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Exhuming the Alps through time: clues from detrital zircon fission-track thermochronology

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ABSTRACT

The European Alps are a mountain belt that is characterized by a series of discrete orogenic events, which have long been recognized. Despite the inherent episodic nature of orogenic evolution, the Alps have been continuously exhumed, mainly by erosion, but also by normal faulting. Since continental collision started in the late Eocene/Early Oligocene evidence for ongoing erosional exhumation has been preserved in synorogenic sediments that accumulated in basins adjacent to the pro- and retro-side of this double-vergent mountain belt. This long-term erosion record can be used to determine exhumation rates. Lag-times calculated from fission-track (FT) ages of detrital zircon from synorogenic sediments are fairly constant for the European Alps since the Oligocene–Late Miocene. Although the fast exhuming areas were unroofed at rates of 0.4–0.7 km Myr^{-1}, the overall average exhumation rate is between 0.2 and 0.3 km Myr^{-1} on a regional scale. The detrital and bedrock zircon FT data of the Alps do not detect the increase in erosion rates since the Pliocene over the past $\sim$5 Myr, as shown elsewhere. This increase cannot be detected yet with the detrital zircon FT method because not enough rock has been removed to widely expose zircons with Pliocene or younger cooling ages in the Alps. Long term (30 Myr) exhumation rates appear to have been approximately constant when averaged over a sliding time window of about 8 Myr, or depth window of 5 to 10 km (ZFT closure depths); shorter-term fluctuations are not identified using this method.

INTRODUCTION

The orogenic evolution of the European Alps (Fig. 1) as a series of discrete periods of tectonic, metamorphic and erosional activity during Cretaceous, Tertiary and Quaternary times is well established and widely accepted (e.g. Hubbard & Mancktelow, 1992; Steck & Hunziker, 1994; Schmid *et al.*, 1996, 2004; Frey & Mählman, 1999; Rosenbaum & Lister, 2005). This understanding of the Alps is derived from numerous local and regional studies of different tectonic and erosional events (e.g. Mancktelow, 1985; Selverstone, 1988; Seward & Mancktelow, 1994; Schlunegger *et al.*, 1997, 2001; Schlunegger, 1999; Kuhlemann *et al.*, 2001 and references therein). Many of these studies presented deformation and exhumation of the Alps by precisely reconstructing the deformation and exhumation of individual particle points. The results of these studies showed that the history of the European Alps is punctuated by orogenic events such as movement of nappes, by times of locally high exhumation rates and times of quiescence. The evolution of the Alps in a series of discrete events is not questioned here, but the intention of this paper is to look at the history of the Alps since the Eocene/Oligocene from a different point of view by studying the overall exhumational signal of the orogen, rather than tracking exhumation of specific material points or rocks within the orogen. For that reason the thermochronology of detrital zircon is used to reconstruct the long-term exhumational history.

Zircons were eroded off the surface of this mountain belt at least since the Eocene and they are preserved in synorogenic sediment to the north, west and south of the Central and Western Alps. Twenty-four samples were collected from the stratigraphic record of synorogenic sediments of Alpine foreland and hinterland basins for detrital zircon fission-track (FT) analysis. The hinterland basin is here defined as the retro-side basin of the Alpine wedge and most of those sediments are nowadays exposed in the northern Apennines. The new FT data are used in combination with previously published results (Bernet *et al.*, 2001, 2004a), to determine Alpine exhumation rates, but not sediment yield, as will be discussed below. There-

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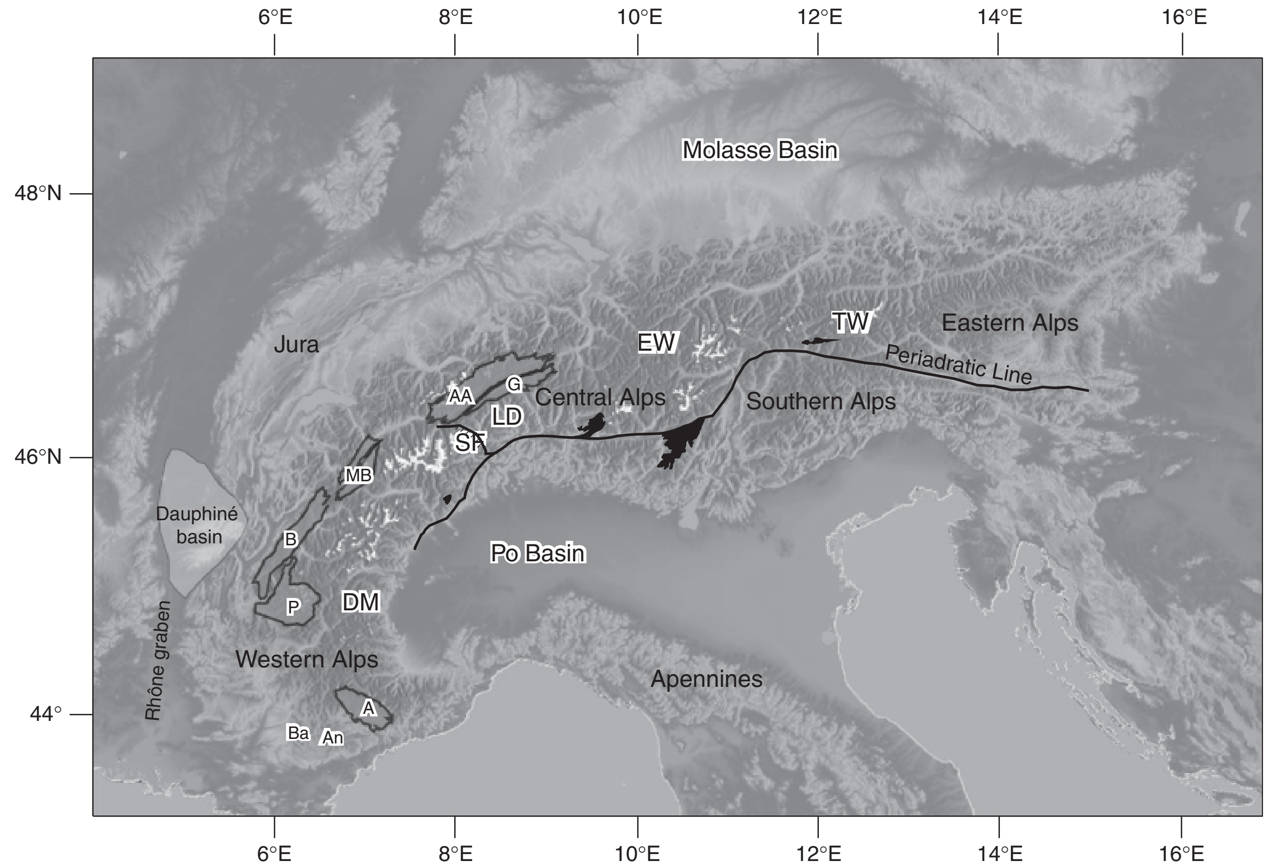


Fig. 1. Digital elevation model of the Alps and northern Apennines. A, Argentera massif; P, Pelvoux massif; B, Belledonne massif; MB, Mont Blanc massif; AA, Aar massif; G, Gotthard massif; DM, Dora Maira massif; LD, Lepontine dome; SF, Simplon Fault; EW, Engadine Window; TW, Tauern Window; Ba, Barrême Basin; An, Annot Basin Shown in black are periadriatic intrusions.

fore, the results of this study are not directly comparable with the work of Kuhlemann (2000) and Kuhlemann *et al.* (2001, 2002) who studied erosion of the Central, Western and Eastern Alps in terms of sediment yield.

