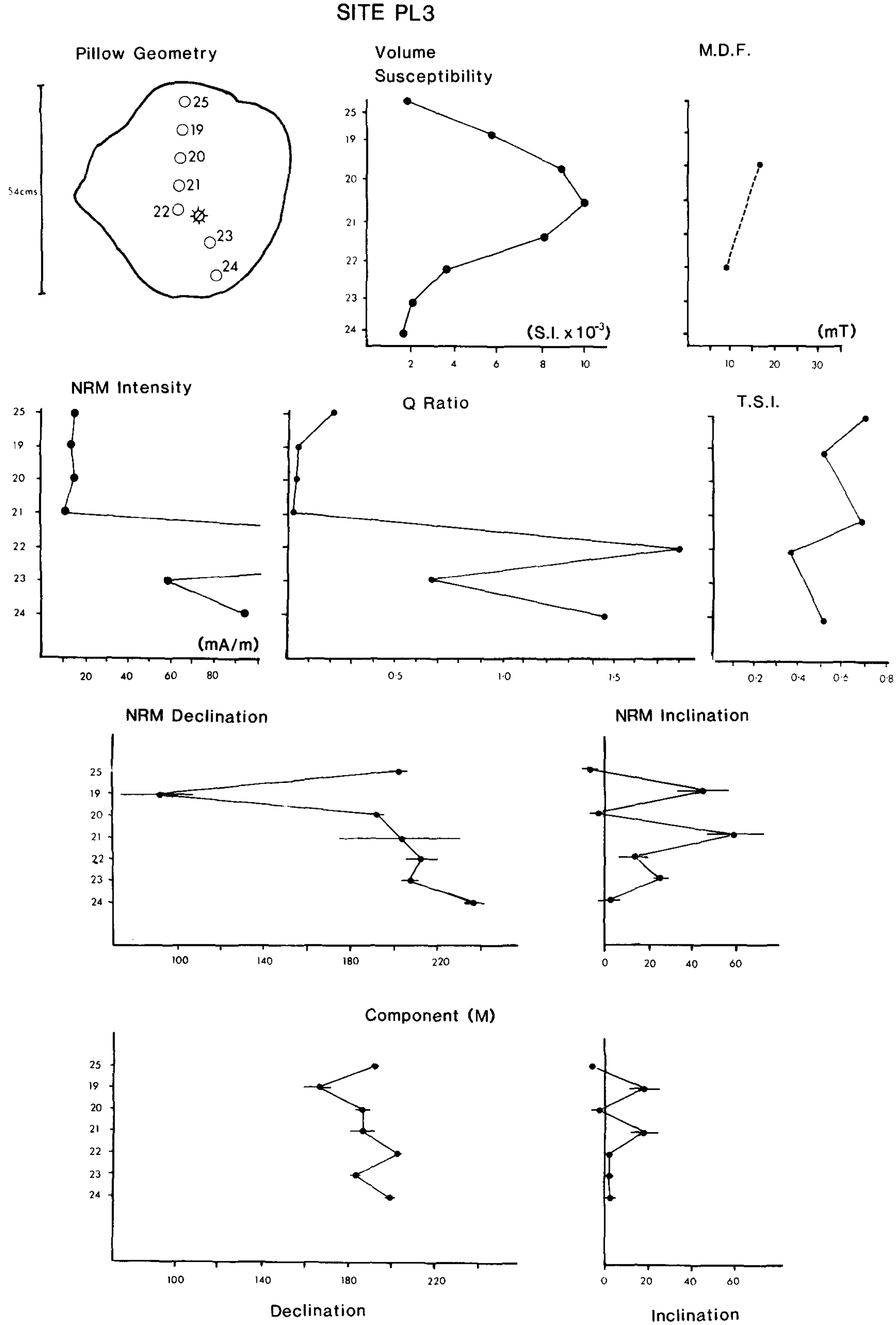


Fig. 4. Magnetic property variations within pillow lava site PL3. See Fig. 2 for explanation of individual graphs. NRM intensity of sample 22 (off-sale) is 355 mA/m.

equivalent to those previously described. Iron (titanium) enrichment appears as a brightening (darkening) of the backscattered electron image. Hematite is prevalent within the silicate groundmass appearing as bright, micron-sized flecks.

