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Letter Section

A 21 ± 2 Ma age for the termination of the ductile Alpine deformation in the internal zone of the Betic Cordilleras, South Spain

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Abstract

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Rb–Sr dating of WR-muscovite pairs from two mica schists and two gneisses from the Vélez-Málaga–Torrox area, 40 km E of Málaga, gives tie lines indicating ages of 23.4 ± 2.7, 19.3 ± 2.2, 19.5 ± 0.7 and 22.4 ± 0.7 Ma, respectively. Geological evaluation suggests that these analytical ages indicate an age of 21 ± 2 (2σ) Ma for the metamorphic culmination connected with the latest phase of ductile deformation in the area. This Early Miocene (Aquitanian) age compares well with published radiometric ages for major orogenic processes in the westernmost Mediterranean and it is suggested that a significant part of the Alpine orogeny in the region took place in the restricted period of 19–23 Ma ago. Uplift rates in the order of 3–5 km/Ma are tentatively suggested.

Introduction

As part of the Alpine Belt, the Betic Cordilleras in South Spain and its counterpart in North Africa were formed in the zone of interaction between the African and the European plates (for a recent review, see Fontboté and Vera, 1986). Andrieux et al. (1971) introduced the term Alborán sub-plate for the western termination of this zone in the western Mediterranean, and Hsü (1977), extending its size and naming it "Alborán microcontinent", suggested that towards the south and north it was delimited by major wrench faults (cf. Weijermars, 1987). The geological history of this crustal domain is different from that of the Iberian Massif and main Africa; its stratigraphic/tectonic/metamorphic development, and thus the orogenic evolution of the Betic Cordilleras, is currently under discussion.

Two striking features of the Alpine Belt in the western Mediterranean are: (1) the continuation of major tectonic-stratigraphic elements of the Betic Cordilleras through Gibraltar into North Africa, a 180° loop, and (2) the variation in direction of apparent dominant nappe transport for the higher Betic nappe complexes: in South Spain from the south, in North Africa from the north. Not surprisingly this has called forward orogenic models based on mantle diapirism in the area of the Alborán Sea, combined with gravity

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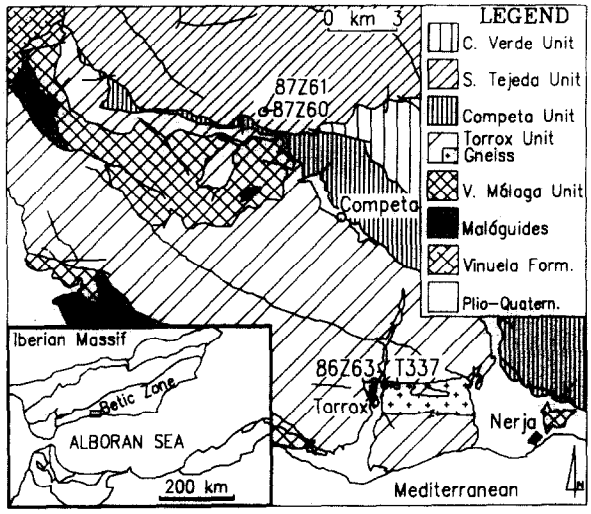

Fig. 1. Geological sketch map of the Vélez-Málaga–Torrox area showing sample locations.

tectonics, nappe shedding towards the north, west and south (Van Bemmelen, 1933; Loomis, 1975; Torres-Roldán, 1979). Geophysical data appear to support such a model (Bonini et al., 1973; Banda et al., 1983; Neugebauer, 1987). However, models involving continental lithosphere thickening by plate convergence and subsequent large-scale, low-angle extensional faulting have also been suggested (Platt, 1986; Dewey, 1988).

The present paper reports Rb–Sr isotopic data which suggest an age for the latest phase of ductile deformation in the Vélez-Málaga–Torrox area in the western-central part of the Betic Chain (Fig. 1). The area consists mainly of tectono-stratigraphic units which belong to the Alpujárride nappe complex. This complex forms the middle part of the Alpine nappe pile in the Betic zone, overlain by the Maláguide Complex and underlain by the Nevado-Filábride Complex.

The geochronological method chosen dates the closure age of the WR-muscovite system. As the closure temperature of muscovite is high (500° ± 50°C), this age should be close to that of the metamorphic culmination.