GEOLOGIC SETTING

A full review of the Alpine orogeny is beyond the scope of this paper, but useful reviews and summaries can be found by Schmid *et al.* (1996, 1997, 2004), and Rosenbaum & Lister (2005). Convergence between the European plate to the north and the Adriatic plate to the south started more or less during the late Cretaceous with rates of 1.5–1.3 cm year^{-1} between the early Palaeocene and late Eocene (Schmid *et al.*, 1996). Successive accretion of terrains and nappe movement culminated with the collision of the European plate and the Adriatic plate during the late Eocene and convergence velocity slowed to on average 0.5–0.3 cm year^{-1} between the late Eocene and the present (Schmid *et al.*, 1996). Continental collision was characterized by widespread barrowvian metamorphism in the Central Alps, slap break-off of the subducting European lithosphere around 33–30 Ma, and back thrusting of the nappe stack on the retro-side of the mountain belt (Blanckenburg & Davies, 1995; Davies & Blanckenburg, 1995; Schmid *et al.*, 1996; Sinclair, 1997b). That convergence continued after collision is documented in the thrusting along the Insubric line and propagation of the Helvetic nappes between 32 and 19 Ma, and foreland propagation in the Southern Alps since 19 Ma (Schmid *et al.*, 1996). If the Alps are still active today is a matter of debate. Accretionary growth by convergence has most likely ceased by now, but metamorphism continues beneath the Alps today (Steck & Hunziker, 1994), and active deformation is observed in earthquake and GPS analyses (e.g. Sue *et al.*, 2000; Martinod *et al.*, 2001). Selverstone (2005) postulated that at least the Central Alps are not collapsing yet, as would be expected in a decaying mountain belt, but extension and normal faulting is a widespread feature in the Western Alps.

Continental collision and continuous convergence since the Eocene led to crustal thickening and loading of the subducting European plate causing the development of a peripheral foreland basin to the north and west of the evolving Alps, where deposition started in the late Eocene and continued throughout the Oligocene until the late Miocene (Homewood *et al.*, 1986; Pfiffner, 1986; Allen *et al.*, 1991; Schmid *et al.*, 1996; Escher & Beaumont, 1997; Gupta & Allen, 2000). The northern foreland basin is well preserved in southern Germany and Switzerland. The western foreland basin, however, is only partially preserved and Eocene to Miocene deposits are either found in thrust-top basins (e.g. Barrême basin and Annot basin) east of the Digne-thrust or along parts of the Rhône graben system and Dauphiné basin in southeastern France (Fig. 1).

The earliest sedimentary deposits in this foreland basin indicate a marine transgression with development of a carbonate ramp (numulitic limestones) on the subducting plate (e.g. Sinclair, 1997a; Gupta & Allen, 2000). These sediments were followed by deeper water deposition of hemipelagic marls and finally by deep-water siliciclastic turbidites in close proximity of the advancing Alpine thrust front (Gupta & Allen, 2000). A typical example of such siliciclastic turbidites are the Annot sandstone units in the western foreland basin. They are interpreted to represents the onset of clastic sedimentation in the foreland basin, indicating early exhumation of Alpine rocks (Sinclair, 1994). The main sediment source of the Annot sandstone is believed to be the Alpine Corsica–Sardinia block now exposed offshore to the southwest (Stanley & Mutti, 1968; Ivaldi, 1974; Sinclair, 1994). Turbidite sedimentation continued in the foreland basin at least until Mid-Oligocene times (Allen *et al.*, 1991; Sinclair, 1997a, b; Gupta & Allen, 2000), and Alpine provenance for Oligocene deposits in the Barrême thrust-top basin of the Digne thrust sheet in southeastern France for example was demonstrated by Evans & Mange-Rajetzky (1991).

The depositional environment in the Alpine foreland basin changed from a deep marine setting during the Oligocene to a shallow marine setting in the Early Miocene, and then to a more continental or molasse setting during the Late Miocene (Allen *et al.*, 1985; Sinclair *et al.*, 1991, 1997a, b; Schlunegger *et al.*, 1997, 2001; Kuhlemann *et al.*, 2001; Schlunegger & Simpson, 2002). The end of sedimentation in the northern foreland basin occurred between 8.5 Ma (Lemcke, 1988; Zweigel, 1998) and 4.5 Ma (Cederbom *et al.*, 2004). Some of the Eocene to Miocene foreland basin deposits were incorporated into the advancing thrust front, exhumed to the surface and are subject to recycling since the Late Miocene (Pfiffner, 1986; Schlunegger *et al.*, 1997; Kuhlemann *et al.*, 2001). Pliocene deposits are very rare in the northern and western part of the foreland basin. Nevertheless, the Pliocene record of Alpine exhumation is preserved in the Rhône and Rhine deltas (Hay *et al.*, 1992; Kuhlemann, 2000).

Not all the sediment eroded off the Alpine mountain belt was deposited in its foreland basin. From Late Eocene to Pliocene times, Alpine derived sediment was also deposited in thick, siliciclastic turbidite sequences in the northern Apennines in Italy (Fig. 1). Turbidite deposition occurred in asymmetric mobile basins developed on continental crust of the western Adriatic margin in various palaeogeographic domains. The Upper Oligocene to Upper Miocene turbidites of the northern Apennines were mainly deposited in the Tuscan domain and Umbrian–Marchean domain (e.g. Vai & Castellarin, 1992 and references therein). The turbidite sequences of the Tuscan domain consist of the Macigno, Monte Modino and Cervarola–Falterona sandstones, whereas the Umbrian–Marchean domain is dominated by the Marnoso-arenacea Formation. Sediment petrologic data and dominant palaeocurrent directions (NW–SE) of turbidites in the Tuscan and Umbrian–Marchean domains indicate strong Alpine sediment delivery. Therefore, most authors conclude that the ultimate sediment source of these turbidite units were in the Western and Central Alps (Bortolotti *et al.*, 1970; Sestini, 1970; Gandolfi *et al.*, 1983; Valloni & Zuffa, 1984; Ricci Lucchi, 1986; Sestini *et al.*, 1986; Bruni *et al.*, 1994; Aruta *et al.*, 1998; Di Giulio, 1999; Cibin *et al.*, 2001; Garzanti & Malusà, 2008). Sediment input from the slowly emerging Apennine thrust front is locally restricted and easily recognizable in olistostromes and tectonosomes intercalated in the turbidite deposits (e.g. Pini, 1999 and references therein).

Exhumation in the Alps, is driven by a combination of erosion and normal faulting, as in any other convergent orogen (Ring *et al.*, 1999), and large-scale extensional features such as the Simplon fault or the Tauern window (Fig. 1) are clear evidence of tectonic exhumation (Mancktelow, 1985; Behrmann, 1988; Selverstone, 1988; Grasemann & Mancktelow, 1993; Fügenschuh *et al.*, 1997; Schlunegger & Willett, 1999). However, extensional tectonics is not necessarily an indication for orogenic collapse of a decaying mountain belt, as horizontal extension has been recognized in many active convergent mountain belts (Ring *et al.*, 1999). Zircon FT ages of currently exposed bedrock reflect a distinct age pattern, with relatively young (<36 Ma) Alpine orogenic cooling ages in large parts of the Central and Western Alps and older cooling ages in many areas of the Eastern and Southern Alps (Fig. 2; e.g. Hunziker *et al.*, 1992). The 36 Ma and younger cooling ages are seen in relation to cooling after widespread Middle to Late Eocene greenschist and amphibolite facies metamorphism in the Central and Western Alps (see papers in Frey *et al.*, 1999).

