Apatite fission track analysis in the Argentera massif: evidence of contrasting denudation rates in the External Crystalline Massifs of the Western Alps

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ABSTRACT

Apatite fission track dating from a central transect in the Argentera massif (southernmost External Crystalline Massif = ECM) yielded ages between 8.05 ± 0.6 and 2.4 ± 0.2 Myr, with a positive age/altitude correlation above 3 Ma, 1200 m. Recognising a thermal peak at c. 250 °C, 33 Ma, based on stratigraphic, metamorphic and 40Ar/39Ar data, the present results suggest a slow cooling rate (8–5 °C) for the Argentera massif during the Oligocene–early Pliocene. This rate compares with that from the Pelvoux massifs, but contrasts with those observed in the northern ECM (Monte-Blanc and Aar: up to 14 °C Myr⁻¹) for the same time interval. This can be related to the different location of the ECM within the collided European margin. At about 3–4 Ma, the denudation rate would have increased up to c. 1 mm yr⁻¹ in the Argentera massif, reaching the same value as in the Belledonne and northern ECM, likely a consequence of Penninic thrust inversion.

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Introduction

The pre-Triassic basement of the external Western Alps (Helvetic–Dauphinois zone) forms elongated crystalline uplifts at the front of the Penninic–Austroalpine orogenic wedge (Fig. 1). These uplifts, referred to as the External Crystalline Massifs (ECM), form two groups in map view (i) a northern branch including the Aar-Gottardo, Mont Blanc–Aiguilles Rouges, and Belledonne massifs, i.e. an almost continuous, NE-trending line of strongly elevated ECM; and (ii) a southern, more discontinuous branch, including the Pelvoux and Argentera–Mercantour massifs. Recent programmes of deep seismic reflection and refraction profiling have decisively increased our knowledge of the deep structure of the northern ECM (Roure et al., 1990; Blandell et al., 1992; Pfeiffer et al., 1997). These massifs correspond to thickened European upper crust (up to 35 km-thick, instead of 20 in the foreland regions) overlying the south-eastward plunging European lower crust (Moho at 40–45 km depth). In contrast, the deep structure of the southern massifs is still conjectural, and the Moho depth itself is constrained by relatively scarce seismic information in the southwestern Alps area (Kissling, 1993; Waldhauser et al., 1998).

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Up to the present, ECM denudation history has been studied more thoroughly in the northern massifs than in the southern ones. The stratigraphy and subsidence history of the flexural Swiss molasse basin were used by Mugnier and Méndez (1986) to suggest that crustal thickening (stacking of crustal slivers over a basal thrust) of the northern ECM was initiated during the early Miocene. Cooling and denudation histories were derived from zircon and/or apatite fission track data for the Aar and Mont Blanc massifs by Soom (1990), Michalski and Soom (1990), and Seward and Mancktelow (1994), and for the Belledonne massif by Lelarge (1995) and Sabli (1995). Preliminary FT data are also available for the Pelvoux massif (Seward et al., 1999). In the present paper, we focus first on the southernmost ECM, i.e. the Argentera–Mercantour massif, based on the published, geological and isotopic (40Ar/39Ar) data, and on apatite fission track (AFT) data thus far only available only in an unpublished doctoral dissertation (Mansour, 1991) and in a preliminary paper by Bigot-Cormier et al. (2000). Then, we address the interpretation of the striking differences, which appear in the Oligocene–Miocene cooling rates of the northern and southern ECM, and the possibility of an acceleration in denudation (and uplift?) rate(s) in the southern ECM during the Pliocene.

Geological outline of the Argentera massif

The Argentera–Mercantour massif (Fig. 2) consists of Variscan high-grade schists, migmatises and granites (Bogdanoff et al., 1991; with ref. therein). Tight synclines of Stephanian conglomerates derived from late Variscan grabens were pinched into the Valetta mylonitic zone during early Permian compressional events. Unconformable Permian red beds are found only at the southern border of the massif. Early Triassic sandstones unconformably overlie the older rocks, either around or within the massif (e.g. La Blache, Soubeyroux, Tortissa synclines), while the remaining of Mesozoic-Cenozoic cover was detached along the lower Mischelkalk and Keuper evaporitic formations. All around the massif, the Mesozoic cover sheet consists of epicontinental sediments, c. 2500 m thick. After a temporary emersion during the Palaeocene–early Eocene (Sturani, 1962; Fries, 1999), marine sedimentation started again with the 1500 m thick, late Eocene–early Oligocene ‘Grès d’Amont’ flysch. The latter sequence accumulated in response to flexural bending of the European crust at the front of the advancing Penninic wedge, and ended with the emplacement of the Embrunais–Ubaye and Maritimes Alps flysch nappes (Kerckhoff, 1969; Ravenne et al., 1987; Vially, 1994; Lickerish and Ford, 1998). The nappe emplacement was re-
Fig. 1 Geological sketch map of Western Alps and adjacent regions, with location of the study area (framed). Abbreviations: A.R., Aiguilles Rouges; B, Barrot dome; L.M., La Mure; R, Remollon; V, Verduches. Geological background after the geological map of France, scale 1/1,000,000, 6th edn (1996). Depth of Moho inland after Waldhauser et al. (1998). Ligurian-Provençal basin after Clamot-Rooke et al. (1997), and Mauffret et al. (1995). Ivrea geophysical body after Schmid and Kissling (2000).