Pillow lavas

Opaque phases proved rare within the pillow lava and occur only at fine grain sizes of less than 5 μm. EDA spectra suggest these grains to be exclusively pyrite and chalcopyrite. As a stable magnetisation is observed from the pillow lavas, we deduce the remanence to be carried by a submicron-sized fraction visible only as a speckling of the groundmass in backscatter image.

Reflected light optical microscopy

Reflected light microscopy was performed to determine the degree of deuteric and low-temperature oxidation affecting the opaque phases. Samples from massive, pillowed (site PL2), and reddened lavas were studied.

The massive basaltic lava contains opaque titanomagnetite grains up to 100 μm in size, comprising exsolution lamellae of anisotropic ilmenite (Fig. 6a, b). Deuteric oxidation is at class II of Ade-Hall et al. (1968). Individual grains commonly show shrinkage cracking, indicating low-temperature oxidation to stage 2 (Johnson and Hall, 1978) and granulation (Ade-Hall et al., 1968,

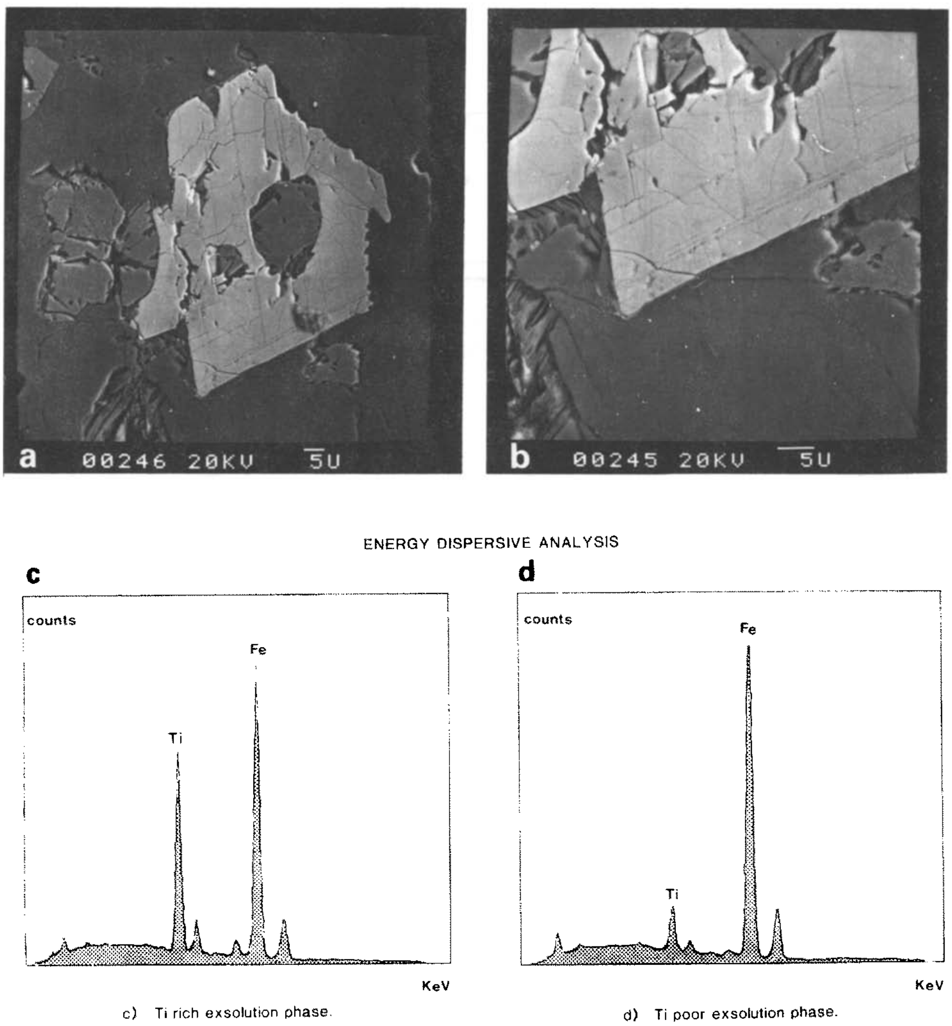


Fig. 5. (a) Backscatter electron photomicrograph of an Fe–Ti oxide grain within a massive basalt at Slockenray. Scale bar is 5 μm. (b) Bottom left-hand corner of same grain showing dark and light-coloured exsolution lamellae. Intra- and inter-grain cracking is also visible. EDA spectra for dark and light "bands" were collected and are shown in (c) and (d) respectively.

1971), which may be linked to hydrothermal alteration.

The reddened lava top shows the remnants of high-temperature oxidation but titanomagnetite grains are extensively corroded within a red hematite matrix. Some grains show substantial replacement by non-opaque phases up to low-temperature alteration stage 5 of Johnson and Hall (1978) (Fig. 6c, d).

The dominant opaque phases in the sampled pillow lava are isotropic sulphide grains that are typically less than 10 μm, but occasionally up to

