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Palaeomagnetism and isotopic age data from Upper Cretaceous igneous rocks of W. Portugal; geological correlation and plate tectonic aspects

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Summary. Palaeomagnetic results and K–Ar age data for the Sintra and Sines intrusive complexes (W. Portugal), and further details on the palaeomagnetic structure of the Lisbon volcanics are reported. The Sintra complex consists of two main intrusive phases having been emplaced in the Upper Cretaceous at around 90 Ma and 75 Ma respectively. The radiometric results show that the Sines complex formed concurrently with the second Sintra magmatism. The early (main) Sintra pluton has a characteristic magnetization of $D = 358^\circ$, $I = 27^\circ$ ($\alpha_{95} = 3.3^\circ$). This remanence direction is defined by gabbros and diorites as the granitic rocks (constituting the bulk of the complex) are shown to possess stable secondary magnetization imposed during Quaternary weathering. The characteristic magnetization of the younger intrusive event, as defined by the Sines rocks, has a mean direction of $D = 041^\circ$, $I = 41^\circ$ ($\alpha_{95} = 3.3^\circ$). These two Upper Cretaceous palaeomagnetic directions are significantly different at the 95 per cent probability level. The Lisbon volcanics show the presence of the '75 Ma' magnetization suggesting that also this volcanic complex dates from the Upper Cretaceous. In addition, the palaeomagnetic results from the Lisbon complex have given further substance to a previously reported magnetization component with shallow inclination and north–northwest declination, now defined by $D = 333^\circ$ $I = 14^\circ$ ($\alpha_{95} = 7.4^\circ$). It is inferred that this magnetization which has a dual-polarity structure formed through low temperature oxidation in late Cretaceous to Lower Tertiary time. The declination difference between the '90 Ma' and

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'75 Ma' magnetizations may be interpreted in terms of a two-phase rotation of Iberia in the Upper Cretaceous prior to the assumed late Cretaceous–early Tertiary secondary magnetic imprint, implying a counterclockwise rotation of minimum 40° followed by a clockwise rotation of similar magnitude. However, if the relevant palaeomagnetic reference frame for Iberia, before and after its independent tectonic movements in the late Cretaceous, is the African plate, then the clockwise rotation is about 20° larger than the counterclockwise one, apparently accounting for the compressive tectonism in the Bay of Biscay.

Key words: palaeomagnetism, K–Ar ages, Cretaceous, Portugal, tectonomagmatic implications

Introduction

Post-Hercynian palaeomagnetic data from Spain and Portugal have traditionally been associated with a counterclockwise rotational opening of the Bay of Biscay (Schwarz 1963; Van Dongen 1967; Girdler 1968; Watkins & Richardson 1968a; Van der Voo 1969; Stauffer & Tarling 1971; Van der Voo & Zijderveld 1971; Storetvedt 1973; Van den Berg 1979), and the Upper Cretaceous has repeatedly been considered as the time during which Iberia detached from the European plate (Van der Voo 1969; Stauffer & Tarling 1971; Williams 1975; Van den Berg 1979; Boillot 1984). However, despite the many geophysical and geological studies, the tectonic setting of Iberia is far from clear, and in a recent discussion on the tectonic evolution of the Pyrenees (see McCaig & Wickham 1984) the need for further detailed palaeomagnetic studies of the Cretaceous of Iberia was stressed. The present paper contributes to this task, reporting new palaeomagnetic and radiometric results from the late Cretaceous Sintra and Sines intrusive complexes of W. Portugal as well as giving further details on the palaeomagnetic structure of the Lisbon volcanics. The latter formation has previously been subjected to palaeomagnetic studies by Watkins & Richardson (1968a), Van der Voo (1969), Van der Voo & Zijderveld (1971) and Storetvedt (1973), but the geophysical interpretation of the results has become a matter of debate (Van der Voo 1968; Watkins & Richardson 1968b; Storetvedt 1970; Watkins & Richardson 1971). The Sines intrusive complex has not previously been studied palaeomagnetically, but results from eight sites of the Sintra granite have been reported (Van der Voo 1969).

The new investigation appears to have clarified some of the controversial magnetization aspects posed by earlier studies of the Lisbon volcanics as well as giving support to preliminary radiometric evidence for a late Cretaceous age of this complex. Furthermore, the results from the Sintra and Sines complexes are very clearcut and suggest that the tectonic setting of Iberia may differ significantly from that previously advocated.

Geological setting

SINTRA AND SINES INTRUSIVE COMPLEXES

The Sintra igneous massif (Fig. 1b), consisting of granites, syenites and rocks of basic composition, forms the core of the Sintra Hills of W. Portugal. Aeromagnetic mapping (Mendes-Victor, private communication) shows that about 1/3 of its total areal extent is covered by the sea. The plutonic mass is approximately elliptical in shape (15 x 5 km) and has a nearly E–W orientation.

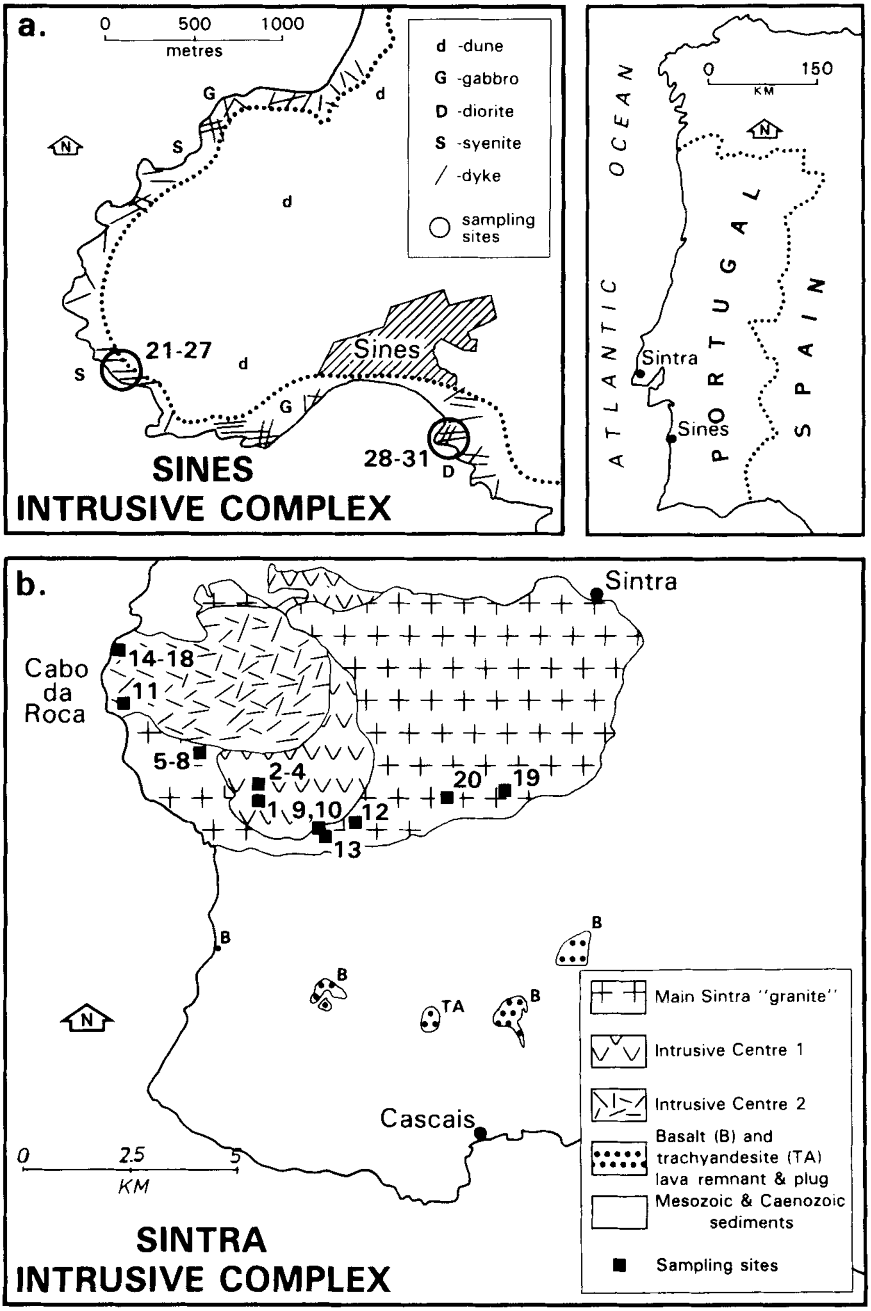

Figure 1. Geological sketch maps of the Sines (a) and Sintra (b) intrusive complexes, simplified after Canilho & Abranches (1982) and Wright (1968), respectively. Sampling locations are numbered.

The igneous emplacement has domed a thick sequence of shallow-water calcareous sediments of mid-Jurassic to late Cretaceous age. The tectonic process responsible for the intrusion generated an asymmetrical dome with overturning of the north limb, the surrounding sediments forming a ring-shaped syncline (Ribeiro 1979). The youngest sediments involved in the doming process are of Upper Cenomanian age, and from field evidence the upper age limit of the Sintra pluton is provided by Oligocene conglomerates, exposed on the north flank and containing pebbles derived from the intrusive body (Torre d'Assunção & Brak-Lamy 1952).