Geological setting and petrography

The Alpujárride tectono-stratigraphic units in the Vélez-Málaga–Torrox area (Fig. 1) have the following basic stratigraphy: (1) pre-Permo-Triassic graphite-rich mica schists and quartzites, (2) Permo-Triassic (Delgado et al., 1981) graphite-poor mica schists with intercalations of calc-silicate rocks and amphibolites, (3) Triassic carbonate rocks. Near the base of the graphite-rich mica schists–quartzite sequence of the Torrox unit a c. 2–4 km large gneiss body occurs (Fig. 1) which locally contains centimetre-size, subhedral K-feldspar crystals and in all samples shows microscopic fabrics reflecting ductile shearing with variable degrees of recrystallization.

In the studied area the tectonic units are sealed by the transgressive Viñuela Formation (Boulin et al., 1973; Fig. 1), at the base of which planktonic Foraminifera have been found (González-Donoso et al., 1982) which belong to biozone N5 of Blow (1969), for which an age interval of 19–22.5 Ma has been suggested (Harland et al., 1982).

Location of the four analyzed samples is indicated in Fig. 1. Two of them (86Z60 and 86Z61) are Permo-Triassic mica schists from the Sierra Tejeda Unit (coordinates 4°01′05″W, 36°52′01″N). The other two (T337 and 86Z63) are gneisses from the Torrox Unit (coordinates: 3°56′05″W, 36°46′01″N).

T337 — Biotite–muscovite augengneiss. Quartz, K-feldspar and plagioclase are the main components, forming 0.5–1 cm large crystals and aggregates preferentially oriented parallel to the foliation; in between, thin muscovite- and biotite-rich schlieren occur. *K-feldspar* crystals essentially represent magmatic crystals rotated into the foliation plane. *Plagioclase* crystals have retained less of their magmatic form; many crystals have been marginally replaced by fine-grained polyhedral aggregates of later plagioclase, quartz and muscovite. *Quartz* forms fine-grained, polycrystalline aggregates often schlieric in form, formed by ductile deformation and recrystallization of larger magmatic crystals. The *biotite-* and *muscovite*-rich schlieren (clots) consist of small, (0.1–0.3 mm) undeformed crystals; larger, relict crystals of muscovite showing some deformation are very rare. Muscovite also occurs as very small subhedral inclusions in plagioclase, and it fills thin cracks in feldspar crystals. It is concluded that the muscovite crystals dated in this survey (125–200 μm fraction) are post-magmatic and formed in direct response to the shearing of the granitic rock: mainly by strain-induced recrystallization of earlier mica crystals.

86Z63 — Muscovite gneiss, better foliated than T337, reflecting higher strain. Up to 1 cm large *K-feldspar* crystals are still present, in most cases within leucocratic schlieren. The major part of the *muscovite* is in 0.4–1 mm large subhedral crystals,

preferentially oriented parallel to the foliation and forming c. 20 vol.% of the rock. Many crystals show signs of minor deformation indicating that they witnessed at least the final stage of shearing. These crystals are set in a schistose matrix consisting of smaller crystals of quartz, plagioclase, biotite and muscovite; smaller biotite and muscovite crystals in part replaced and grew on the larger muscovite crystals. Small, anhedral *garnet* crystals are rare, and are partly replaced by late, yellow-brown biotite (and very rare chlorite), muscovite and quartz—the only obvious retrogressive effect in the rock. Quartz fabrics show a higher degree of recrystallization than in T337, but the ductile flow heritage is still apparent. One thin, c. 3 mm large schlieren is present in the section studied, consisting of biotite, andalusite and sillimanite, without any apparent reaction relation between these minerals or with the surrounding matrix; it might represent an Al-rich, restitic inclusion. The occurrence of some small crystals of pink-colourless pleochroic dumortierite replacing biotite may be mentioned as a curiosity.

86Z60 — Kyanite mica-schist, showing polygenetic structures, interference of several, small-folded S-planes. The main paragenesis of the rock consists of *muscovite, biotite, plagioclase* (An_{40-45}), *quartz* and *magnetite* in 0.2–1.5 mm large crystals which post-date these structures, overprinting them in mimetic fashion. Kyanite might predate this paragenesis which may be classified in the middle amphibolite facies. Within the nodes of the interferring S-planes, relict crystals of *plagioclase* and *garnet* occur. These crystals are marginally replaced by crystals of the main paragenesis and they are characterized by the abundance of small, partly helicitic inclusions. These crystals are suggested to belong to an older paragenesis. *Fibrolite* occurs in thin relict wisps and schlieren, usually related to biotite crystals; its paragenetic relations are uncertain. Traces of a late, retrogressive paragenesis are present and consist of very small crystals of *chlorite* and *epidote*. The muscovite