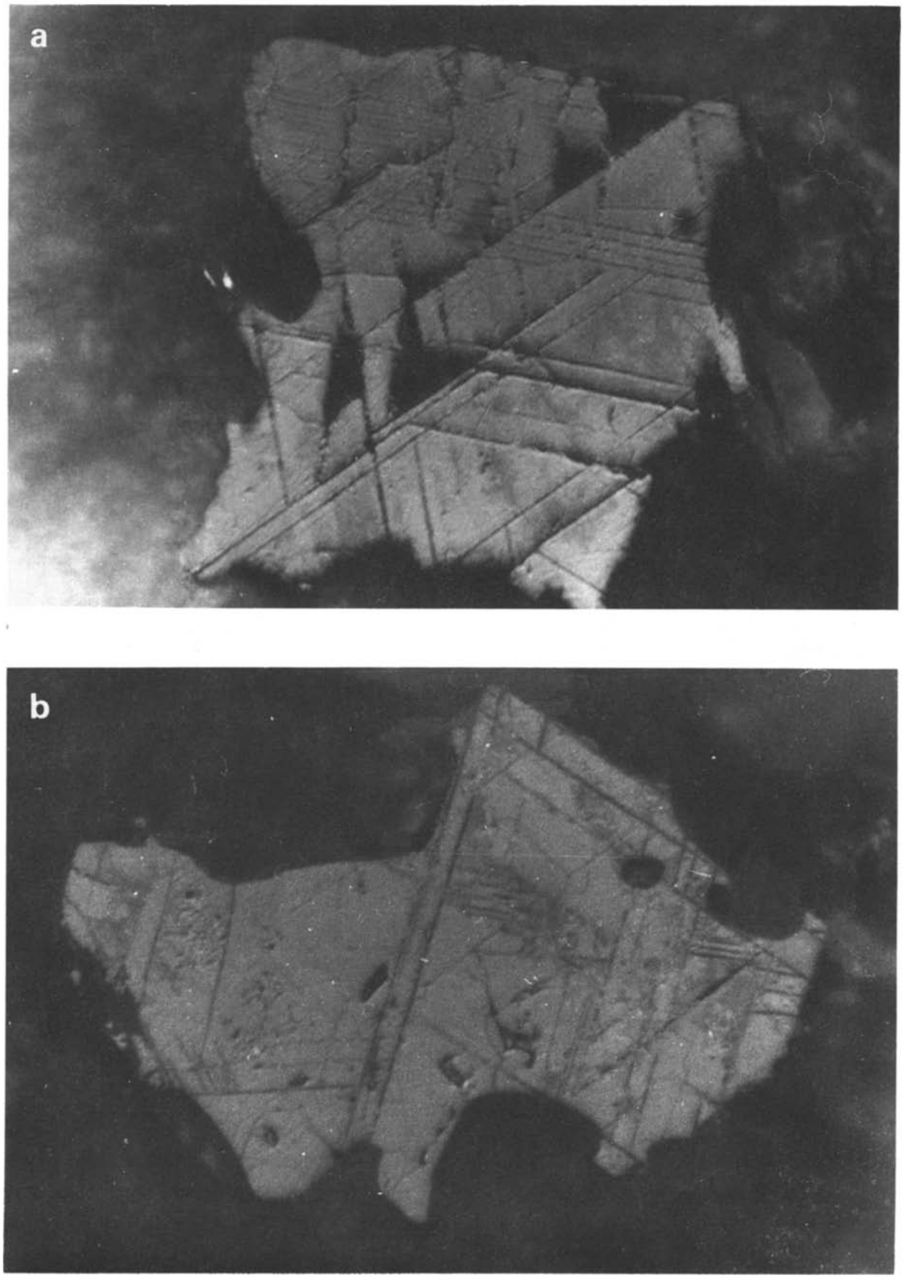


Fig. 6. (a, b) Exsolution lamellae of ilmenite and magnetite to oxidation state 2 of Ade-Hall et al. (1968). Massive lava, Slockenray. Plate widths: a = 130 μm, b = 95 μm. (c, d) Low-temperature alteration of Fe–Ti oxide grains containing relict exsolution lamellae. Reddened flow top, Slockenray. Plate widths: c = 120 μm, d = 130 μm.