Steeply close to the surface, particularly in the inner (north-east) part of the massif where back-thrust structures are observed (Fry, 1989). Compared with the northern ECM (Ménard, 1988; Guell et al., 1990; Hitz and Pfiffner, 1994), it is suggested here that the fault dip progressively decreases at depth below the Argentera, such that the faults would branch into a major shear zone at the top of the lower crust. In our interpretation, the deepest part of the Argentera upper crust is shifted to the NE relative to the topographic crest, consistent with the crustal model calculated by Calais et al. (1993) from geodetic data.

Argentera cooling and denudation history

From partial resetting of the $^{39}$Ar/$^{40}$Ar chronometer in plagioclase from NW Argentera rocks, Monić and Maluski (1983) concluded that temperature in these rocks reached 220–250°C between c. 40 and 33 Ma. We retain an age close to 33 Myr for the temperature maximum, because the burial of the Argentera basement was at a maximum during the earliest Oligocene. Assuming a 25°C km$^{-1}$ mean superficial geotherm (cf. Soom, 1990), and a burial of 9–10 km (summing 4–5 km of autochthonous–parautochthonous cover, and at least 5 km of Penninic nappes), the Argentera rocks would have been at least at 225–250°C at about 33 Ma. More precisely, the studied rocks of central Argentera (sampled at some depth beneath the former basement top) were heated at 250°C < T < 300°C as their zircon grains were reset for FT dating (yielding ages between 29 and 20 Myr; Bigot-Cornier et al., 2000), while their biotite ages are not for K/Ar (yielding Variscan ages; see Hunziker et al., 1992).

Apatite fission track analysis from central Argentera allows us to constrain the latest stages of cooling of the massif below ±100°C (e.g. Hunziker et al., 1992). Sixteen samples were collected from the NE-trending transect Isola–Pratolongo, and one additional sample (F3) from western Argentera (Figs 2, 4). After separation, the apatite grains were dated by the population method, using zeta calibration (Table 1, footnote). The ages vary from 8.05 ± 0.6–2.4 ± 0.2 Myr, with the older samples from
Fig. 2. Geological setting of the Argentera massif, after the geological map of France, scale 1:250,000, sheet 'Gap', and pers. obs. (S. B.). AA, cross-section Fig. 3. Framed, sampled area shown in Fig. 4a; the additional sample F3 (cf. Table 1) is shown close to Saint Etienne de Tinée. Abbreviations: B.F., Bersezio fault; C.F.F., Camp des Fourches fault; I.F., Inciamo fault.

Fig. 3. Generalized cross-section of the Argentera massif (location Fig. 2). Framed, sampled area and cross-section shown in Fig. 4. The interpretation at depth is speculative (see text). Sinistral shear between the Ivrea mantle backstop and the European basement after Ricou and Siddans (1986), and Schmid and Kissling (2000).
relatively higher topographic locations (Mansour, 1991).

The denudation rate for late Miocene–Pliocene time can be deduced from the age/altitude diagram (Fig. 5). Between c. 8 Ma and 3 Ma, the basement samples lie between two lines which give bounding limits for the denudation rate (e.g. Brown et al., 1994), 0.20 and 0.34 mm yr$^{-1}$, respectively. This indicates a mean denudation rate close to 0.25 mm yr$^{-1}$ in a moderately differentiated area. A closely similar age/altitude correlation (0.2 mm yr$^{-1}$) is obtained from independent AFT data by Bigot-Cormier et al. (2000) in central–eastern Argentera. A break in the slope of the age/altitude plot (Fig. 5) at about 3 Ma suggests an increase of the exhumation rate up to 0.8–1.5 mm yr$^{-1}$ (dashed lines), with a mean rate of c. 1 mm yr$^{-1}$, nicely consistent with the results obtained by Bigot-Cormier et al. (2000) in central and eastern Argentera.
Table 1 Apatite fission track analytical data and ages