The main part of the Sintra massif consists of granite but in its western areas later intrusive bodies of syenitic and basic compositions occur. This main plutonic division is hereafter called the Sintra complex. Wright (1968) distinguished between two main phases of younger magmatic activity (including sub-volcanic breccias), Centres 1 & 2 (Fig. 1b), but from the present study (see below) it is only the latter intrusive centre, here named the Cabo da Roca complex, that appears to be significantly younger than the Sintra pluton. The plutonic rocks and their host sediments are cut by a dense network of dykes and sills. Radiometric age dating (see below) concurs with the field evidence, arriving at an Upper Cretaceous (Turonian–Senonian) age for the Sintra and Cabo da Roca intrusive events.

The Sines complex, located some 105 km south–southeast of the Sintra hills (see Fig. 1a) shows several characteristic features similar to the Sintra massif, i.e. elliptical shape, nearly east–west orientation, sub-volcanic ring-structure (Teixeira 1962; Ribeiro 1979) etc. The major part of the complex lies offshore. The subaerial part is largely covered by Quaternary deposits, the exposed rocks constituting only a narrow littoral section. Both on the north and south side of the complex thermal aureoles have been generated in the adjacent sedimentary rocks, causing various types of thermometamorphism (Canilho 1971). The most important petrographic types are gabbro, diorite and syenite. The complex is cut by numerous dykes of varied composition. Radiometric studies (see below) suggest that the Sines intrusive mass formed contemporaneously with the Cabo da Roca complex. This seems to fit the hypothesis of Ribeiro (1979) that the Sines and Sintra intrusions are connected by a common strike-slip fracture. These complexes form part of the late Cretaceous alkaline igneous province of Iberia (Rock 1982).

THE LISBON VOLCANICS

The Lisbon volcanic complex (Fig. 2) is situated east of the Sintra massif, forming a north–south trending belt of isolated volcanic vents, cutting across Cenomanian sedimentary strata, and numerous lava flows alternating with less abundant tuff layers. There is also a dyke network, almost exclusively vertical, and a few late stage volcanic vents piercing lava flows.

Petrographically, the Lisbon volcanics are mostly composed of basaltic rocks, but hawaiites and mugearites are also present. Coarser grained varieties include ankaramites and ankaratrites, and aphanitic rocks include basanites and basanitoides (Alves *et al.* 1980).

The volcanic complex comprises at least 15 eruptive centres. Their pipes (plugs, necks and breccias) and part of their cones are still visible, forming an almost continuous belt between the Cascais/Oeiras and Mafra/Malveira districts. Reconstruction of some of the more pronounced volcanic centres suggests heights of the volcanic edifices of the order of 2.5 km, with cones of about 20 km in radius and average flow dips of around 8°. Due to later erosion the volcanic vents have been separated from the lava fields.

The flows which are all of subaerial origin rest on Cenomanian limestones, but they are overstepped by the conglomerates of the Benfica formation (Eocene–Oligocene). The pyroclastic horizons, constituting lapilli and ash, are, with few exceptions, relatively thin (0.8–2.0 m) in comparison with the lavas for which thicknesses normally vary between 5 and 10 metres. The pyroclastic horizons, originally probably covering the entire area of the volcanic edifices, overlie deeply weathered basalts and/or palaeosoils. Each tuff/palaeosoil layer nearly always covers a sequence of lava flows that are relatively well preserved at the

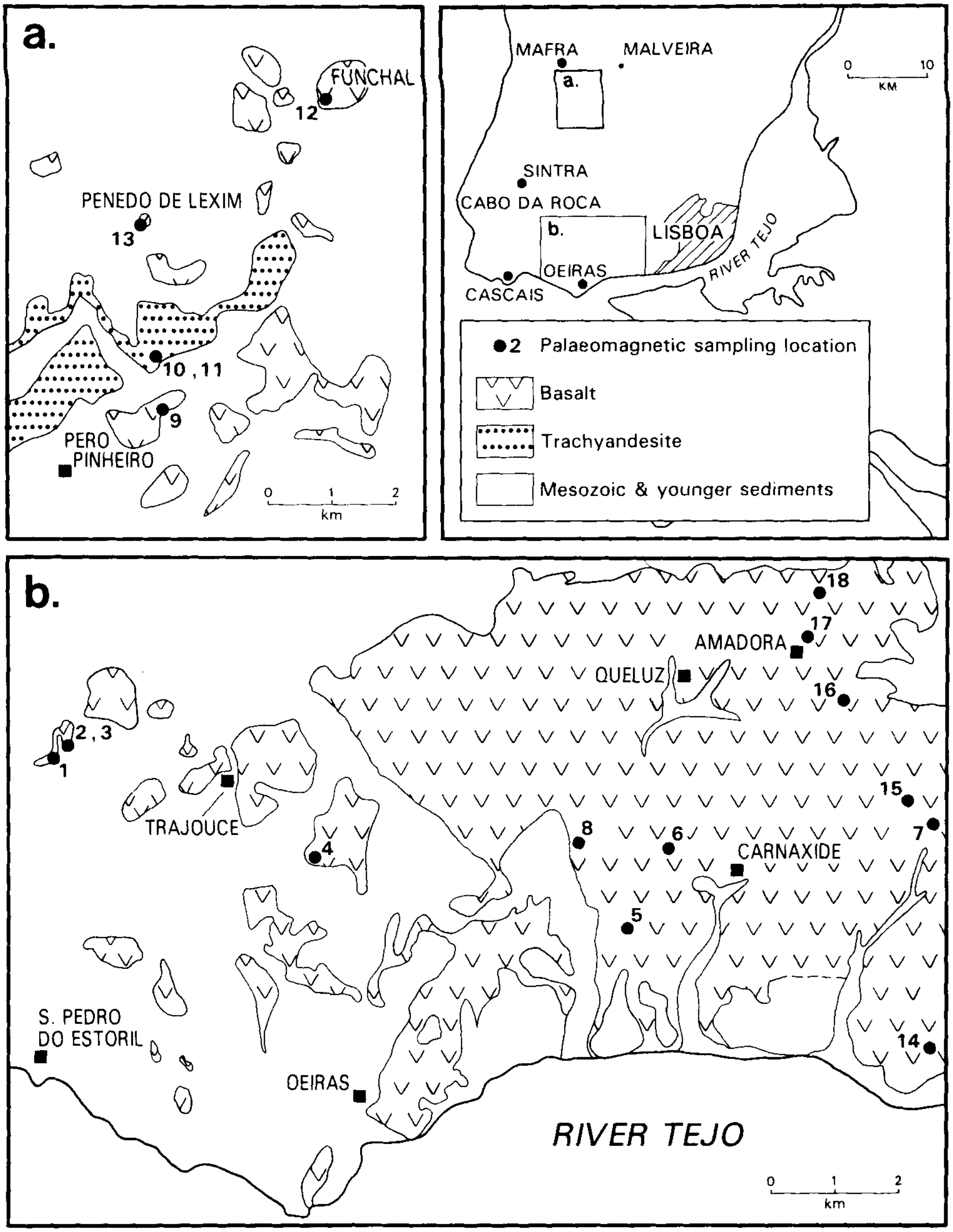

Figure 2. Simplified maps showing the distribution of the Lisbon volcanics in the two specific areas covered by the present palaeomagnetic sampling. Numbered points refer to sampling sites.

base but progressively more weathered towards the top. This suggests strongly that the Lisbon volcanism developed as a series of episodic events separated by longer quiescent periods (Serralheiro 1978).

The volcanic complex has been disrupted by faulting, but on the whole the lava pile is practically horizontal (dip $< 10°$), and it must be recalled that the average original flow dips are *c.* 8°. Near major faults however, flexuring of the lava sequence has created dips ranging upwards to 80°. This folding includes the overlying Eocene–Oligocene strata and is therefore probably an integral part of the late Tertiary (Miocene and younger) orogenic movements in the Iberian domain.

Freshwater fossils in intercalated sediments of the volcanic sequence cannot fix the age more precisely than between Upper Cretaceous (Senonian) and Eocene (Zbyszewski & Assunção 1965; Ribeiro, private communication), but preliminary radiometric studies (Ferreira & Macedo 1979; Abranches, unpublished results) suggest Upper Cretaceous ages for the Lisbon igneous complex (see below).

Sampling and laboratory procedures

The rock material for the present study was collected during two field seasons; Sintra, Cabo da Roca and Sines complexes in 1980 and the Lisbon volcanics in 1984. Due to break-down of the drilling equipment sampling of most Sintra and Sines sites (except Sintra sites 2–7) was through hand samples. In the Sintra and Cabo da Roca complexes the granitic rocks show a higher degree of visible weathering than the rest of the sampled material. The bulk of the samples from the three plutonic complexes have been obtained by both magnetic and sun compass orientation methods. The average declination based on the two types of reading is 3°W as compared with the regional declination of around 8°W. This difference is attributed to the pronounced magnetic field anomalies surrounding these intrusive masses. A few samples have only been magnetically oriented, and these have therefore been subjected to a declination correction of 3°W.

The material from the Lisbon volcanics was sampled by field drill and for orientation a sun compass was used throughout. Apart from site 18, where local flexuring has tilted the volcanics by about 58° in a southwesterly direction, the tectonic tilting of the sampled sites is minimal. Based on previously encountered magnetization complexities due to inferred long-lasting low temperature oxidation (Storetvedt 1970, 1973) a strict sampling criteria, accepting only macroscopically fresh samples, was adopted. The deep weathering of the lava pile eliminated many potential sampling locations, and without a field drill fresh lava samples would have been very difficult to obtain. On the other hand, most volcanic plugs (not previously studied) had an unaltered appearance in the field.

Further rock collection details are given in Table 1.