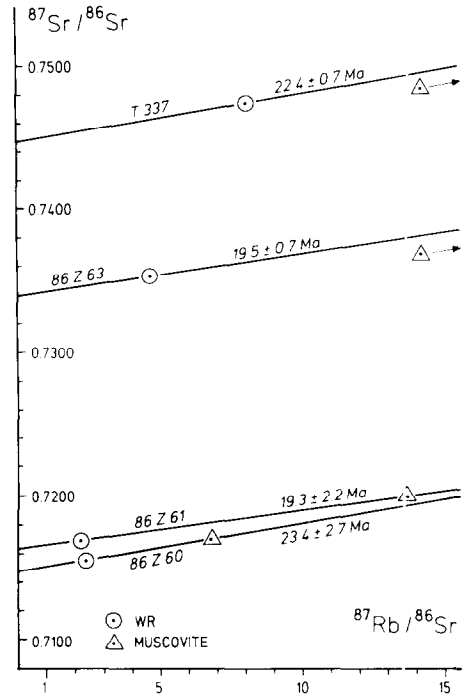

Fig. 2. Nicolaysen diagram showing the dating results for the four samples from the Vélez-Málaga–Torrox area. The four WR (whole rock)–muscovite tie lines are close to parallel. The slope of the lines dates the closure time of the muscovite Rb–Sr isotopic systems.

crystals dated (125–200 μm fraction) were probably formed directly after the latest folding, mainly by strain-induced recrystallization of pre-existing mica crystals.

86Z61 — Kyanite-bearing mica schist, similar to 86Z60, but the retrogressive paragenesis is more extensive. Almost all kyanite

TABLE 1

Rb–Sr analytical data

Sample No.	Type of sample [a]	Rb (ppm)	Total Sr (ppm)	^{86}Sr (μ mol/g)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$ [b]
86Z60	WR	231	279	0.314	2.394	0.71555 ± 2
	mu	345	148	0.166	6.767	0.71700 ± 3
86Z61	WR	205	270	0.304	2.200	0.71687 ± 3
	mu	388	82.2	0.093	13.652	0.7200 ± 1
86Z63	WR	222	138	0.156	4.649	0.73517 ± 5
	mu	539	21.5	0.024	72.95	0.75406 ± 7
T337	WR	266	95.8	0.107	8.073	0.74730 ± 6
	mu	523	12.4	0.014	123.02	0.78391 ± 5

[a] WR = whole rock, mu = muscovite.

[b] $^{87}Sr/^{86}Sr$ normalized to $^{86}Sr/^{88}Sr$ of 0.1194; $\pm 2\sigma$ error in last digit.

was replaced by aggregates of very fine muscovite (which was not included in the dated material).

Sample preparation, analytical procedures and results

A detailed description of sample preparation and analytical procedures is given in Zeck and Hansen (1988). Before crushing the samples were cleaned in an ultrasonic bath. Muscovite was separated in the 125–200 μm size fraction. The separates only have a few tenths of a per cent impurities (small inclusions of biotite, tourmaline and opaque material). The analytical results are given in Table 1 and Fig. 2. The least squares method of York (1969) was used for the calculation of the regression lines. The error on $^{87}Rb/^{86}Sr$ is 1%, and on $^{87}Sr/^{86}Sr$ it is 2σ of the run mean (Table 1). Errors given on the ages are at the 2σ level.

Discussion and conclusions; regional geological implications

The four tie lines in Fig. 2 are virtually parallel. Weighing the four ages according to the reciprocal square of their sigma error results in an average of 21.0 ± 0.8 Ma which represents the closure age of the WR (whole rock)–muscovite systems. Dodson and McClelland-Brown (1985) suggested $500 \pm 50\,^{\circ}C$ for the closure temperature of muscovite Rb–Sr isotopic systems at a cooling rate of $30\,^{\circ}C/Ma$.