SAMPLE COLLECTION, PROCESSING AND DATA HANDLING

Twenty-four medium to coarse-grained sandstone samples of 4–7 kg each were collected in the field. The sampling strategy was to obtain samples from different stratigraphic horizons of the Eocene to Miocene foreland basin deposits and the Oligocene to Pliocene hinterland basin deposits. Twelve samples from outcrops of turbidite sequences and shallow marine or fluvial sandstone of known stratigraphic position in the western foreland basin (southeastern France) and the northern foreland basin (southern Germany) were collected (Fig. 3a). Twelve hinterland samples were collected from turbidite sequences of known stratigraphic age (Fig. 3b). The stratigraphic positions of all sampled units are shown in Fig. 4. Stratigraphic ages are known on the bases of biozones and have an estimated error of $\pm$ 1 Myr.

All samples were processed in the same way at the FT laboratory at Union College, Schenectady, NY. Heavy minerals were separated with standard heavy liquid and magnetic separation techniques after crushing and pulverizing. Zircon aliquots of each sample were mounted in

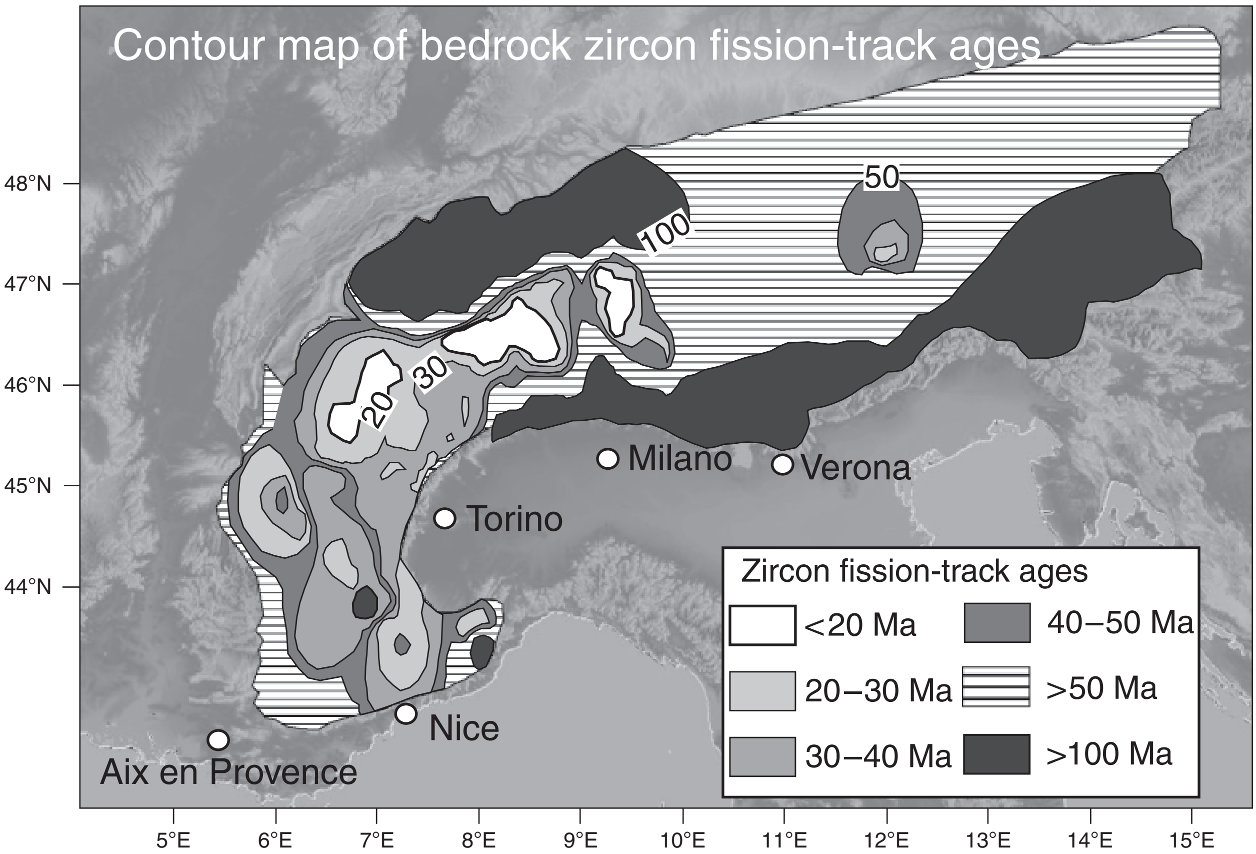


Fig. 2. Bedrock zircon fission-track (FT) age contour map (modified from Bernet *et al*., 2001).

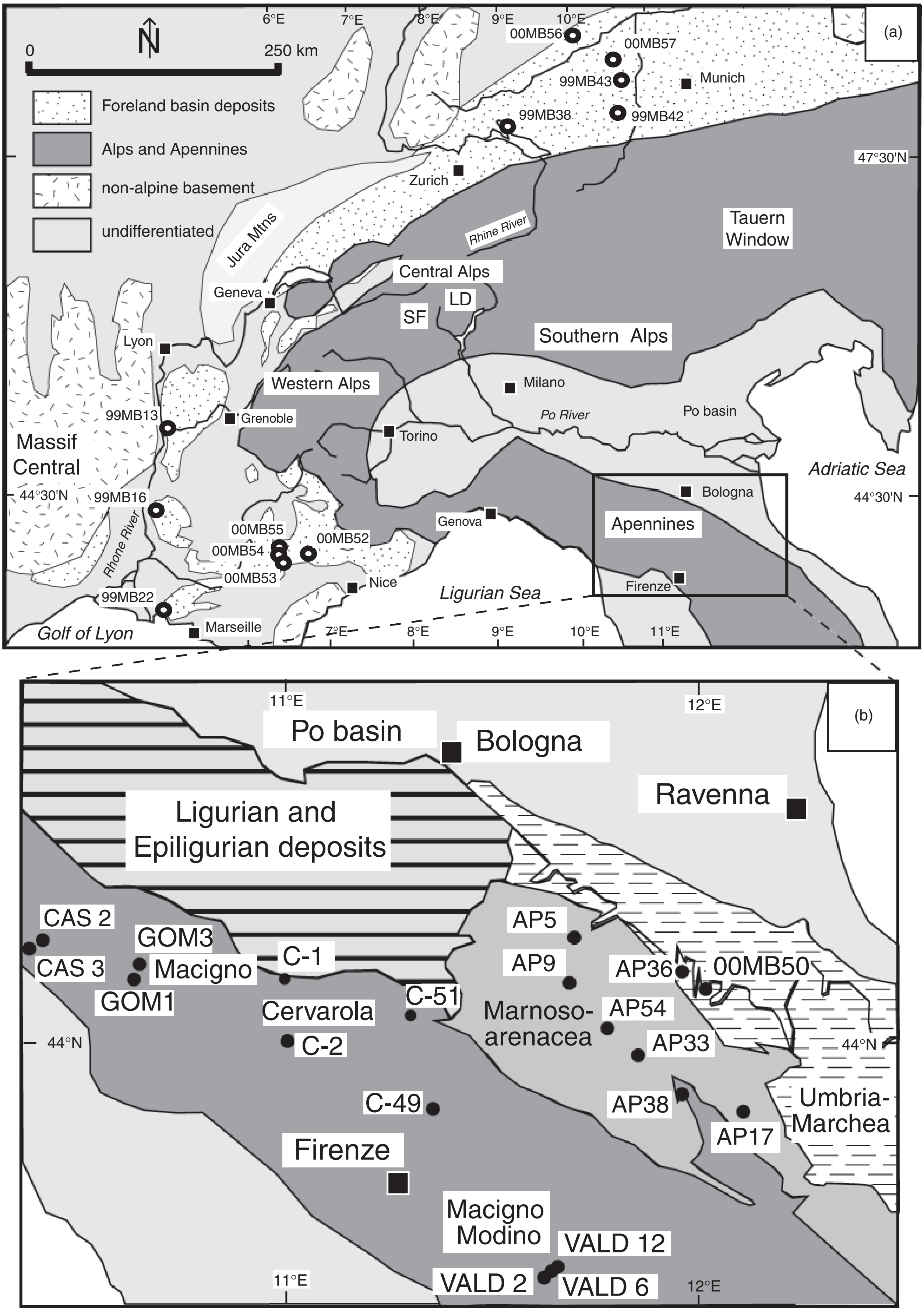
Teflon® (Du Pont, Nemors, Luxembourg). Following Naeser *et al*. (1987) and Bernet *et al*. (2004b) a multi-mount technique with two mounts per sample was used, each mount containing 500–1000 grains. Mounts were polished to expose flat zircon surfaces and then etched. One mount per sample was etched in a short etch (10 h) and one mount in long etch (24 h) in a NaOH–KOH eutectic melt at 228 °C in a thermostatically controlled oven. All samples were first pre-etched for 1 h to remove metamict zircon, and other impurities; then they were etched in fresh etchant for the rest of the selected etch time.