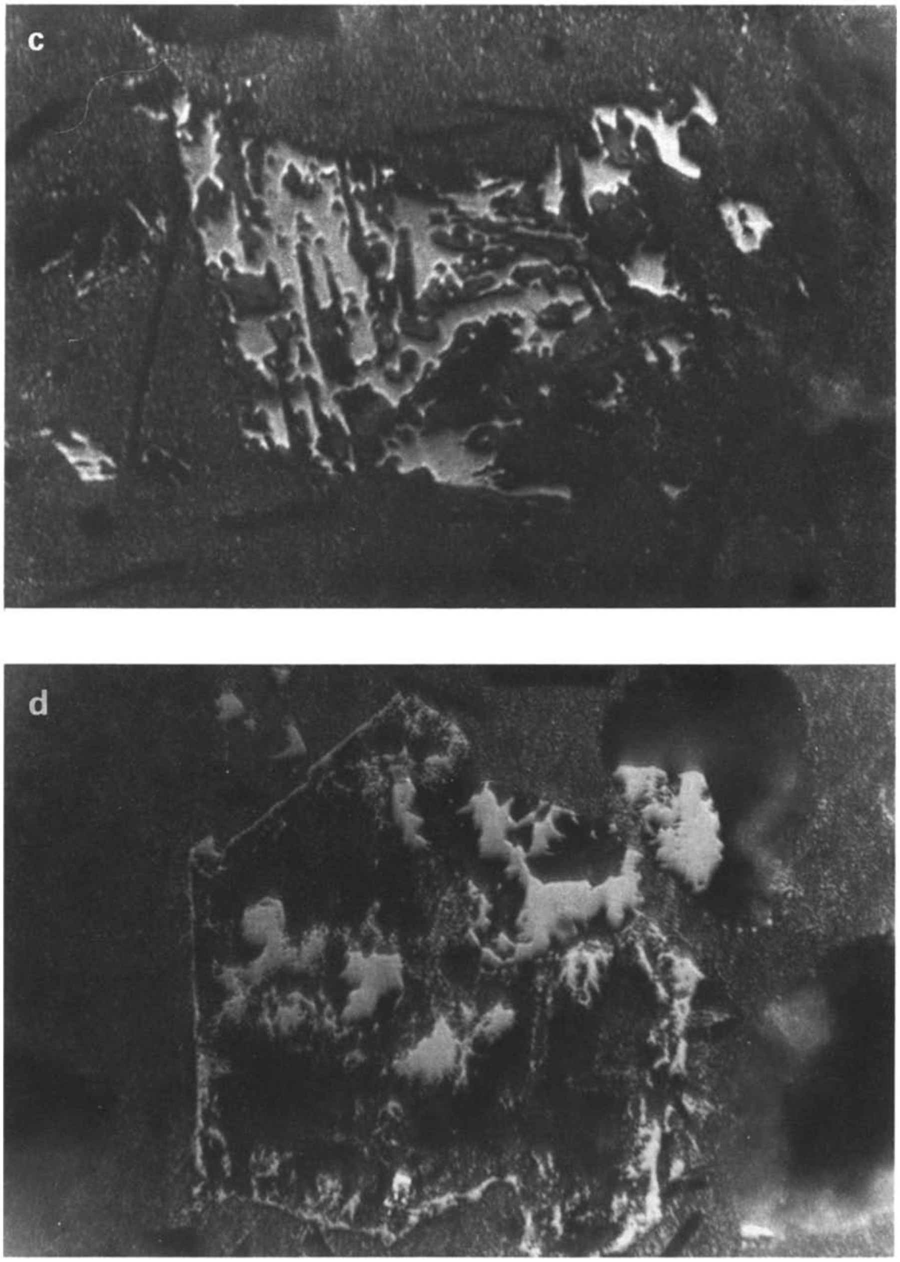


Fig. 6 (continued).

100 μm. Opaque grains near the limit of optical resolution were also observed but not classified.

Thermomagnetic analysis

Saturation magnetisation versus temperature characteristics were investigated using a horizontal translation balance. Samples were heated in air and then cooled from temperatures between 700° and 750°C. Once again, samples from a massive flow, pillowed flow and reddened lava top were processed. Representative thermomagnetic curves for these samples are shown in Fig. 7 together with their respective remanence decay curves during thermal demagnetisation. The characteristic J_{sat} versus T curves are described individually below:

(1) The thermomagnetic curve for the massive lava flow (Fig. 7a) is reversible and shows a single Curie temperature approaching 580°C. This suggests a low-Ti titanomagnetite close to pure magnetite to be the principal magnetic constituent in the lava. Maximum unblocking temperatures in excess of 500°C during thermal demagnetisation are in accordance with this observation.