In the laboratory cylindrical specimens, diameter and height of 1.9 cm, were cut for palaeomagnetic measurements, and the remaining material from 12 of the sampled Sintra and Sines blocks were used for potassium–argon age determination. For the radiometric studies the fine grained diabases, andesites and some of the syenites were analysed as 'whole rocks', while one granite and two syenite samples were dated by separating a K-feldspar fraction. Whole rock analyses were performed by crushing the samples and taking the number 20–40 fraction in order to homogenize the material prior to argon analysis. Part of this material was then reduced to about number 200 size for potassium analysis. K-feldspar was separated from the granite and syenite samples by an electromagnetic separator following crushing to number 40–80. Potassium contents were determined by flame photometry on an EEL 450 instrument with lithium internal standard. Argon isotopic analysis were performed by isotope dilution on a modified MS 10 mass spectrometer, using an ^{38}Ar 'spike'. Gas extractions were by RF induction heating and cleaning with Ti getters, the extraction system being coupled directly to the mass spectrometer.

For measurements of remanent magnetization a shielded version Digico spinner magnetometer (Sintra and Sines intrusives) and a Molspin magnetometer (Lisbon volcanics) have

Table 1. Details of sampling sites. See Figs 1 and 2 for locations.

Site no.	Sample nos	Rock type	Site no.	Sample nos	Rock type
SINTRA (Fig. 1b)					
1	Si 1–6	Syenite (fine grained)	11	Si 58–62	Granite
2	7–11	Gabbro	12	63–65	Granite
3	12–17	Gabbro	13	65–68	Syenite
4	18–23	Gabbro	14	69–72	Gabbro
5	24–28	Gabbro (coarse grained)	15	73–76	Gabbro
6	29–33	Syenite	16	77–79	Diabase
7	34–43	Gabbro	17	80–83	Gabbro
8	44–48	Syenite (fine grained)	18	84–87	Syenite
9	49–52	Diorite	19	88–90	Granite
10	53–57	Diabase	20	91–93	Granite
SINES (Fig. 1a)					
21	Sin 1–3	Syenite	27	Sin 20–22	Diabase
22	4–6	Diabase	28	23	Diorite (fine grained)
23	7–9	Syenite	29	24–26	Diorite
24	10–13	Diabase	30	27–29	Diorite
25	14–16	Diabase	31	29–31	Diorite (fine grained)
26	17–19	Syenite			
LISBON VOLCANICS (Fig. 2)					
1	LV 1–5	Basaltic dyke	10	LV 46–50	Trachyandesite sill
2	6–8	Basaltic plug	11	51–55	Trachyandesite sill
3	10–13	Basaltic plug	12	56–60	Basaltic plug
4	14–19	Basaltic plug	13	61–65	Basaltic plug
5	20–24	Basaltic plug	14	66–72	Lava flow
6	25–29	Basaltic flow	15	73–77	Lava flow
7	30–35	Basaltic flow	16	78–82	Lava flow
8	36–40	Basaltic flow	17	83–87	Lava flow
9	41–45	Basaltic flow	18	88–97	Lava flow

been used. Alternating field (AF) demagnetization was carried out using a two-axis tumbler and for thermal demagnetization a Schonsted TM-1 was employed.

In an attempt to define remanence components other than stable end points, careful vector analysis has been carried out on all specimens defining great circle trends on demagnetization.

Radiometric ages and their interpretation

Conventional potassium-argon analyses, following the procedures outlined above, have been carried out on twelve samples from the Sines, Sintra and Cabo da Roca complexes. The mean potassium and argon analyses are included in Table 2 along with the mean age and one standard deviation error based on the analytical reproducibility.

Of the seven Sines samples studied four come from two separate diabase dykes (sites 22 and 24). The two site 24 samples show chemical variation manifested by a 25 per cent difference in their K_2O's. Their ages are concordant, however, at close to 75 Myr. On the other hand, the two site 22 samples exhibit only a 10 per cent difference in K_2O but yield discordant K–Ar ages of 62.0 ± 1.3 and 78.2 ± 1.7 Myr, respectively. The concordancy of

Table 2. Radiometric age data from the Sintra and Sines complexes. Below the new Sintra and Sines results earlier radiometric data from these formations are listed for comparison.

Site no.	Rock	Method	K_2O (wt per cent)	Rad. $^{40}Ar(mm^3 gm^{-1})$	Atm.conta. (per cent)	Age ± 1σ Myr	References
SINTRA COMPLEX							
20	Granite	K–Ar, K-feldspar	4.68 ± 0.01	$(1.298 \pm 0.016)10^{-2}$	18.4	84.0 ± 1.1	Present study
6	Syenite	K–Ar, whole rock	4.89 ± 0.03	$(1.231 \pm 0.020)10^{-2}$	16.8	76.4 ± 1.4	Present study
Previous radiom.		Rb–Sr, biotite and whole rock				88.7 ± 0.6	Mendes (1968)
results from		Rb–Sr, biotite				88 ± 8 ≠	Bonhomme *et al.* (1964)
Sintra		Rb–Sr, isochron				95 ± 4	Abranches & Canilho (1982)
		K–Ar, biotite				81.9 ± 0.4	Macintyre & Berger (1982)
CABO DA ROCA complex							
17	Gabbro	K–Ar, whole rock	1.92 ± 0.01	$(4.74 \pm 0.06)10^{-3}$	23.9	74.9 ± 1.0	Present study
18	Syenite	K–Ar, whole rock	5.54 ± 0.01	$(1.39 \pm 0.02)10^{-2}$	52.4	76.1 ± 1.1	Present study
18	Syenite	K–Ar, whole rock	5.42 ± 0.11	$(1.40 \pm 0.02)10^{-2}$	60.9	78.3 ± 1.9	Present study
SINES complex							
24	Diabase	K–Ar, whole rock	0.591 ± 0.002	$(1.47 \pm 0.02)10^{-3}$	66.4	75.5 ± 1.1	Present study
24	Diabase	K–Ar, whole rock	0.476 ± 0.001	$(1.17 \pm 0.03)10^{-3}$	74.8	74.6 ± 1.9	Present study
22	Diabase	K–Ar, whole rock	0.551 ± 0.001	$(1.42 \pm 0.03)10^{-3}$	65.1	78.2 ± 1.7	Present study
29	Diorite	K–Ar, whole rock	4.34 ± 0.07	$(1.333 \pm 0.012)10^{-2}$	48.3	79.1 ± 1.5	Present study
26	Syenite	K–Ar, K-feldspar	5.26 ± 0.02	$(1.304 \pm 0.013)10^{-2}$	45.4	75.2 ± 0.8	Present study
22	Diabase	K–Ar, whole rock	0.609 ± 0.008	$(1.24 \pm 0.02)10^{-3}$	30.4	62.0 ± 1.3	Present study
23	Syenite	K–Ar, K-feldspar	5.64 ± 0.02	$(1.181 \pm 0.014)10^{-2}$	24.0	63.8 ± 0.8	Present study
Previous radiom.		K–Ar, whole rock				72.8	Canilho (1971)
results from		Rb–Sr, isochron				72 ± 1.5	Canilho & Abranches (1982)
Sines		Rb–Sr, biotite and whole rock				72.1 ± 0.7	Mendes (1968)

$\lambda_e = 0.581 \times 10^{-10} a^{-1}$; $\lambda_\beta = 4.962 \times 10^{-10} a^{-1}$; $^{40}K/K = 1.167 \times 10^{-2}$ atom per cent; ≠ age corrected to new decay constant, $\lambda = 1.42 \times 10^{-11} a^{-1}$.

the three diabase ages strongly suggests that argon loss is responsible for the discrepantly low age of the fourth sample. The three diabase ages are also concordant with the whole rock age of the site 29 sample investigated (diorite) and with the K-feldspar age for site 26 (syenite). The mean of these five ages from the Sines complex (concordant at the 2σ level) is 76.5 ± 2 Myr. A second K-feldspar analysis on a sample separated from the syenite of Site 23 yielded 63.8 ± 0.8 Myr. The coincidence of this K-feldspar age with the discordant dyke age (site 22) is most likely the consequence of a later overprinting event in these rocks.

We present five new radiometric dates from the Sintra massif, two from the main Sintra pluton (sites 6 and 20) and three from the inferred latest intrusive phase (Cabo da Roca complex). The Cabo da Roca samples which are of diverse petrological types (gabbro and syenite) yield concordant K–Ar ages (at the 2σ level) whose mean is 76.4 ± 2.9 Ma. This is in perfect agreement with the main Sines age group, as well as with the K–Ar data from Sintra site 6 (76.4 ± 1.4 Myr) which is located in close proximity to the Cabo da Roca complex (see Fig. 1). The second sample from the Sintra intrusion (site 20) which comes from a location that is much more distant from the Cabo da Roca complex, gave the significantly higher age of 84.0 ± 1.1 Myr. Bonhomme, Mendes & Valette (1961) have reported biotite 'model' ages for the Sintra granite using the Rb–Sr technique, and their mean age of 88 ± 8 Myr is indistinguishable from the feldspar K–Ar age from our site 20. On the other hand, Abranches & Canilho (1982) have defined a Rb–Sr whole rock isochron for the Sintra complex corresponding to an age of 95 ± 4 Myr, and Mendes (1968), using the same technique, arrived at an age close to 90 Myr. Thus, our K–Ar age of 84.0 ± 1.1 Myr is significantly younger than the Rb–Sr dates. In line with this pattern is the Macintyre & Berger (1982) K–Ar age of 81.9 ± 0.4 Myr on biotite from the Sintra granite. The fact that the Rb–Sr data tend to define an older age than do the K–Ar ages, and that one of our two K–Ar ages shows a perfect match with those of the younger Cabo da Roca complex (Intrusive Centre 2 of Fig. 1b) suggest argon loss in the Sintra complex during the later igneous event. The relatively high weathering disintegration of the granites may be another reason for the lower K–Ar age of site 20 as compared with the Rb–Sr ages for the Sintra complex.