Because the external detector method for FT dating was used, pre-annealed, low-uranium mica was attached to each sample to record induced tracks as an external detector during irradiation. In general, between 50 and 100 grains were dated per sample. All observed FT grain-age distributions were decomposed into main grain-age components using the binomial peak-fit method after Galbraith & Green (1990) and Brandon (1992, 1996). Peak ages (P1, P2, P3, etc.) were calculated with the BINOMFIT program from Brandon, as described in Ehlers *et al*. (2005).

RESULTS

The range of single grain ages, the mean ages and the different peak ages of each sample are presented in Tables 1 and 2, in addition to previously published results. Single grain ages of all of the ~2000 zircons dated for this study are available from the corresponding author upon request. Looking at the data, the following observations are remarkable: (1) the spread of ages of about one order of magnitude in all samples is typical for detrital samples and each sample fails the χ^2-test. Even if every sample contains a few grains that are close to or slightly younger than the depositional age, no indication of partial or full resetting of zircons after deposition has been observed. The minimum age, the youngest population of grain-ages in each sample is older than the depositional age. Therefore, all grain-age components are regarded to reflect pre-depositional cooling. (2) The percentage of grains with Alpine cooling ages, again here defined as <36 Ma, increases overall up-section in both the foreland and hinterland deposits (Fig. 5). (3) The mean ages of each sample of the foreland and hinterland suites are shown in

Fig. 3. Simplified map showing sample locations of the western and northern Alpine foreland basin and of the hinterland basin in the Northern Apennines (Tables 1 and 2). (a) Foreland basin samples were collected from a variety of locations in southeastern France and southern Germany. Northern foreland basin samples 99MB42 and 43 were obtained from cores of the Kaufbeuren-1 and Siebnach-1 drill holes (courtesy of BEB Hannover). (b) Hinterland samples were collected from turbidite sequences of the Oligocene to Miocene M. Senario Formation of the sub-Ligurian units, from the Macigno and Cervarola–Falterona sandstones of the Tuscan domain, and from the Pliocene Borello Sandstone of the Umbrian–Marchean domain (Table 2). Previously published detrital zircon fission-track (FT) ages from the Miocene Marnoso-arenacea Formation (AP samples from Bernet *et al*., 2001) are shown for completeness.


(a)
0
250 km
Foreland basin deposits
Alps and Apennines
non-alpine basement
undifferentiated
6°E
7°E
8°E
9°E
10°E
00MB56
00MB57
99MB43
Munich
99MB38
99MB42
47°30'N
Zurich
Jura Mtns
Rhine River
Tauern Window
Central Alps
Geneva
LD
SF
Southern Alps
Lyon
Western Alps
Milano
Po River
Po basin
99MB13
Grenoble
Torino
Massif Central
Adriatic Sea
44°30'N
99MB16
Bologna
Rhone River
00MB55
Genova
Apennines
00MB54
00MB52
00MB53
Nice
Ligurian Sea
Firenze
99MB22
Golf of Lyon
Marseille
11°E
(b)
11°E
12°E
Po basin
Bologna
Ravenna
Ligurian and Epiligurian deposits
CAS 2
GOM3
AP5
CAS 3
Macigno
C-1
AP9
AP36
00MB50
GOM1
C-51
Cervarola
Marnoso-arenacea
AP54
44°N
C-2
AP33
C-49
AP38
Umbria-Marchea
AP17
Firenze
Macigno Modino
VALD 12
VALD 2
VALD 6

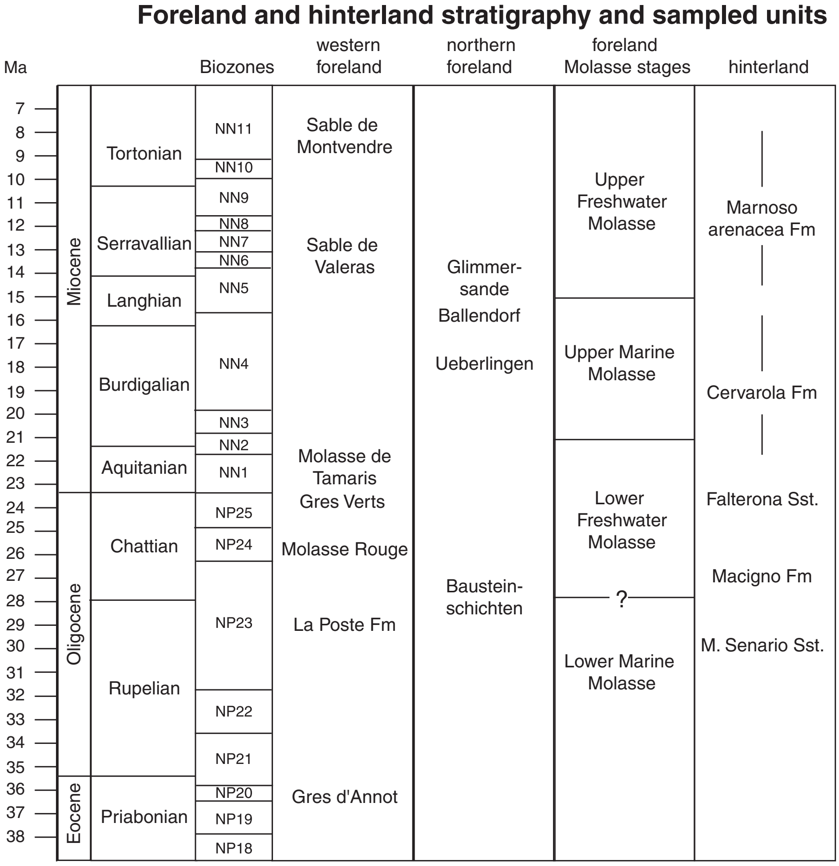


Fig. 4. Simplified stratigraphic column of the northern and western Alpine foreland basin based on Lemcke (1988), Zweigel (1998) and Pomerol (1980), and for the Alpine hinterland. Note that only units and locations sampled for this study are shown. Sample numbers are given in Tables 1 and 2. Biostratigraphic zones are based on Harland *et al.* (1989) and Berggren *et al.* (1992).