(2) Thermomagnetic analysis of a sample from

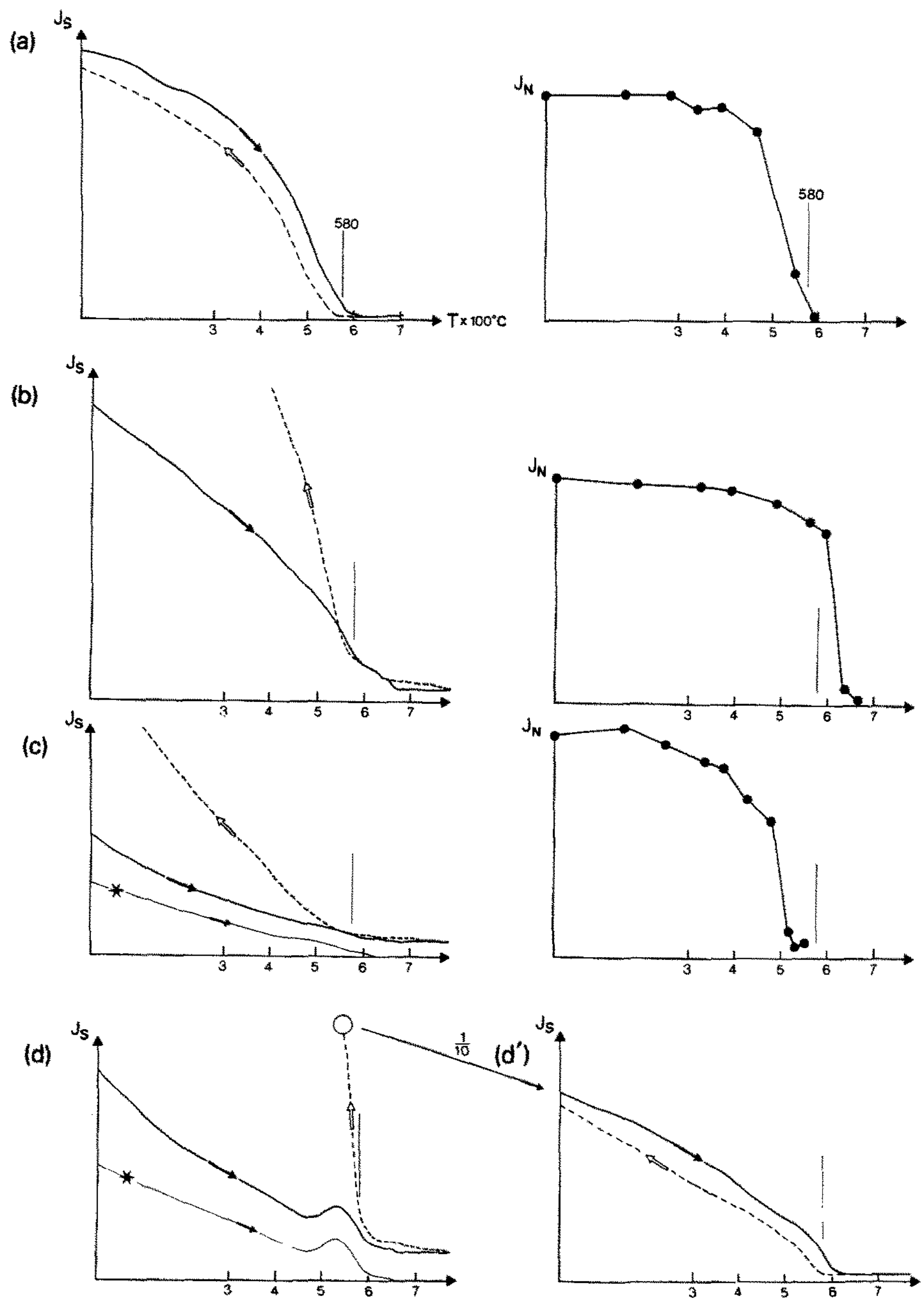


Fig. 7. Thermomagnetic analyses (J_S vs. T) and thermal demagnetisations (J_N vs. T) from the Slockenray Formation. (a) Massive lava flow. (b) Reddened lava top. (c, d) Pillow lavas ("starred" curves after subtraction of the paramagnetic component). (d') Re-run daughter product from (d).

the reddened lava top (Fig. 7b) is irreversible and shows two Curie temperatures of approximately 570°C and 650°C on heating. These are attributed to (low-Ti) magnetite and hematite phases, respectively. A significant increase in saturation magnetisation upon cooling indicates additional production of magnetite, possibly by reduction of hematite during heating or the thermal breakdown of non-magnetic phases to form magnetite. Thermal demagnetisation confirms the presence of hematite with unblocking temperatures persisting above the Curie point of magnetite.

(3) Pillow lava samples (adjacent to site PL2) show two types of behaviour during heating/cooling, both of which produce irreversible curves (Fig. 7c, d). The heating curves in Fig. 7c and d are shown both before, and after (starred), subtraction of the paramagnetic component.

The curve in Fig. 7c is dominated by the paramagnetic contribution and hence the magnetic phases are difficult to resolve. A proportion of magnetisation remains to temperatures around the Curie temperature of magnetite however. Unblocking temperatures greater than 500°C on thermal demagnetisation suggest the presence of low-Ti magnetite as the dominant magnetic phase.