Another possibility is that the Sintra pluton has suffered crustal contamination resulting in too high Rb–Sr ages. However, in the Sines intrusive province there is no such age discrepancy, and this agreement between the two dating methods apply also for the Monchique complex (see Macintyre & Berger 1982). On balance, a *c.* 90 Myr emplacement age for the Sintra granite seems most likely, but it is pertinent to stress that more radiometric work is needed before a precise age for the Sintra massif can be established. The ultimate question seems to be whether the age difference between the two intrusive events is $\leqslant$ 10 Myr, as suggested by the K–Ar dates (see also Macintyre & Berger 1982), or *c.* 15 Myr indicated by the present evaluation of all available radiometric age information. The small error in the Rb–Sr age of Mendes (1968) (88.7 ± 0.6 Myr), may be an important point in substantiating the '90 Myr' age for Sintra.

The extremely consistent K–Ar ages of the Cabo da Roca and Sines intrusions, at around 75 Myr, covering diverse petrological rock types, suggest strongly that these igneous 'units' are contemporaneous. These ages are also indistinguishable from that of the Monchique massif of S. Portugal (for example five K–Ar mineral ages give a mean age of 72 ± 2 Myr; Macintyre & Berger 1982). This age correspondence gives further substance to the proposition of Ribeiro (1979) that the Sintra, Sines and Monchique complexes developed along a common shear zone. Considering the evidence for repeated magmatic activity in these centres, a minor thermal event in the Sines complex in the Late Cretaceous or Early Tertiary, giving rise to the two *c.* 63 Ma K–Ar dates, cannot be excluded.

Properties of remanent magnetization

Magnetizations have been recognized from linear segments of orthogonal vector projections. Practically all of these magnetizations are those of highest stability (end-points) as consistent components of lower resistance have been successfully extracted from only six specimens. For about 25 per cent of the investigated material, terminal directions are associated with relatively low magnetic moments, and determination of their end-point magnetizations are generally obscured by experimental noise, viscosity effects etc. The problematic sites in this respect come primarily from the granitic intrusions (Sintra and Sines) and from the Lisbon basalts. These rocks have suffered a higher degree of secondary alteration (weathering) than the remaining part of the collection (see also below). However, stable remanence directions have been determined from nearly all investigated sites (most sampling locations are represented by many end-point results), and these data form the basis of the following discussion.

SINTRA COMPLEX

All stable end-point results, encompassing granite/syenite, gabbro/diorite, and a single diabase dyke, are depicted in Fig. 3(b). The total population is relatively well grouped but a certain spread towards the direction of the present axial dipole field is noted. Of the 14 sites investigated from this complex, two sites, 12 and 13 (granite and syenite), did not give sensible results, and directional data are therefore not available from these sampling locations. As can be seen from the site mean directions of Fig. 3(d) the granite and syenite rocks have on average steeper inclinations than the gabbro and diorite material. From field evidence there is also a fairly clear distinction between the two groups of rocks in that the gabbros and diorites have a very fresh appearance as compared with the granitic rocks which frequently show a high degree of secondary alteration. Sites 19 and 20 in particular, i.e. those with mean directions closest to that of the present axial geocentric dipole field, come from an area of the Sintra granite in which rock disintegration due to weathering is pronounced. The within-site distribution of stable end-point results for site 1 (syenite) forms a nearly perfect linear spread between the shallow end of the total specimen population (Fig. 3b), represented by gabbros and diorites, and the present dipole field direction (Fig. 3e). The magnetic mineralogy of site 1 is also dominated by haematite (cf. specimens Si 3-A2 and Si 5-A1 of Fig. 3f), suggesting that the rock has been subjected to strong oxidation.

The investigated dyke, site 10, has an overall stable magnetization located in the steeper half of the site mean population (Fig. 3d). Measurement of saturation magnetization, J_S, versus temperature, T, shows a marked 'kink' at around 150°C (cf. specimen Si 55-B, Fig. 3h). This feature was originally described by Ade-Hall, Palmer & Hubbard (1971) and related to either high deuteric or to advanced low temperature oxidation. The rapid decay of saturation magnetization at around 340°C, and the relatively low J_S on subsequent cooling (after heating to 600°C) are indicative of break-down of maghaemite. The suggested transformation of maghaemite is also supported by the marked intensity drop at around 350°C during thermal demagnetization (see Si 55-A2, Fig. 3f). This suggests that the dyke must have suffered extensive low temperature oxidation of its original iron–titanium mineralogy.

On the other hand, most of the gabbro and diorite material shows a much nearer virgin state of magnetic mineralogy, and $J_S - T$ measurements (cf. Si 49-C, Fig 3h) and thermal demagnetization (see Si 50-A2, Fig. 3f) suggest that the most important remanence-carrying phase is pure magnetite. Stable end-point directions are generally easily obtained after

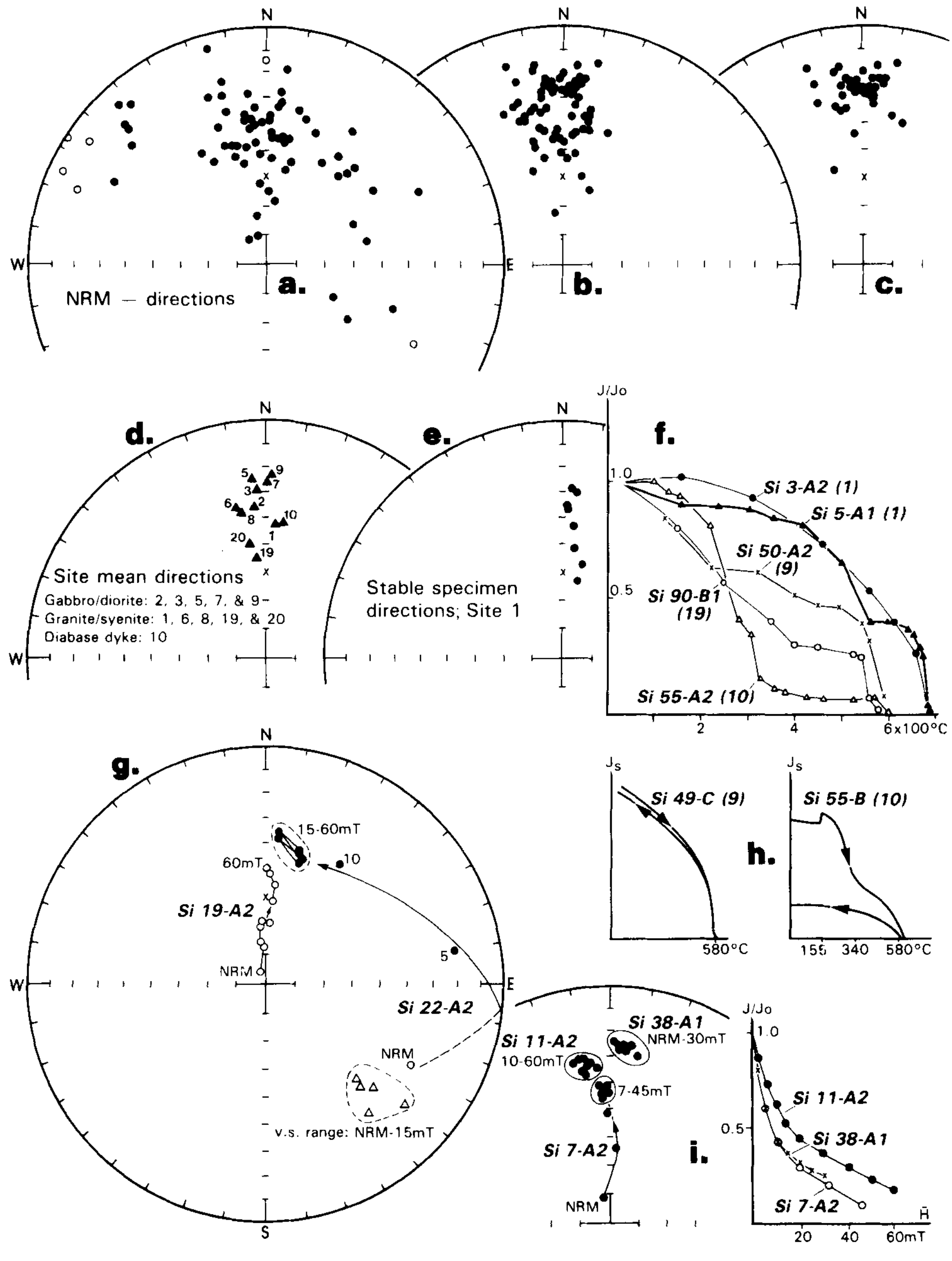

Figure 3. Laboratory results from the Sintra complex. The various diagram sections are as follows: (a) distribution of NRM directions of individual specimens; (b) all stable specimen directions; (c) stable specimen results from the gabbros and diorites; (d) site mean directions; (e) within-site distribution of stable remanence directions for site 1; (f) examples of intensity decay patterns on thermal demagnetization; (g) two examples illustrating the complex magnetization build-up for site 4; (h) examples of saturation magnetization versus temperature ($J_S - T$); (i) examples of direction and intensity variation during progressive alternating field demagnetization (field in milli-Tesla) of gabbro specimens. The elongated distribution for site 1 (diagram e) signifies the co-existence of characteristic gabbro/diorite and Upper Tertiary–Quaternary field components, i.e. most of the stable directions do not represent univectorial magnetizations. Note the clustered vector subtracted directions (triangles) for the low-stability magnetization of specimen Si22-A2, defining one of the components probably creating the markedly smeared 'NNW–SSE' distribution of NRM directions [see (a)]. This low-stability magnetization is thought to reside in maghaemite. The crosses of projections represent the direction of the present relative geocentric dipole field. Projection is equal area and closed (open) symbols represent directions in the lower (upper) hemisphere.