Table 1. Detrital zircon fission-track (FT) data from western and northern foreland basin sediments

Samples	Biozones	Deposition (Ma)	*N*	Age range (Ma)	P1	P2	P3	P4
Western foreland basin								
99MB13	NN11	8.0	60	11.5–184	17.2 ± 1.8	28.3 ± 5.1	99.8 ± 11.9	
Sable de Montvendre					52%	25%	23%	
99MB16	NN7	13.0	60	14.2–195	22.0 ± 1.8	63.9 ± 9.8	121.8 ± 26.7	
Sable de Valeras					66%	19%	15%	
99MB22	NN4	16.0	60	18.3–142	24.0 ± 2.0	66.0 ± 10.7	116.4 ± 18.5	
Etang du Lavalduc					59%	19%	22%	
00MB53	NP24	26.0	11	31.9–221		no peaks fitted		
Grès Verts								
00MB54	NP23	28.0	60	23.5–266	34.4 ± 3.3	59.7 ± 7.7	152.9 ± 17.5	
Molasse Rouge					45%	26%	29%	
00MB55	NP23	30.5	50	22.7–160	32.4 ± 2.8	60.4 ± 10	111.1 ± 14.4	
La Poste					50%	16%	34%	
00MB52	NP20	36.0	50	35.3–341	60.9 ± 9.0	122.3 ± 19.5	–	
Grés d'Annot					51%	49%		
Northern foreland basin								
00MB57	NN5	14.0	60	17.2–196	23.4 ± 4.2	45.0 ± 4.3	82.5 ± 19.2	140.2 ± 64.1
Glimmersande					12%	53%	27%	8%
99MB38	NN4	16.0	60	12.9–320	25.6 ± 3.2	69.8 ± 8.5	144.3 ± 17.7	
Überlingen					30%	37%	33%	
00MB56	NN4	18.0	60	17.3–308	22.2 ± 2.8	35.8 ± 4.8	78.7 ± 22.0	129.6 ± 17.8
Ballendorf					29%	21%	16%	34%
99MB42	NP23	27.0	60	34.3–188	41.3 ± 4.3	72.6 ± 14.7	124.9 ± 14.6	
Baustein-schichten					33%	24%	44%	
99MB43	NP23	27.0	60	29.1–293	42.4 ± 4.9	85.6 ± 13.6	140.9 ± 36.2	
Baustein-schichten					37%	40%	23%	

n = total number of grains counted; binomial peak-fit ages are given ± 2 SE. Also given is the percentage of grains in a specific peak. All samples were counted at 1250 × dry (× 100 objective, 1.25 tube factor, 10 oculars) by M. Bernet using a zeta (CN-5) of 334.22 ± 3.40 (± 1 SE). Depositional ages after Pomerol (1980), Lemcke (1988), Evans & Mange-Rajetzky (1991) and Zweigel (1998). Biozones after Harland *et al.* (1989) and Berggren *et al.* (1992).

Table 2. Detrital zircon fission-track (FT) data from hinterland basin sediments

Samples	Biozones	Deposition (Ma)	n	Age range (Ma)	P1	P2	P3	P4
00MB50 Borello Sst.	NN14	3.7	90	8.8–114	13.9 ± 2.1 30%	24.2 ± 3.1 42%	38.5 ± 7.9 21%	72.9 ± 22.1 7%
C1 Cervarola Sst.	NN4	16.0	60	14.4–112	27.2 ± 2.3 83%	82.1 ± 12.5 17%	–	–
C51 Cervarola Sst.	NN4	18.0	100	19.0–104	22.1 ± 3.5 15%	33.2 ± 3.1 69%	72.9 ± 10.3 16%	.
C2 Cervarola Sst.	NN2	21.5	60	16.1–124	28.8 ± 3.3 52%	43.0 ± 6.2 40%	80.2 ± 17.2 8%	–
VALD 12 Cervarola Sst.	NN1	22.0	49	22.0–113	29.2 ± 2.1 77%	39.6 ± 6.5 12%	65.0 ± 16.4 5%	106.8 ± 23.5 6%
VALD 2 Falterona Sst.	NP25	24.0	46	23.1–147	29.4 ± 2.7 49%	75.1 ± 7.1 51%	–	–
VALD 6 Falterona Sst.	NP25	25.0	51	20.3–141	29.7 ± 2.2 70%	57.4 ± 9.0 20%	103.9 ± 22.1 10%	–
GOM-1 Macigno Fm	NP24	26.0	46	21.2–104	33.9 ± 2.2 83%	57.2 ± 8.1 14%	100.7 ± 38.6 3%	–
GOM-3 Macigno Fm	NP24	26.0	50	26.0–69	33.3 ± 2.3 92%	50.9 ± 19.5 8%	–	–
CAS-3 Macigno Fm	NP23	28.0	49	28.0–109	35.5 ± 2.5 87%	59.0 ± 10.9 11%	111.2 ± 80.5 2%	–
CAS-2 Macigno Fm	NP23	28.0	49	29.5–153	41.0 ± 3.7 65%	83.1 ± 11.7 35%	–	–
C49 M. Senario Sst.	NP23	30.0	100	19.2–154	32.3 ± 5.0 36%	50.8 ± 9.8 43%	83.9 ± 14.6 21%	–

n, total number of grains counted; binomial peak-fit ages are given ± 2 SE. Also given is the percentage of grains in a specific peak. All samples were counted at 1250 × dry (× 100 objective, 1.25 tube factor, 10 oculars) by M. Bernet using a zeta (CN-5) of 334.22 ± 3.40 (± 1 SE). Biozones after Harland *et al.* (1989) and Berggren *et al.* (1992).

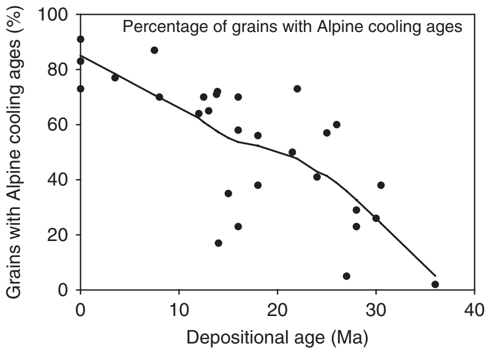


Fig. 5. Diagram showing the increase of dated grains with Alpine (<36 Ma) cooling ages in the detrital samples from synorogenic and modern river sediment (Bernet *et al.*, 2004b).