The curve in Fig. 7d is similar to that observed in submarine lavas (e.g. fig. 5b of Smith and Banerjee, 1985b) where the characteristic magnetic mineral is a low-temperature oxidized titanomagnetite or titanomaghemite. This phase is metastable during heating and inverts to form separate Ti-rich and Ti-poor phases close to ilmenite and magnetite end member compositions. The formation of magnetite is indicated by the increase in saturation magnetisation at 500°C on heating and during subsequent cooling. Conversion to low-Ti magnetite is confirmed by the reversible 580°C Curie temperature observed on reheating of the sample (Fig. 7d′). The tenfold increase in saturation magnetisation may not indicate magnetite formation solely from inversion of titanomaghemite. Oxidation of iron-sulfides perhaps combined with titanomaghemite inversion could account for this however. Iron-pyrites and chalcopyrite are clearly observed during SEM and reflected light microscopy.

Discussion.

Magnetic mineralogy.

The combined use of microscopic and rock magnetic techniques leads us to suggest that three principal magnetic phases are present in the lavas. These are: (1) low-Ti titanomagnetite, almost pure magnetite in composition, (2) hematite, and (3) titanomaghemite. Of these, only the last is routinely reported from in-situ oceanic crust (e.g. Johnson and Hall, 1978; Marshall, 1978). Low-Ti titanomagnetite has been reported from deeply buried oceanic lavas (Smith and Banerjee, 1985b) and from postulated shallow-water extrusives however (Cockerham and Hall, 1976), but appears more commonly in the lavas of ophiolites (e.g. Levi and Banerjee, 1977; Swift and Johnson, 1984). The reason for its occurrence within ophiolites may be two-fold. Butler et al. (1976) suggest its formation by inversion of titanomaghemite during hydrothermal circulation at the spreading centre. Levi et al. (1978) suggest metamorphic heating during or after ophiolite obduction. Whilst the former explanation predicts that zones of low-Ti titanomagnetite and titanomaghemite should be preserved within a lava sequence, the latter predicts a uniform distribution of low-Ti titanomagnetite throughout the lava pile. As we observe an intimate occurrence of both titanomaghemite and low-Ti titanomagnetite, we prefer the former explanation in this case. Indeed, hydrothermal circulation within the Slockenray succession has previously been proposed purely on geological grounds (Lewis, 1975). Hematite has not been previously reported as a significant remanence-carrying phase within lavas of other ophiolites or from oceanic basalts. Its presence as an alteration product has been noted, however (Alt et al., 1984; Hall and Fisher, 1987).

All three magnetic phases are explicable as an oceanic magneto-mineralogy when rock magnetic, palaeomagnetic and geological evidence is combined. The preservation of titanomaghemite implies that temperatures in excess of approximately 150°C have not affected parts of the sequence since its formation (Johnson and Merrill, 1973;

Levi et al., 1978). This is consistent with only zeolite facies metamorphism having affected the sampling area (Oliver et al., 1984). Remagnetisation of the Slockenray Conglomerate may therefore represent a local, rather than a pervasive hydrothermal effect (cf. Trench et al., 1988).

Magnetite–ilmenite exsolution lamellae are observed from a massive basalt and a reddened lava top. Whilst deuteric oxidation is a common feature of subaerial lavas, it occurs only rarely within submarine basalts where it has been linked with shallow eruption (Cockerham and Hall, 1976). In this respect, there are several lines of geological evidence to suggest shallow-water eruption for the Slockenray lavas. These include the considerable thicknesses of hyaloclastites, the presence of large-scale foresets and the occurrence of current-rounded conglomerate clasts (Bluck, 1982).

The presence of low-Ti titanomagnetite can thus be explained either as the product of local hydrothermal inversion of titanomaghemite and/or as a deuteric oxidation product within the Slockenray lavas. In both cases, this would support the contention of a pre-obduction remanence previously proposed on palaeomagnetic grounds (Trench et al., 1988). A post-obduction metamorphic origin for the (low-Ti) magnetite is untenable given the local survival of titanomaghemite within the succession and the low metamorphic grade. The presence of grain size-related magnetic properties in the pillow lavas also suggests retention of an original magneto-mineralogy and argues against post-obduction metamorphism.