elimination of a minor low stability component (Fig. 3i). The only site with a complex remanence is site 4 (Fig. 3g). Most of its specimens have resistant upward-pointing magnetizations that move in a northerly direction upon progressive demagnetization (see Si 19-A2, Fig. 3g) but without attaining stable end-points at the highest available fields or before the onset of the erratic stage on thermal demagnetization. It is believed that this site has a dual-polarity magnetization with strong component overlap. Only one of the site 4 specimens (Si 22-A2) shows the characteristic behaviour, i.e. reaching the northerly directed stable end-point after removal of a low-stability magnetization. However, this is the only specimen from the Sintra and Sines collections for which the direction of a low-stability component has been established. Various vector-subtracted directions within the NRM–15 mT range define a clustered group (corresponding to linear segments of orthogonal vector plots), with SE declination and shallow upward inclination. This latter magnetization is probably in part responsible for the elongated distribution of the NRM data (Fig. 3a), corresponding to the inferred ancient secondary magnetization of the Lisbon lavas (see below). However, due to too much component overlap with the principal magnetization this low-stability remanence could not be adequately defined from the Sintra rocks.

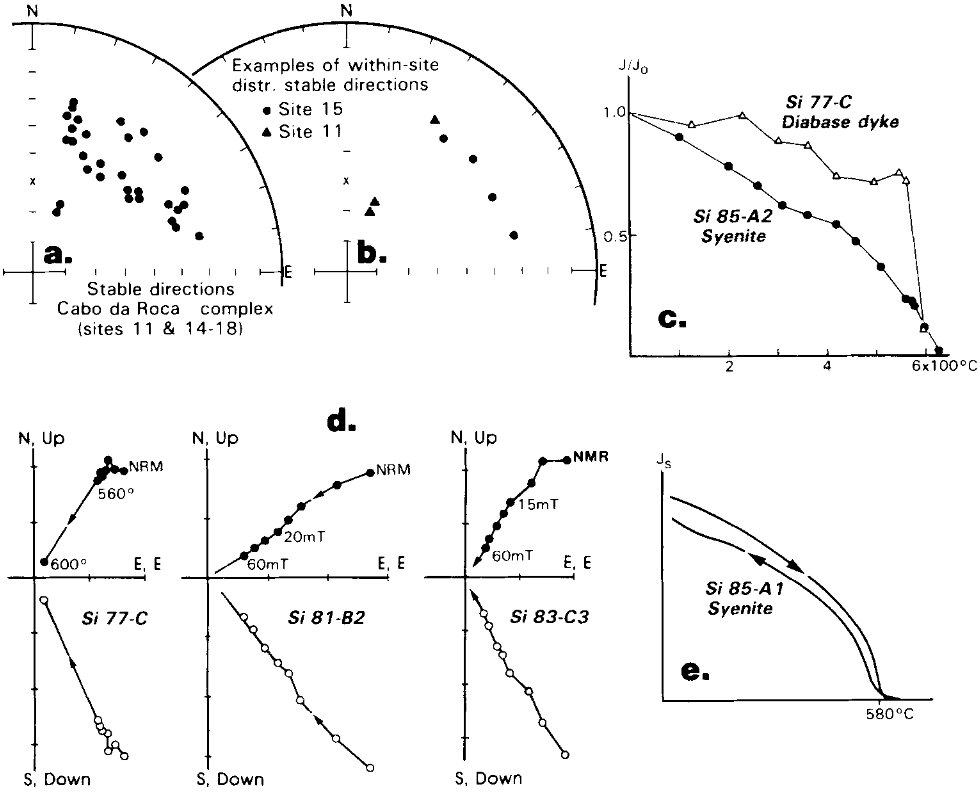

Figure 4. Experimental results from the Cabo da Roca complex. The diagram parts are as follows: (a) distribution of stable remanence directions; (b) examples of within-site stable magnetization; (c) intensity decay patterns on thermal demagnetization; (d) examples of orthogonal vector diagrams defining stable end-point directions; (e) the J_S-T behaviour (thermal cycling) for a syenite sample. Note the NNW–SSE stretching of the stable direction group, fig. a. See text for further details. In the vector plots (fig. d) projections in the vertical and horizontal planes are marked by open and closed symbols respectively. Projections and other conventions as for Fig. 3.

In conclusion, the primary magnetization of the Sintra complex is best represented by the gabbros and diorites. Subsequent low temperature oxidation has primarily affected the granitic rocks, and in particular Quaternary weathering has added a chemical magnetization component of high magnetic stability, aligned along the direction of the present axial dipole field.

CABO DA ROCA COMPLEX

This minor intrusive complex, injected into the Sintra pluton, has been studied at six sites, 11 and 14–18 (Fig. 1). Site 11 represents a relatively strongly weathered granite, and two of the three stable end-point results obtained are located close to the present dipole field direction (Fig. 4b). Many of the stable magnetization directions are well defined (see Fig. 4d) and pure magnetite appears to be the important remanence carrier (*cf.* Fig. 4c and e). The total distribution of stable end-point directions differ from that of the Sintra complex by showing more easterly declinations. However, the population is clearly smeared in declination (Fig. 4a) and the same feature is demonstrated in the within-site distribution of site 15 (Fig. 4b). This suggests that at least in part the stable remanence directions do not constitute single-component vectors. The following two schemes of partial remagnetization seem most appropriate;

(1) The remanence population consists of the original component of the Sintra complex (discussed above) plus a reversely magnetized component. If this is the case, all the stable remanence directions are unresolved composite magnetizations, and the results can hardly be used for geological correlation purposes, tectonic studies etc.;

(2) The majority of the stable directions are univectorial, having been imposed at the time of cooling, but superimposition of minor secondary remanences have led to a slight 'stretching' of the population in southeasterly and northwesterly directions. Comparison with data from the Sines complex (see below) suggests strongly that this is the appropriate explanation. This means that the overall direction of the Cabo da Roca complex closely approximates to that of the relative axial dipole field at the time of cooling (ca. 75 Ma).

SINES COMPLEX

The stable end-point directions for this formation constitute a well-defined circular grouping (Fig. 5b), the mean direction ($D_m = 041$, $I_m = 041$, $\alpha_{95} = 3.3^\circ$) being in nearly perfect agreement with that of the Cabo da Roca complex ($D_m = 039$, $I_m = 43$, $\alpha_{95} = 6.2^\circ$). High quality stable remanence directions have been obtained (cf. Fig. 5d), and nearly pure magnetite appears to be the most important remanence carrying mineral (cf. Fig. 5c–e). However, the presence of maghaemite is clearly in evidence, notably for syenite and diabase rocks. This mineral is identified by its thermal instability, mineral transformation frequently taking place at around 350°C, as suggested by $J_S - T$ measurements (see Fig. 5e, sample Sin 2). Another possible demonstration of the presence of a low temperature magnetic mineralogy is shown by the intensity decay pattern on thermal demagnetization in Fig. 5c. A steady intensity reduction below *c.* 350°C is superseded by a 'magnetite platform', and the 350°C $J_n - T$ discordance is associated with a 'sudden' reduction in low field magnetic susceptibility, K. These observations are most likely related to the break-down of a maghemite structure.

It is an important observation that the distribution of Sines NRM directions (Fig. 5a) is practically identical to the distribution of stable end-point data for the Cabo da Roca com-