<table>
<thead>
<tr>
<th>Sample</th>
<th>m</th>
<th>( n_f )</th>
<th>( n_i ) ( (\sigma) )</th>
<th>( n_t ) ( (\sigma) ) ( (10^2 ; t^{-2}) )</th>
<th>( n_i ) ( (\sigma) ) ( (10^2 ; t^{-2}) )</th>
<th>( N_d )</th>
<th>( \rho_f ) ( (10^3 ; t^{-1}) )</th>
<th>( \tau ) ( (1 ; t^{-1}) ) ( (\tau) )</th>
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<td>99</td>
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<td>130</td>
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<td>92</td>
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<td>180</td>
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<td>330</td>
<td>360</td>
<td>2.89 ± 0.08</td>
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<td>330</td>
<td>360</td>
<td>0.70 ± 0.06</td>
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</table>

\( m \), altitude, in meters; \( n_f, n_i, N_d \), numbers of apatite grains (n) and number of tracks counted (N), with suffixes f and i for fossil and induced tracks, respectively; \( N_d \), number of tracks counted in muscovite micas associated to glass doismeters NBS 1562. Errors on track densities were calculated following McGee et al. (1983).

Sample 15 was counted twice by the same observer (M. Mansouf) and sample 16 by two observers (weighted means shown in bold characters). Ages were calculated using zeta values of, respectively, 314 ± 4.4 and for the one with an age of 313 ± 20.

However, the dispersion of the ages of the low-elevation samples could be at least partly a consequence of lateral cooling effect of topographic relief (Stüwe et al., 1994; Mancktelow and Grasemann, 1997). If we apply to the quoted apparent rates the correction suggested by Stüwe et al. (1994), the real denudation range of central Argentiera in the last 3,5–4 Ma would be close to 0.8–1.0 mm yr⁻¹. Remarkably, during the same time interval, the massif elevation increased by about 1 km (Fauquette et al., 1999).

The earliest cooling stages of the Argentiera massif through erosional denudation may be reasonably dated from the early Oligocene, since pebbles from the Penninic nappe (which formed the upper levels of the lid overlying the Argentiera–Pelvoux basement at that time) are abundant in the San-niostean (Rupelian) conglomerates of the Bârène syncline (Griciasans, 1972; Evans and Mange-Rajetzky, 1991), at about 31 Ma (Artomi and Meckel, 1998). Erosional denudation went on during the late Oligocene-Miocene, affecting the Dauphinois cover (cf. pebbles in the Lower–Middle Miocene molasses of the Varensole basin; Haeckel et al., 1989), and finally the Permian and basement rocks in the late Miocene (pebbles in the Roquebrune conglomerates close to Nice; Fauquier and Couris, 1960). Tectonic denudation likely played an increasing role in the denudation process from late Miocene onward (see below). For the 33–8 Ma interval, corresponding approximately to cooling of the Argentiera rocks from 250°C < T < 300°C to

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$T = 100^\circ C$, we may calculate a mean cooling rate of 7–8°C Myr$^{-1}$ (Fig. 6, bold curve). This relatively slow cooling rate is similar to that deduced from the AFT age/altitude diagram (Fig. 5) for the 8–3 Ma interval, assuming a 25°C km$^{-1}$ geotherm (6°C Myr$^{-1}$). Both rates contrast with the post 3 Ma cooling rate calculated from the same diagram (25–35°C Myr$^{-1}$).

Discussion and conclusions

Cooling and denudation rates

The juxtaposition of the various ECM cooling paths (Fig. 6) shows that the northern massifs followed much steeper paths than the southern ones during Oligocene–Miocene times. Indeed, burial depth was much greater in the north than in the south. Temperature remained in the 220–250°C range in the Argentera and Pelvoux massifs, but reached 350°C in the Mont Blanc and Aar massifs (e.g. Desmons et al., 1999). This longitudinal thermal gradient is strikingly visible on the K/Ar and Rb/Sm age compilation maps by Hunziker et al. (1992), with Varscan or partially reset biotite ages in the Argentera, Pelvoux, southern Belledonne and Aiguilles Rouges massifs, and Eocene–Oligocene biotite ages in the Mont Blanc and Aar massifs. The change in burial depth is related to the change in thickness of the tectonic overload (Penninic wedge), as demonstrated by the concomitant change in the metamorphic grade of the Dauphinois–Helvetic flysch formations, from diagenesis in the Argentera area to anchizone/low-grade metamorphism in internal Pelvoux and Belledonne areas (Lekin et al., 1983; Tricart, 1984; Desmons et al., 1999), to epidote–greenish schist facies north-west of Mont Blanc, and, finally, to chloritoid-bearing greenish schist facies around the Aar massif (Desmons et al., 1999). Hence, the Penninic wedge overloads varied from 3 to 5 km in the Argentera area, to 4–8 km in the Pelvoux and south Belledonne, to 10–14 km in the Mont Blanc and Aar, and the maximum burial depth varied from 8 to 10 km for the southern ECM, to 12–16 km for the northernmost ones. On the other hand, denudation of the ECM was roughly contemporaneous all along the Western Alps, as shown by the sedimentary record in the peri-Alpine molasses. Denudation started during the Oligocene (erosion of the Penninic overloads), and was achieved during the Miocene (erosion of the Dauphinois–Helvetic cover, then of the crystalline basement itself), such as pebbles originating from the ECM appear elsewhere in the late Miocene molasses (Triumphy, 1980; Mugnier and Ménard, 1986). Accordingly, the mean denudation and cooling rates between c. 30 and 10 Ma had to be much greater in the northern ECM than in the southern ones. Using a 25°C km$^{-1}$ geotherm, the mean cooling rates reported for the Oligocene–Miocene, i.e. for the high temperature range (Fig. 6), correspond to denudation rates of 0.2–0.3 mm yr$^{-1}$ for the southern ECM, and 0.5–0.6 mm yr$^{-1}$ for the northern ECM.