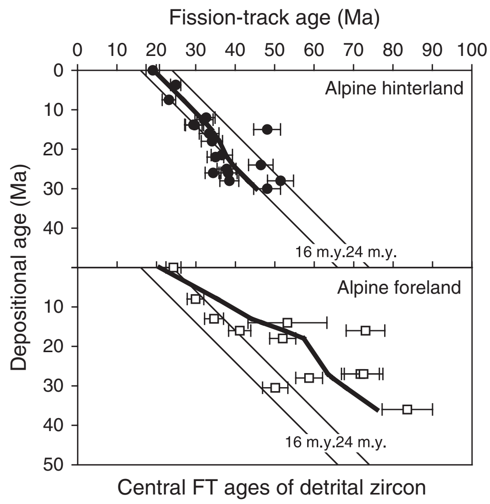


Fig. 6. This graph shows a mean-age lag-time plot for the foreland and hinterland samples. Hinterland samples demonstrate a similar trend as the P1 and P2 lag-time plots in Fig. 7, whereas the foreland samples contain a larger fraction of older grains, which is evidence for the removal of non-reset cover-units towards the foreland.

Fig. 6. Both suites show similar trends of younging up-section. (4) Binomial peak fitting revealed that each sample contains three to four main peaks and the different peaks show similar up-section age trends in both the foreland basin and hinterland basin deposits. The younger peaks, P1 and P2, are moving peaks because they are getting considerably younger up-section (Fig. 7a and b). The older peaks of about 60–140 Ma, usually P3 and P4, show less systematic variation up-section (Tables 1 and 2). The actual peak ages are of more importance than peak sizes. Variation in peak sizes from sample to sample is a common observation, because in addition to sampling error, the actual peak size depends on short-term erosion rates, which are decoupled from long-term exhumation rates (Bernet *et al.*, 2004b).

It should be noted that a peak age (for example P3) of one sample might fall within the age group of a different peak

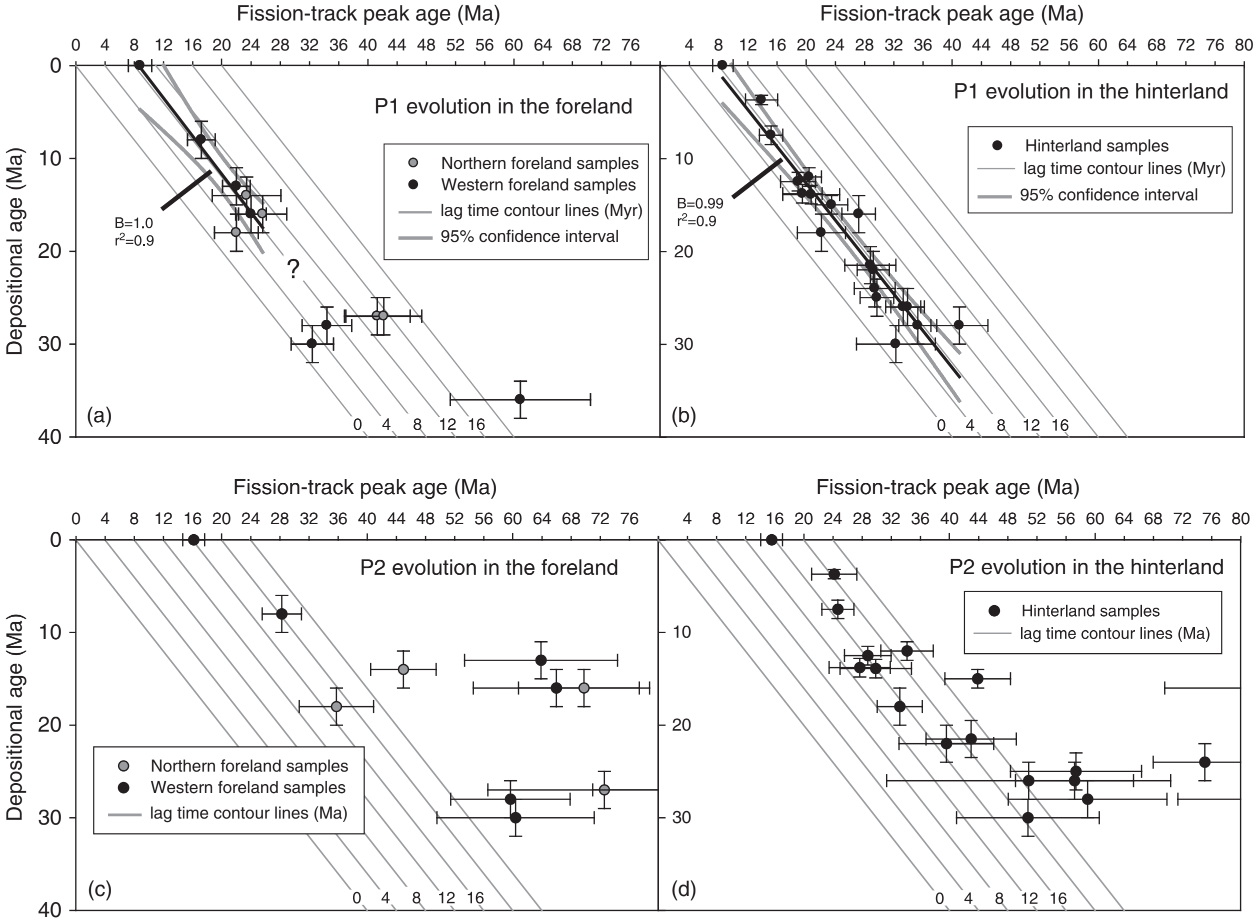


Fig. 7. Evolution of the youngest grain-age components (P1 and P2) throughout the stratigraphic record in the western and northern foreland basin, and the hinterland basin. Contour lines indicate lag times (peak age – depositional age). Both, foreland and hinterland display a fairly steady P1 and P2 evolution since the Late Oligocene/Early Miocene, with relatively constant lag times.

(for example P4) of other samples (see Table 1 for examples). This is a result of labelling peaks with respect to the individual sample, and was done to emphasize up-section changes in the two younger age components P1 and P2. In studies where the detrital zircon FT grain-age distributions and peak ages from samples of the same depositional age are compared P1, P2, etc., are assigned with respect to peak-age groups of all samples to allow for better comparison (Bernet *et al*., 2004a). The different peak ages for the foreland and hinterland samples given in Tables 1 and 2 can now be used to gain provenance information (Brandon & Vance, 1992; Garver *et al*., 1999; Bernet *et al*., 2004a; Stewart & Brandon, 2004), or for calculating lag times (e.g. Bernet & Garver, 2005).

THE MEAN EROSION RATE AND LAG TIME CONCEPTS

Relation between mean age and mean erosion rate

Even if mean ages of detrital FT samples are generally regarded as meaningless, they can provide a first-order estimate of a mean erosion rate of a drainage area. Considering a landscape with spatially varying erosion rates $\dot{\varepsilon}(x, y)$, where x and y indicate position within the landscape covering an area A. In this scenario, the detrital cooling ages in sediment eroded from the landscape are mainly determined by erosion. In other words, the influence of tectonic exhumation on the grain age distribution is insignificant.