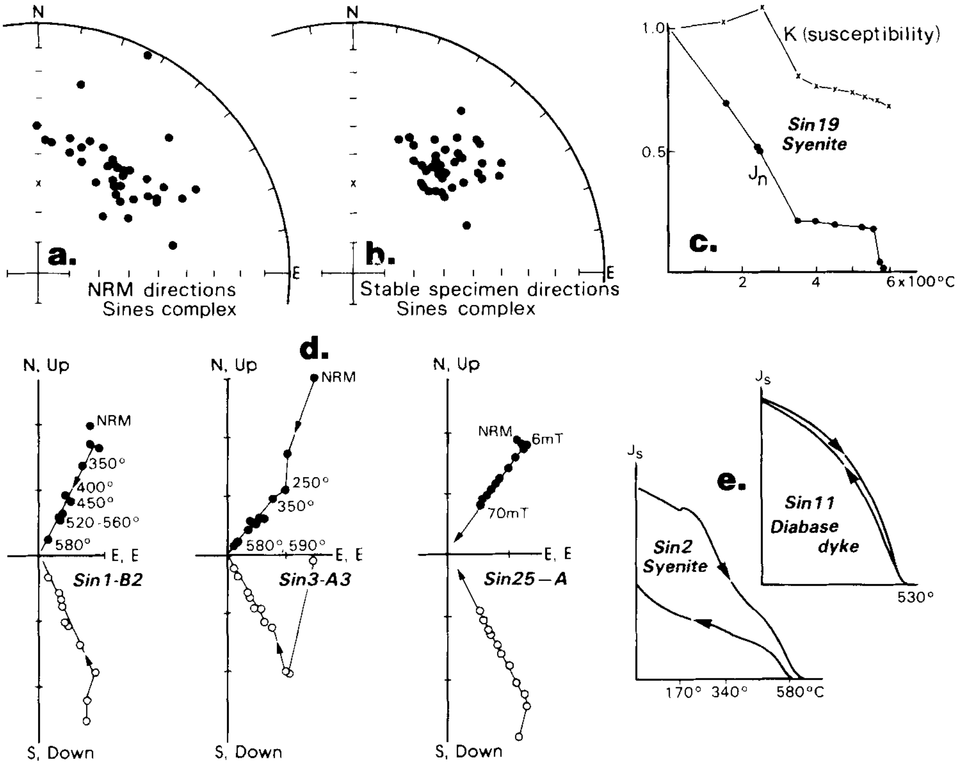

Figure 5. Results from the Sines complex. The diagram constitutes the following: (a) distribution of NRM directions; (b) stable end-point magnetizations; (c) remanence intensity and low-field susceptibility versus increasing temperature for sample Sin 19; (d) vector diagrams showing characteristic examples of magnetization stability; (e) examples of $J_S - T$ results. Note again the NNW–SSE smear of the NRM population, i.e. the low-stability magnetization is not aligned along the direction of the present geomagnetic field. Conventions etc. as for Figs 3 and 4.

plex (Fig. 4a). These elongated groupings, also recognized in the NRM population of the Sintra pluton (Fig. 3a), are regarded here as having been imposed during ancient partial remagnetization associated with maghaemitization (low temperature oxidation). For the Sines and Sintra complexes these NNW/SSE directed secondary components (see below) are easily removed on partial demagnetization, but these components have clearly not been entirely eliminated in some of the Cabo da Roca specimens. However, the stable but minor secondary overprinting of the latter formation is apparently symmetrical, i.e. the southeasterly ‘drag’ in the remanence population is about equal to the northwesterly one, so that the overall mean direction is not being noticeably affected by the unresolved secondary remanences.

LISBON VOLCANICS

The complex remanence build-up of this formation as suggested by previous analyses (Storetvedt 1970, 1973) has been reconfirmed by the present study. The magnetic mineralogy appears strongly dominated by maghaemite (see Fig. 6l and m and fig. 1 of Storetvedt 1973), suggesting that chemical magnetization of weathering origin provides a

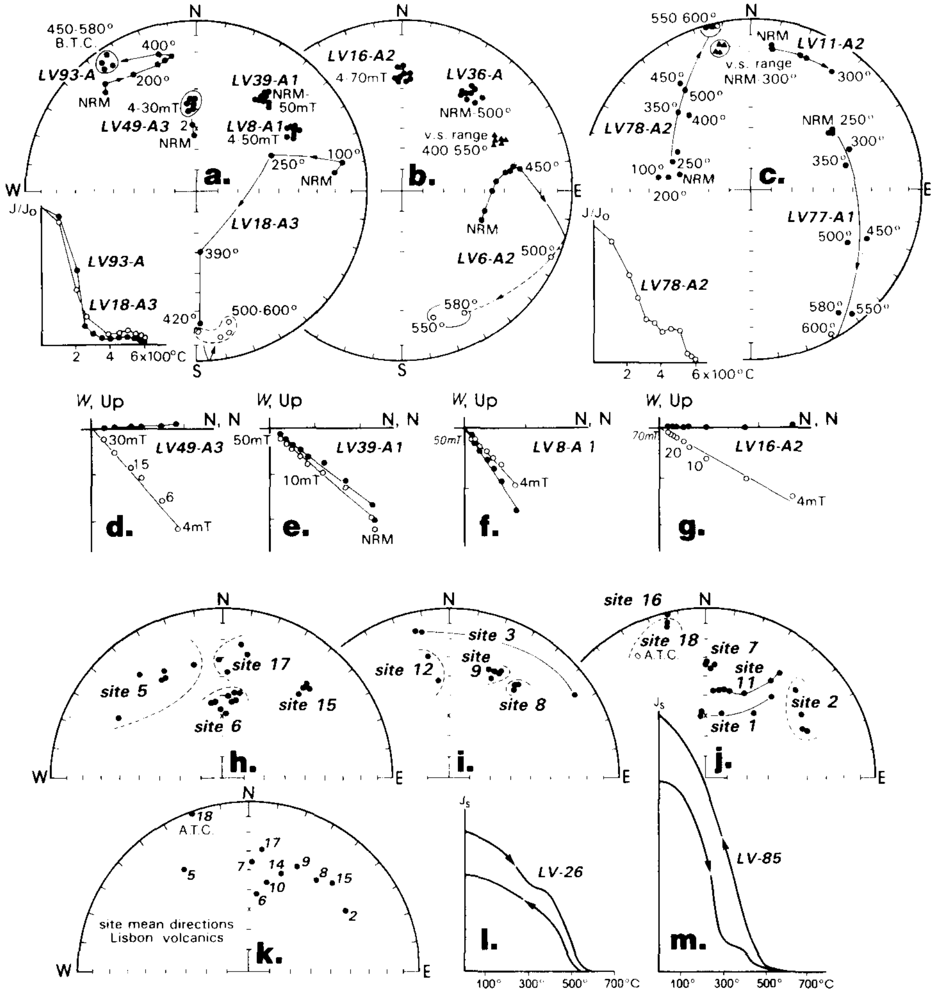

Figure 6. Composite diagram showing a variety of results from the Lisbon volcanics. The sub-figures are: (a–c) examples of intensity and directions of magnetization upon progressive demagnetization. Vector subtracted directions (triangles) over specified demagnetization ranges are shown for specimens LV6-A2 (b) and LV11-A2 (c); (d–g) orthogonal vector plots for highly stable specimens (from Figs a and b); (h–j) within-site grouping of remanence directions. Note the composite magnetization of sites 1 and 11 (Fig. j) and of site 3 (Fig. i); (k), site mean directions for all sites having a minimum of three 'clustered' specimen results; (l and m) examples of J_S–T results indicating the presence of maghaemite. A.T.C. and B.T.C. denote *after* and *before* tectonic correction (structural correction for the Lisbon volcanics results apply only for site 18, Figs a, j and k). Projections and conventions as for Figs 3 and 4.

substantial contribution to the present palaeomagnetic record. After elimination of minor low stability components several specimens exhibit well-defined stable end-points (cf. Fig. 6d–g). The majority of these stable directions of magnetization are located in the NE quadrant (Fig. 6h–j). These moderately inclined northeasterly magnetizations have not previously been reported from the Lisbon volcanics. The reason for this will be discussed below.

Another group of results define a fairly shallow NNW/SSE directed axis of magnetization. In some cases the remanence is characterized by stability throughout practically the full

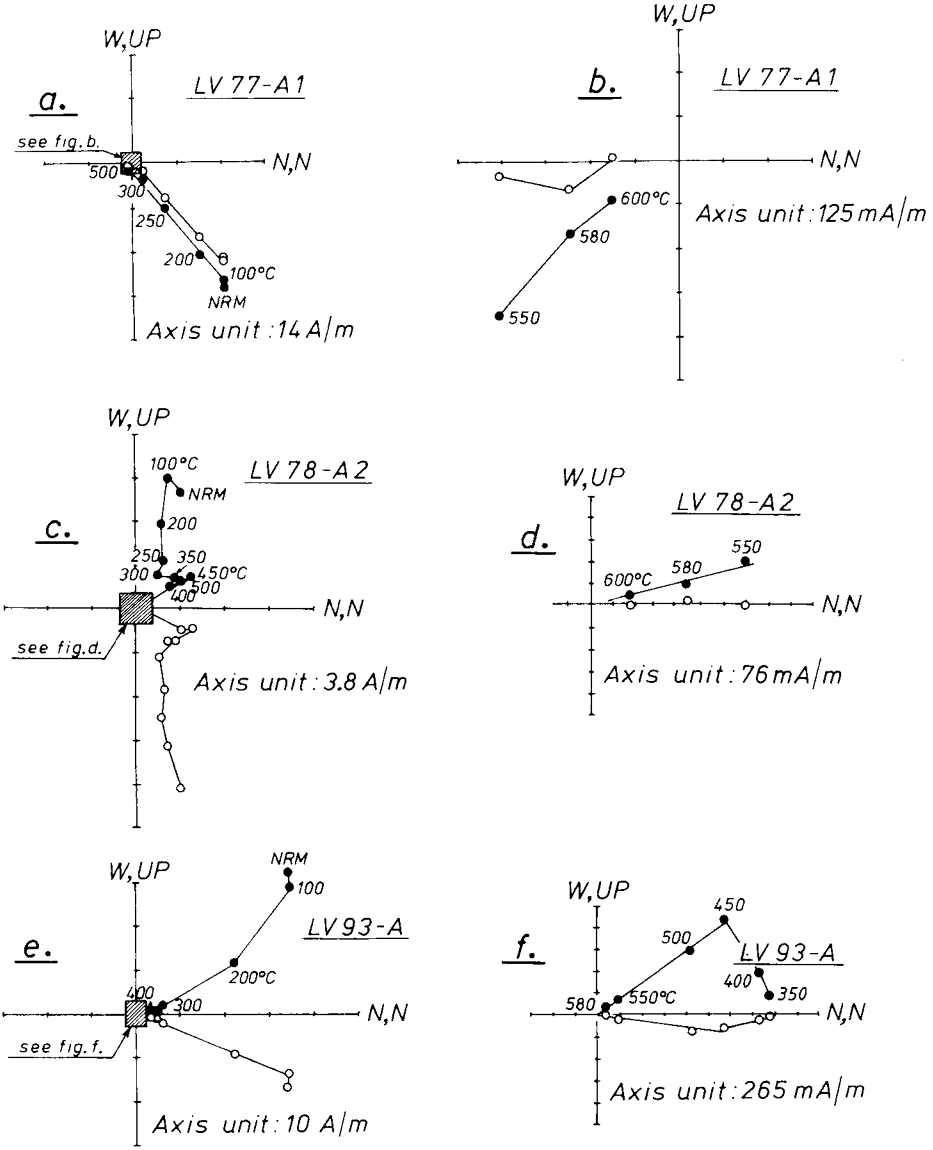

Figure 7. Orthogonal vector plots of three specimens demonstrating multicomponent magnetizations. Diagrams b, d and f display the remanence behaviour at high temperatures. Note that all specimens possess both the NE and the NNW/SSE magnetizations. Diagram convention as for Fig. 3.