In the low temperature range, AFT age–elevation diagrams from the Belledonne and Argentera massifs would
suggest an increase of the denudation rate at 3–4 Ma (Fig. 7). Some track length optimization runs (Gallagher, 1995) for Belledonne samples also yielded evidence of cooling acceleration at about 2–3 Ma (Sabił, 1995). From that time up to the Present, the denudation rate would have reached 1.0–1.5 mm yr$^{-1}$ in the Argentera and Belledonne as in the northern ECM. This apparent rate is overestimated as a result of the effect of erosional topography (Stüwe et al., 1994; Mancktelow and Grasemann, 1997), and would correspond to a real denudation rate of 0.8–1.2 mm yr$^{-1}$ after Stüwe et al. (1994) correction. If we accept a broadly steady-state uplift-denudation regime, then a Pliocene denudation rate of 0.1 mm yr$^{-1}$ would be remarkably consistent with the present-day uplift rate of 0.9–1 mm yr$^{-1}$ observed in the Aar–Mont Blanc (Kahle et al., 1997) and Belledonne massifs (Fourniquet in Ménard, 1988).

Tectonic processes

Uplift and exhumation of the northern ECM were driven by compressional stacking of upper crustal slices detached from the internally dipping lower crust, and a similar thickening process is likely in the Argentera (see above). However, shortening of the Alpine realm was more important along the leading (northern) edge of the Adriatic block than along its western boundary (e.g. Schmid et al., 1996). From balanced cross-sections at crustal scale, shortening of the external European upper crust was evaluated at about 30 km and 50 km in the Belledonne and Mont Blanc transects, respectively (Ménard and Thouvenot, 1987; Sommaruga, 1997), while Hitz and Pfiffner (1994) suggest a 25-km shortening for the Aar massif. In the Argentera case, Lickorish and Ford (1998) provide an estimate of c. 10 km, based on the structure of the Mesozoic–Cenozoic sheet. Although there are only a few constraints on the individual basement faults, summing their reverse throw along a NE-trending transect (Fig. 3) would rather provide a bulk shortening of about 15 km, which is approximately half of the shortening estimate for the northern ECM. The fact that the Argentera displays about the same crustal thickness (c. 40 km, cf. Figure 1) as the northern ECM, in spite of its lesser collisional shortening, may be explained by the greater thickness of the Argentera crust at the end of the Mesozoic rifting, with respect to the typical Dauphinois–Helvetic crust. Indeed, Mesozoic extension of the European margin was less in the Argentera than in the other ECM, as the corresponding Jurassic–Cretaceous sedimentary facies of the eastern Digne sheet (Maritimes Alpes) are thinner and more calcareous than that of the typical Dauphinois domain, more to the west or north-west (Dardeau, 1988; Friès, 1999).

Isostatic uplift of the thickening crust triggered erosion of the sedimentary lid over the Argentera basement during the Oligocene (see above), while tectonic denudation lagged until the Miocene, with the development of SW-dipping normal faults along the SW border of the massif (cf. Campdes-Fourches fault, Fig. 2; Labaune et al., 1989). Eventually, it can be expected that the post 5 Ma inversion of the Penninic thrusts (Penninic Front, Houiller Front) documented on the Mont-Blanc and Belledonne transects (Seward and Mancktelow, 1994; Fügenschuh et al., 1999) also occurred on the Argentera transect, hence accounting for the high, recent denudation rate (Fig. 7).

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References


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