Amorphous hematite is abundantly present in the matrix of the visibly reddened lava top. The iron–titanium phases from this lava top are extensively corroded and only relict exsolution textures are preserved. This corrosion suggests extensive low-temperature alteration of the lava to have occurred. Hematite and low-Ti titanomagnetite phases carry an identical remanence direction within the Slockenray Formation and imply their magnetisation to have been acquired near-synchronously (Trench et al., 1988). The palaeomagnetic and microscopic observations therefore support hematisation by sub-aerial exposure during Ordovician times (Smellie, 1984b).

Reliability criteria

The application of *palaeomagnetic* criteria to the Ballantrae studies is straightforward (Table 1). However, the *overriding* criterion in any palaeomagnetic assessment requires that the studied rocks, when magnetised, formed a rigid part of the unit for which an APWP is being constructed. Thus, a "quality factor" of 5 for the Slockenray Formation (Table 1) implies the data to be *reliable* but of only moderate *significance* in large-scale APWP construction.

New geochronological data have become available for the Byne Hill Gabbro since the first palaeomagnetic results were reported by Piper (1978). Bluck et al. (1980) report a U–Pb age of 483 ± 4 Ma for the Byne Hill trondhjemite which is transitional with the gabbro. The rock age *is* therefore known "to within half a geological period" and now satisfies this criterion in the Van der Voo (1988) scheme. This point was overlooked by Stearns et al. (1989) and has been amended in Table 1.

Assessment of the Ballantrae extrusive layer in terms of Levi et al's (1978) *rock magnetic* constraints is problematical as our results bridge each of their specified minimum criteria (Table 2). With the recent recovery of low NRM intensities, Koenigsberger ratios less than unity, and reversible thermomagnetic curves on deeper sampling of the oceanic crust (e.g. Smith, 1985; Smith and Banerjee, 1985b), these minimum criteria could justifiably be lowered. In so doing, they would lose much of their usefulness in determining "representative" from "altered" ophiolite sequences, however. We therefore contend that whilst the Slockenray extrusive layer has maintained its original oceanic magnetic properties, these are somewhat atypical to those specified by Levi et al. due to the shallow-water depositional environment and the effects of a hydrothermal system on parts of the Slockenray sequence. Notably, all lithologies provided good quality demagnetisation results irrespective of their contrasting Koenigsberger ratios (Table 2).

Concluding remarks

The combination of palaeomagnetic, rock magnetic and geological observations yields a better understanding of the magnetisation in the extrusive layer of the Ballantrae ophiolite than could have been obtained using any one technique in isolation.

Intra-pillow magnetic property variations, the occurrence of titanomaghemite, low-Ti titanomagnetite and hematite are all explained as original facets of ocean crust magnetisation processes. These observations support the arguments for a pre-obduction remanence, initially developed on palaeomagnetic grounds. In the light of this new evidence, remagnetisation of the Slockenray conglomerate is now better interpreted as a "local hydrothermal" rather than "pervasive" overprinting.

The study of magnetic properties within pillow lavas provides an additional field test in the assessment of remanence stability. If systematic trends concur with pillow lava geometry, then an original grain size variation may be present. This precludes complete recrystallisation of magnetic phases by post-obduction metamorphism in this case.

Palaeomagnetic reliability criteria are easily applicable to the palaeomagnetic study of ophiolitic rocks. A pre-obduction remanence is of limited use in APWP construction however as it positions only the ophiolite terrane itself. The single most important palaeomagnetic reliability criterion therefore concerns the possibility of relative rotation or translation of the study area with respect to its host continent.

The application of *rock magnetic criteria* to studies of ophiolite magnetism can be ambiguous. This reflects an increased diversity of magnetic properties now reported from directly sampled ocean floor basalts. The integrated approach adopted here is therefore preferred to determine the reliability of ophiolite magnetic properties.

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