range of demagnetization (both thermal and AF), for example all specimens of site 5 and two of the site 18 specimens, but in other cases the NNW/SSE directed remanences are determined either by high temperature end-points (cf. LV18-A3, Fig. 6a; LV6-A2, Fig. 6b; LV78-A2, Fig. 6c) or by the directions of erased vectors below *c.* 350°C (see for example the clustered directions of subtracted vectors for specimen LV11-A2, Fig. 6c). Low temperature oxidation and remagnetization appear to have affected the Lisbon volcanics throughout the history of the formation. The resulting palaeomagnetic diversity that may occur at specimen level is adequately illustrated by the following examples.

As shown in Figs 6(a) and 7(e) and (f) the shallow NNW directed magnetization of sample LV93-A is apparently carried by both maghemite (blocking temperatures (T_b) < 350°C) and pure magnetite. Like maghemite, magnetite may have formed at low temperatures too as a disintegration product of original titanomagnetite (Readman & O'Reilly 1970). A second palaeomagnetic component of this sample, corresponding to that of Sines and Cabo da Roca formations, is held by the intermediate T_b-range (350°–450°C). The same two-axis magnetization is also exhibited by specimens LV6-A2 (Fig. 6b) and LV77-A1 (Figs 6c and 7a and b). The latter sample suggests that the northeasterly remanence is carried either by maghemite or by a virgin state titanomagnetite, while magnetite or rather a thermally stable titanomaghemite (note that $T_b > 600°$C) carries the high temperature direction (shallow southeast). Sample LV78-A2 has a shallow NNW high temperature ($T > 550°$C) magnetization (Fig. 6a and 7d), but at intermediate temperatures (300°–500°C) a shallow reversed component is erased (see Fig. 7c). Thus, the fairly stable but deviating bulk magnetization of this sample (below 300°C) is regarded as resulting from superimposed normal and reverse components probably carried by maghemite. Such multicomponent magnetizations are fully compatible with the $J_S - T$ evidence for extensive low temperature oxidation of the original opaque mineralogy.

Most of the investigated locations show fairly good within-site consistency (cf. Fig. 6h–j) of estimated component directions, but sites 1 and 11 (Fig. 6j) show smeared distributions towards the present dipole field direction, suggesting the co-existence of (unresolved) Sines/Cabo da Roca and Upper Tertiary–Quaternary geomagnetic field components. Site 3 (Fig. 6i) appears to possess the Sines/Cabo da Roca and the shallow NNW directed magnetizations, i.e. in complete agreement with the magnetization buildup of individual specimens discussed above. A directional distribution like that of site 3 should not be difficult to understand in that the acquisition of chemical remanence must be a function of micro-environmental oxidizing conditions which in turn is time-dependent, i.e. small volumes of a rock body may suffer complete magnetic resetting in a secondary field while other parts of the rock may be hardly affected at all.

Fig. 6(k) depicts the distribution of site mean directions for the best grouped sites, including only those with at least three individual directions of magnetization and excluding locations for which the internal magnetizations is clearly composite (smeared). However, also the distribution of these critically selected site means show a clear evidence of smearing: (1) towards the NNW directions, and (2) towards the present axial dipole field direction. This probably implies that at least some of the apparent univectorial magnetizations of Fig. 6(d–g) in fact are multivectorial but lack the necessary component resolution. Thus, Fig. 6(k) and other evidence discussed above suggest that the Lisbon volcanics constitute a three-axis magnetization. These components are: (1) a NE magnetization, corresponding to that of the Sines/Cabo da Roca complexes; (2) a shallow NNW/SSE magnetization; and finally (3) a component of more recent origin, aligned along the Upper Tertiary–Quaternary dipole field direction.

The NE magnetization has not been recognized in previous studies of the Lisbon volcanics. The reason for this discrepancy is probably that earlier collections were principally obtained by hand sampling, while the present material has been collected by field drilling in carefully selected and freshest-looking material. This sampling of the most compact parts of the various outcrops seems to have provided samples with a stronger component of original magnetization than that achieved in earlier collections. However, this primary magnetization has not been cleanly separated from the secondary components present (cf. the smeared distribution of site mean directions of Fig. 6k).

With one exception (site 18) the invested sites have random but insignificant 'tectonic' dips, and structural corrections had therefore no real effect. Site 18 has a stratal dip of 58°, but the characteristic NNW directed magnetization for this location undergoes only a minor adjustment through the tectonic 'unfolding'. From the present collection it is impossible, therefore, to decide whether the shallow NNW/SSE directed magnetization of assumed secondary origin pre- or post-dates the limited local tectonic disturbance. Even if one includes the few relevant data from a previous study (Storetvedt 1973), the fold test becomes insignificant. However, due to the fairly recent age of the structural deformation (Miocene and younger) it seems most appropriate that the shallow NNW/SSE magnetization, which is significantly different from the Upper Tertiary field, should be subjected to tectonic adjustment wherever such corrections can be applied.

The NNW/SSE magnetization is clearly present in all formations studied here but it is only in the Lisbon volcanics that this remanence component can be numerically estimated. Combining the data from the present study with the results of an earlier investigation (Storetvedt 1973) [totally 23 individual specimen results (Fig. 8)] give an overall magnetization (reversed directions inverted) of D = 333, I = 14 (α_{95} = 7.4). The NNW magnetization is clearly separated from the remaining remanence population.

Palaeomagnetic data from both Europe and Africa suggest a marked polar shift in Caenozoic time, the present relative dipole axis probably being established by Middle Tertiary time (Storetvedt 1978, 1980). The significant climatic deterioration as revealed for Europe at around 30 Ma (Lower Oligocene) may be directly linked to this polar shift. Hence, even the youngest palaeomagnetic component of the Lisbon volcanics (i.e. that directed along or close to the direction of the present axial dipole field) may at least in part pre-date the local tectonic deformation. Therefore, the late Tertiary magnetization should in principle respond positively to structural unfolding provided the pretectonic component constitutes more than 50 per cent of the 'stable remanence'. This is apparently the case in the study by Van der Voo & Zijderveld (1971). These authors only present two examples of their thermal demagnetization data, but one of these specimens (VIM 195) shows that haematite is an important remanence carrier and the second specimen too has blocking temperatures well above 600°C. These results do indeed signify an advanced stage of low temperature (weathering) oxidation. It is not difficult to understand therefore why the palaeomagnetic record provided by Van der Voo & Zijderveld represent predominantly late Tertiary secondary magnetization (cf. Storetvedt 1973).

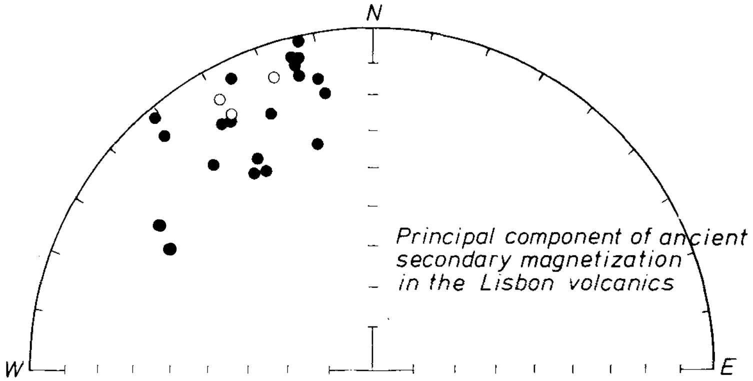

Figure 8. Principal component of secondary magnetization as unravelled from the Lisbon volcanics, including nine specimen directions from a previous study (Storetvedt 1973). Directions are after structural correction. This magnetization which fits the late Cretaceous palaeomagnetic field for Africa, was probably imposed after the tectonic rotations of Iberia had ceased.

The palaeomagnetic record; geological correlation and plate tectonic considerations

The experimental data outlined above suggest that the Sintra and Sines complexes and the Lisbon volcanics have recorded magnetizations along altogether four different relative field axes, covering an age span from Turonian (*c.* 90 Myr) to Quaternary. The overall directions of the three principal magnetizations, including statistical parameters and pole locations, are listed in Table 3.