The Rb–Sr isotopic muscovite systems, which are the subject of the present study, relate to the latest ductile deformation which is suggested to be connected with the nappe emplacement in the area studied. The metamorphic development since has been concordant for the various rocks within the tectonic units in the area. The presence of sillimanite and andalusite without reaction relations in one of the gneiss samples suggests conditions close to the sillimanite–andalusite field boundary. There is disagreement on the paragenetical position of kyanite in the mica schists; in case it belongs to the muscovite-bearing paragenesis, conditions near the Al_2SiO_5 triple point (c. 4 kbar/500 °C; Holdaway, 1971) would be indicated. A pressure range of 2–4 kbar in a temperature interval of 500–600 °C is suggested, indicating that the WR–muscovite systems should have closed for Rb–Sr diffusion simultaneously with or shortly after the formation of the muscovite in the mica schists and gneiss sample T337. In the gneiss sample 86Z63 the dated muscovite crystals predate the latest ductile deformation, but their age indicates that the Rb–Sr isotopic system has been reset.

A further estimate of the possible time gap between the formation of the muscovite in the mica schists and gneiss T337 and its isotopic closure may be based on the 19–22.5 Ma paleontological date for the basal part of the transgressive Viñuela Formation (above). Considering the 21 ± 1 Ma muscovite closure age a minimum cooling rate of c. 200 °C/Ma would be suggested, and consequently the time gap would be very small—the more so because such high cooling rates would increase the muscovite closure temperature (Dodson and McClelland-Brown, 1985). The minimum average uplift rate would be 2.5–5 km/Ma.

These minimum estimates for cooling and uplift rates are rather high, and they should be considered with caution as the uncertainties on the input data are considerable. The figures, however, compare fairly well with estimates of 3–5 km/Ma for early stages of uplift in the Eastern Alps (Cliff et al., 1985; von Blankenburg et al., 1989); the high cooling rate suggested by our data would be a result of the higher geothermal gradient in the Betic system. In conclusion: taking into account that the geological considerations add to the analytical error, it is suggested that the best estimate for the age of the metamorphic culmination is 21 ± 2 Ma.

The 21 ± 2 Ma age is statistically indistinguishable from the 22 ± 4 Ma Rb–Sr WR intrusion age (Priem et al., 1979) for a leucogranitic body at Penas Blancas within the marginal parts of the Ronda peridotite, Sierra Bermeja, c. 120 km west of Torrox. The leucogranitic magma there is thought to have formed by contact anatexis adjacent to mantle diapir material intruded into relatively high levels of the crust in the Alborán Sea area (Loomis, 1972, 1975; Torres-Roldán, 1981). Subsequent lateral displacement has delivered a

weakly S-dipping Alpujárride thrust slab of which the present Ronda peridotite body forms the lower part (Lundeen, 1978; Tubía and Cuevas, 1986: Los Reales nappe). We suggest that the 21 ± 2 Ma ductile deformation in the Alpujárride domain in the Vélez-Málaga–Torrox area can be correlated with the peridotite slab emplacement in the Ronda area, and is thus geologically (slightly) younger than the 22 ± 4 Ma age for the leucogranitic magma back-veining into the Ronda peridotite. This interpretation is somewhat different from that of Priem et al. (1979) who correlated the leucogranitic magma production with the lateral thrusting itself.

A direct age indication for the emplacement of the Los Reales peridotite nappe may be derived from WR–biotite ages, both Rb–Sr (3 measurements) and K–Ar (5 measurements), by Priem et al. (1979) on mylonitic rocks underlying the peridotite thrust sheet in the Sierra Alpujata, about 25 km east of the Ronda peridotite proper in the Sierra Bermeja. These ages are concordant, with an average age of 19.2 ± 0.3 Ma. Considering biotite closure temperatures of c. 300 °C (Jäger et al., 1967, see also Dodson and McClelland-Brown, 1985), a cooling rate of c. 200 °C/Ma (above) and metamorphic temperatures of 725 °C (Westerhof, 1977), this analytical age suggests a metamorphic culmination of c. 21 Ma, which would support our proposed correlation. Stratigraphic–tectonic relations in this eastern area also support the correlation. The transgressive San Pedro de Alcántara Formation in the Ronda–Marbella area which seals the trust plane between the Alpujárride and the Maláguide nappe complexes, yields nannoplankton floras (Aguado et al., in press) which belong to zone NN2 of Martini (1971) which corresponds to the later part of zone N5 of Blow (1969) with an inferred age interval of 19–21 Ma (Harland et al., 1982).

It is concluded that radiometric ages from the Alpine Belt in the western Mediterranean indicate that major orogenic processes such as intrusion of mantle material and nappe emplacement over a large area took place in a restricted period of time, probably between 19 and 23 Ma ago. A 19–20 Ma age for transgressive, nappe-sealing sediments suggests uplift rates in the order of 3–5 km/Ma. A picture arises of a contracted orogenic evolution, much quicker paced than currently evisaged.

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