It can be argued that given the expected variation in erosion rates, the closure depth Z_c can be taken to be approximately constant. If correct, it can be stated that

$$\tau(x, y) \approx Z_c/\dot{\varepsilon}(x, y) \quad (1)$$

where $\dot{\varepsilon}(x, y)$ now indicates the long-term average erosion rate at x, y averaged over the time interval indicated by τ. We observe the grain ages as a function of their yield in the sediment. Thus, the mean of the measured grain ages $\bar{\tau}$ in a sediment sample is given by

$$\bar{\tau} = Y^{-1} \int_A \tau(x, y)\, \dot{\varepsilon}(x, y)\, \mathrm{d}A \quad (2)$$

where the sediment yield Y from A is given by

$$Y = \int_A \dot{\varepsilon}(x, y)\, dA \qquad (3)$$

The mean erosion rate from the drain is defined by

$$\bar{\dot{\varepsilon}} = Y A^{-1} \qquad (4)$$

Combining Eqns (1)–(4) gives

$$\bar{\dot{\varepsilon}} \approx Z_c / \bar{\tau} \qquad (5)$$

This relationship is directly analogous to that used to relate cosmogenic isotope production to the mean erosion rate in the drainage (Brown *et al.*, 1995; Granger *et al.*, 1996). The sediment eroded from the drainage has a 'mean concentration' that changes approximately proportional with time, as represented by a mean of the grain cooling ages $\bar{\tau}$ or the mean of the cosmogenic isotope concentration. The production process has a vertical length scale over which the 'concentration' increases, which is the average closure depth Z_c or the attenuation length scale for the cosmogenic method. Thus, we can consider that the mean of a distribution of detrital cooling ages is approximately proportional to the inverse of the mean erosion rate. When plotting mean ages of each sample on a lag-time plot, more or less steady trends starting in the foreland basin around 18 and around 28 Ma in the hinterland (Fig. 6). The mean age lag times translate into average erosion rates of 0.3–0.2 km Myr^{-1} during these steady periods.

Lag time concept

The lag-time concept, defined as the cooling age minus the depositional age, can be used to determine exhumation rates from cooling ages. This concept has been described in detail by Garver *et al.* (1999) and is reviewed here (Fig. 8). Three requirements have to be met to apply the lag-time concept when using the FT method: (1) the detrital grains should not have been partially or fully reset after deposition due to a subsequent heating event or deep burial; (2) it is important to know the depositional age of the detrital grains that are dated, at least within the range of precision of their determined cooling ages; and (3) it is of advantage if no active volcanism is present during the time of deposition, which could lead to input of very young volcanic zircons that are not related to exhumational cooling and might obscure the orogenic signal. If these requirements are fulfilled, as they are through most of the history in case of the European Alps, where detrital zircons have not been reset after deposition in synorogenic sediments, the depositional ages are fairly well constrained, and basically no active volcanism occurred since

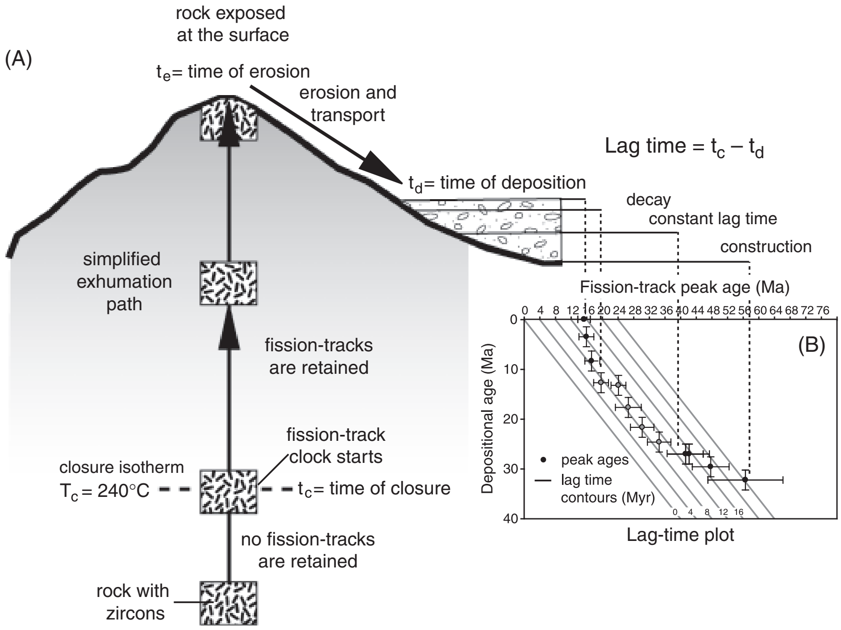


Fig. 8. (a) The basic idea of the lag-time concept is that a rock is exhumed from deeper crustal levels because of erosion, normal faulting or ductile thinning at or near the surface. On its way upwards the rock cools below the closure temperatures of the zircon fission-track (FT) system. As soon as this happens the FT clock is started and continues to tick on the way to the surface. Eventually the rock will reach the surface where it is subject to erosion. The exposed rock will be weathered and eroded and its apatite, zircon and mica grains are released into sediment and transported by glaciers and rivers into the adjacent basins, where they are deposited. The time for erosion and sediment transport is regarded as geologically instantaneous (Heller *et al.*, 1992; Bernet *et al.*, 2004a). Therefore, lag time integrates between the time of closure and the time of deposition and mainly represents the time needed to exhume the rock to the surface. After deposition the FT clock will continue to record the time since cooling below the closure temperature until the grains are heated again close to or above the closure temperature, which would lead to partial or full resetting. (b) The up-section trend in lag time of different grain-age components or peaks (here only the youngest peaks are shown) are important for interpreting the exhumational history of the mountain belt.

Oligocene times (Blanckenburg & Davies, 1995), the lag-time concept can be applied.

DISCUSSION

Provenance of individual peaks

A large number of provenance studies were done in the Alps over the years (e.g. Schlanke, 1974; von Eynatten *et al*., 1996, 1999; Schlunegger *et al*., 1998 and references therein; Garzanti & Malusà, 2008). FT data of detrital zircon can be used for provenance analysis as well (e.g. Hurford *et al*., 1984; Hurford & Carter, 1991; Carter, 1999). That is, peak ages of all samples in Tables 1 and 2 and their associated lag times can be used to identify their potential sources in the European Alps by comparing them to present bedrock FT ages (Fig. 2). Zircons in a FT grain-age peak can have been derived only from areas where the present bedrock zircon FT ages are younger or equal to the age of the peak. This relation allows isolating candidate source areas.

In general, peak ages between 8 and 36 Ma are mainly derived from deeply exhumed rocks of the Central and Western Alps, and the external massifs, or the Tauern Window in the Eastern Alps (Fig. 2). Input of volcanic zircon with cooling ages of $\sim$30 Ma from Oligocene volcanism is insignificant. In Fig. 9, a comparison of FT cooling ages of euhedral and rounded zircons is shown for three samples that were deposited between 30 and 16 Ma in the hinterland basin. Because volcanic zircons have in general an euhedral grain shape, we would expect to find a strong shape-age relation of euhedral grains with only 30 Ma cooling ages, if volcanic input is significant. While each sample contains euhedral grains with 30 Ma cooling ages, younger and older euhedral grains were observed as well as rounded grains with the same age spectrum. Therefore, a strong shape-age relation is not confirmed, which means that not all euhedral grains with 30 Ma cooling ages were necessarily derived from Oligocene volcanism (also see Dunkl *et al*., 2001). Combined FT and U/Pb dating on the same grains is needed to identify grains of Oligocene volcanic origin (e.g. Carter & Moss, 1999), and should be the objective of future research.