The (inferred) oldest palaeomagnetic direction is that represented by the gabbro and diorite sites of the Sintra complex (axis I of Fig. 9). The granitic rocks of this complex show a much more advanced low temperature alteration (weathering) than the basic–intermediate rock units, and a stable chemical remanent magnetization has apparently been added along the direction of the present geomagnetic dipole field. Hence, the granite and syenite material has steeper inclinations than the gabbro and diorite rocks owing to unresolved dual-component magnetizations. In an earlier study of the Sintra complex (Van der Voo 1969) seven of the eight sampling sites were granitic rocks. The only gabbro investigated gave consistently shallower inclinations than those characterizing the granites, i.e. in complete agreement with the results of the present study. We conclude that in view of the available evidence for at least partial chemical remagnetization of the Sintra granitic rocks the mean direction given by Van der Voo (1969), $D = 359^\circ$, $I = 43^\circ$ ($\alpha_{95} = 5^\circ$) is now to be superseded by the mean gabbro/diorite magnetization of the present study, $D = 359^\circ$, $I = 27.3^\circ$ ($\alpha_{95} = 3.3^\circ$). It is believed that the latter magnetization represents the true relative palaeomagnetic field at the time the Sintra pluton cooled some 90 Ma.

The second oldest magnetization (axis II of Fig. 9) is that characterizing the Sines and Cabo da Roca complexes, $D = 041^\circ$, $I = 41^\circ$. This palaeomagnetic direction is also present in the Lisbon volcanics though it is here somewhat blurred due to interaction with unresolved secondary remanences. This northeasterly component appears to be an extremely important one for evaluating the tectonic history of Iberia (see below). The consistent palaeomagnetic and radiometric results from the Sines and Cabo da Roca complexes, despite a highly variable petrography, suggest that the fossil magnetization dates from the time of their origin at around 75 Ma (Senonian).

The third magnetization (axis III of Fig. 9) is considered as secondary in origin and related to ancient low temperature oxidation processes (maghemitization). This shallow

Table 3. Summary of palaeomagnetic results.

Formation	Magnetic age (Myr)	*N*	D_m	I_m	α_{95}	*R*	*K*	Pole	dp/dm
Sintra a	~ 90	34	358.0	27.3	3.3	33.4	57	65N,176E	2.0/3.6
Sintra b	~ 90	5	357.7	27.1	5.7	5.0	180	65N,176E	3.4/6.2
Sintra c	~ 90	70	357.6	35.0	2.9	68.0	35	70N,178E	1.9/3.4
Sintra d	~ 90	11	357.0	36.2	6.7	10.8	48	71N,180E	4.5/7.7
Cabo da Roca a	~ 75	30	039.4	43.4	6.2	28.5	19	54N,092E	4.8/7.7
Cabo da Roca b	~ 75	6	041.0	44.9	13.4	5.8	26	54N,089E	10.7/17.0
Sines a	~ 75	36	040.8	41.4	3.3	35.4	54	53N,092E	2.4/4.0
Sines b	~ 75	9	041.4	41.9	5.4	8.9	92	52N,091E	4.0/6.6
Lisbon volc. a	Overprint	23	333.4	13.5	7.4	21.8	18	50N,215E	3.9/7.6

Groups a and b are with unit weight on specimens and sites respectively. Sintra a and b represent gabbro and diorite material only, while groups c and d include also the granite and syenite rocks which apparently have suffered partial chemical remagnetization (weathering) in Quaternary time. The Lisbon volcanics overprint is most likely of late Cretaceous age. The symbols are: *N* unit vectors (specimens or sites); D_m, mean declination; I_m, mean inclination, α_{95}, half-angle of the cone of 95 per cent confidence about the mean (in degrees); *R*, length of resultant vector; *K*, precision parameter; dp, dm, the semi-axes of the oval of 95 per cent confidence about the mean pole. Statistics according to Fisher (1953).

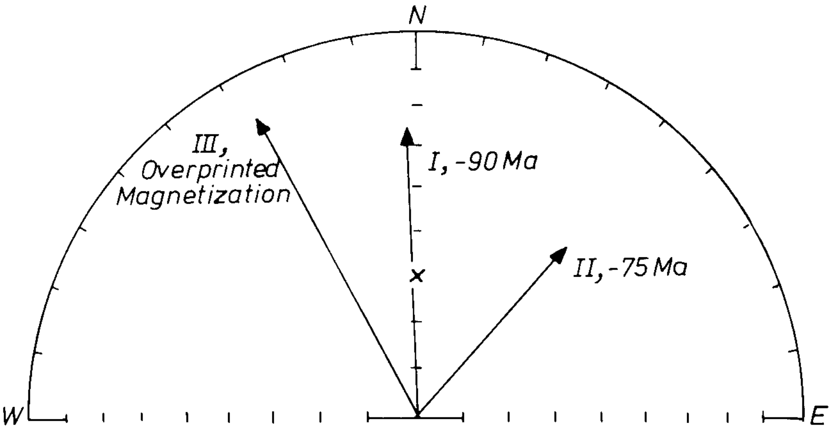

Figure 9. Upper Cretaceous palaeomagnetic axes relative to Iberia as determined by the present investigation: axis I, Sintra complex; axis II, Sines and Cabo da Roca complexes (this magnetization is also present in the Lisbon volcanics); axis III, major overprinted magnetization.

NNW–SSE directed remanence which has a dual-polarity build-up can only be determined numerically from the deeply weathered Lisbon volcanics, yet the presence of this component is also recognized by a 'symmetrical' NNW–SSE smear in the NRM populations of the intrusive complexes. In the Sintra and Sines rocks this overprint is easily removed, while in the Cabo da Roca complex the secondary magnetization has in part overlapping stability with the characteristic magnetization, providing certain NNW–SSE elongation of the stable remanence grouping. Axis III is less well defined than the two other palaeomagnetic field directions, but the results of the present study confirm those of an earlier investigation (Storetvedt 1973) based on entirely different measuring and demagnetization equipment. The combined results of the two studies give after structural correction an overall direction of $D = 333°$, $I = 14°$ ($\alpha_{95} = 7.4°$). This magnetization has a shallower inclination than the other two axes of magnetization. Both European and African palaeomagnetic data show a certain shallowing of the palaeomagnetic field at the end of the Cretaceous or early Tertiary (see Storetvedt 1978, 1980). It is suggested therefore that the axis III magnetization most likely dates from around the Cretaceous–Tertiary boundary (~ 65 Ma), but a lower Tertiary (pre-Oligocene) age cannot be ruled out at this stage. Storetvedt (1973) regarded the shallow NNW direction to be the one closest in age to that of the original magnetization, but the present evidence for at least an axis II magnetization in the Lisbon volcanics (though not cleanly separated) renders the previous conclusion invalid. The hand sampling of earlier investigations inevitably encompassed the more fractured and hence the more weathered sections of the individual flows, the secondary processes having apparently erased the original magnetization beyond recognition. From the available palaeomagnetic and radiometric age evidence (Ferreira & Macedo 1979 report a mean of age of 72.6 ± 3.5 Myr from five whole rock K–Ar results) it appears reasonable to conclude that the Lisbon volcanics developed concurrently with the Sines and Cabo da Roca intrusive activity (axis II). Hence, the coexistence of axes II and III magnetizations in individual specimens is further evidence in favour of a secondary origin for axis III.

The late Cretaceous palaeomagnetic reference frames for Africa and notably Europe are not sufficiently precisely defined to justify any dynamic evaluation of the *c.* 15° inclination difference between the axes I and II magnetizations, but the declination discrepancy is most likely related to the tectonic history of Iberia. Whatever reference frame is being referred to

(Europe or Africa) a major counterclockwise rotation of Iberia followed by a clockwise one must be invoked. Both rotations are likely to be of Upper Cretaceous age, predating the overprinted axis III magnetization (see below). This tectonic instability of Iberia may have resulted from plate boundary irregularities during a phase of Europe/Africa plate convergence. The Upper Cretaceous probably provides the most important time span for post-Palaeozoic seafloor spreading and global tectonics (Storetvedt 1985), representing: (1) the most significant phase of coastal onlap in Phanerozoic time; (2) major sedimentological and tectonic discordances in the oceanic basins; (3) formation of major intracontinental sedimentary basins, etc. From a radiometric study of dredged gabbros from the Gorringe Ridge (just to the southwest of Iberia) Prichard & Mitchell (1979) obtained (in addition to other results) a mean age of 82 ± 3 Myr for three deformed plagioclases. This age was interpreted as related to a phase of shearing, an explanation that would be fully compatible with the inferred Upper Cretateous rotation of Iberia.

The axis III pole, which is in close correspondence with the Upper Cretaceous apparent polar wander path (APW) for Africa, may have been impressed after the two-phase tectonic rotation of Iberia had ceased. Also older Mesozoic and Permian poles for Iberia have a closer match with the African APW than with the European one. This raises the question of whether Africa is the reference plate of relevance here (before and after the independent two-phase rotation of Iberia), i.e. whether the true Africa/Europe plate boundary in the region is approximated by the North Pyrenean Fracture Zone. If so, the clockwise rotation would be about 20° larger than the preceding counterclockwise movement, hence accounting for the compressive tectonic deformation in the Bay of Biscay. It is indeed pertinent to ask whether the Bay actually developed by stretching and continental attenuation processes, possibly reflected in the accelerated subsidence in the Aquitanian basin during Triassic/early Jurassic and late Jurassic/early Cretaceous (Brunet 1984), rather than being the result of Iberia having rotated away from adjacent Europe. For example, the matching of Hercynian trends from Brittany to Iberia certainly does not require a closing of the Bay, in that the present configuration may define a Variscan arc equally well. However, the dynamic and structural aspects of the plate tectonic model alluded to here will be dealt with in a later paper.

Acknowledgments

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