Peak ages between 40 and 50 Ma are likely derived from Austroalpine units that were eroded off the Central Alps during the Oligocene, but still occur today in parts of the Eastern Alps (Kuhlemann *et al*., 1999). This result is supported by other provenance studies using white mica Ar–Ar dating (von Eynatten *et al*., 1996, 1999), or the work by Schlanke (1974) and Schlunegger *et al*. (1998). All older peaks of $\sim$60–140 Ma or older are related to material from the less deeply exhumed units of the Alps, including possible recycling from partially or non-reset cover units and synorogenic sediment (Fig. 2, Bernet *et al*., 2004a).

Long-term exhumation

The evolution of P1 and P2 lag times in the foreland and hinterland deposits through time provides insight into the long-term exhumational history of the Alps (Fig. 7).

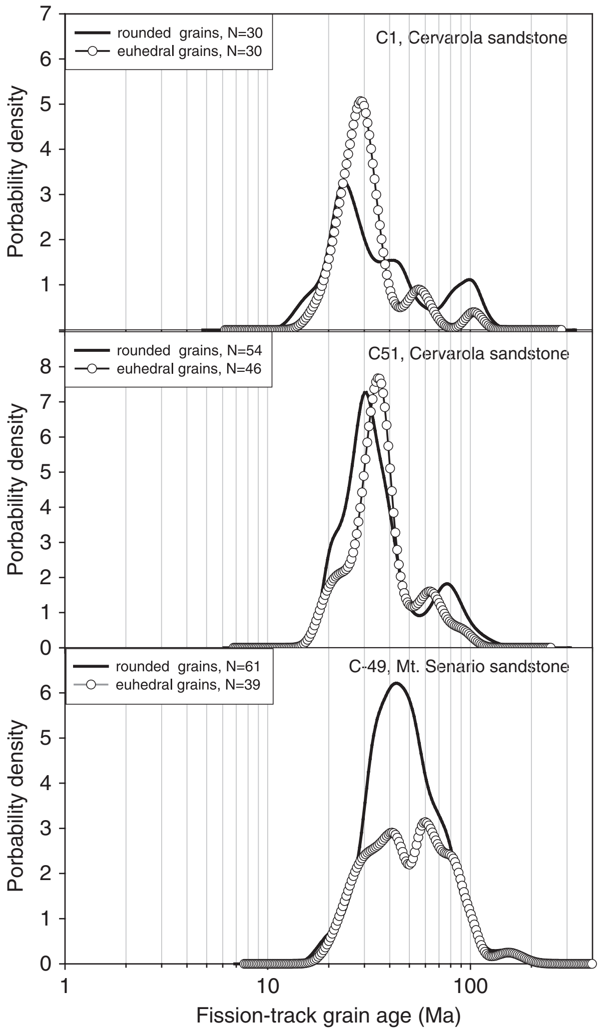


Fig. 9. Probability density plot of one M. Senario and two Cervarola samples, comparing zircon grain shapes (euhedral vs. rounded) with fission-track (FT) cooling ages. There is no significant correlation between grain shape and cooling age. Both grain shapes represent the whole age range in the observed grain age distribution. Therefore, euhedral grains cannot be directly related to volcanic input, but could also be derived from exhumationally cooled meta-sedimentary units or granites (also see Dunkl *et al*., 2001).

Zircon FT data from modern-river samples of the Rhône River (Bernet *et al*., 2004a) and the Ticino River (Bernet *et al*., 2001) were added in Fig. 7 for a more complete data set.

Overall, P1 lag times show a fairly stable trend since the mid-Miocene ($\sim$18 Ma) in the foreland samples and since the Oligocene ($\sim$28 Ma) in the hinterland samples (Fig. 7a and b). Such constant lag-time trends are evidence for zircon with consistently young cooling ages being shed to both sides of the orogen. This result can be interpreted in two ways. The first option is that certain source areas were rapidly exhumed at a specific time, providing zircons with short lag times, but than fast exhumation changed to a different area from where young zircons with short lag times were supplied to the basins. The second, and here preferred interpretation of the data is that in some parts of the Central and maybe even the Western Alps zircons have cooled in a fairly steady fashion. Willett & Brandon (2002) defined such a thermochronologic exhumational steady state as the time invariant production of cooling ages within a specified spatial domain. In other words, zircon sources existed which were continuously exhumed at fairly constant rates. The average lag time of P1 for the Oligocene or mid-Miocene to present exhumation periods in both the foreland and hinterland is $\sim 7.4 \pm 1.3$ Myr. This average lag time can be translated into a long-term average exhumation rate of $\sim 0.7 \pm 0.1\,\mathrm{km\,Myr^{-1}}$, using a one-dimensional (1D) advective thermal model (AgeToEdot programme of Brandon, as described in Ehlers *et al.*, 2005), assuming an initial average geothermal gradient of $\sim 30\,^{\circ}\mathrm{C\,km^{-1}}$ and a zircon FT closure temperature of 240 °C for the Central Alps (Hurford, 1986). Such exhumation rates must have been established earlier than documented by the first occurrence of zircon with a ~ 8 Myr lag time, because to see young cooling ages in the sedimentary record, the non- or partially reset cover-units need to be removed first. One obvious question is how stable was the thermal structure of the Alps since the Eocene, on the scale of the whole orogen, to justify this interpretation? The thermal structure could be perturbed by relief change and growing topography, fast exhumation or a thermal event such as widespread magmatism. When discussing the exhumation in terms of apatite FT thermochronology it is necessary considering the influence of exhumation rates and topography on the closure depth (Stüwe *et al.*, 1994; Braun *et al.*, 2006). For the zircon FT system, however, the closure depth seems to remain fairly uniform over a range of exhumation rates, up to 1.5–2 $\mathrm{km\,Myr^{-1}}$. Given the exhumation rates calculated here the closure depth was probably not much perturbed and stayed at around 7–8 km depth. The closure isotherm is also deep enough so that it is not significantly affected by Alpine topography. More important could be the thermal perturbation of the Alps by magmatic activity following slab-break off during the Oligocene. One can argue that the cooling recorded in zircon FT ages is at least to some part the result of thermal relaxation following perdiadriatic magmatism, and not related to exhumational cooling. This argument is likely to be true for zircons derived directly from periadriatic intrusions, and may be true for zircons with Oligocene FT ages from areas close to the intrusions (contact aureoles). However, on the basis of the available detrital and bedrock FT data, it cannot be distinguished how much cooling of such zircons was post-magmatic and/or exhumational. Post-magmatic cooling seems not applicable to most zircons with Miocene or younger cooling ages. At least so far, the thermal overprint related to slab break-off and periadriatic magmatism remains poorly constraint and is only locally documented in parts of the Southern and the internal Western Alps (Desmons *et al.*, 1999). The Lepontine dome of the Central Alps, the most important P1 source area was undoubtedly affected by post-metamorphic cooling related to exhumation (Hunziker *et al.*, 1989; Frey & Mählman, 1999, and references therein) rather than by post-magmatic thermal relaxation. The external massifs, which are important sources of zircons with Alpine cooling ages, seem to be little or non-effected at all by the Oligocene periadriatic thermal event (Frey *et al.*, 1999, and papers therein). Therefore, this thermal event seems to have little influence on the exhumation signal of the Alps derived from the sedimentary record.

Why is a longer constant P1 trend observed in the hinterland basin than in the foreland basin (Fig. 7)? The obvious answer is that sample coverage is more complete in the hinterland basin, but more important is that the hinterland samples are exclusively collected from turbidite sequences which have the advantage of being better mixed and more likely to contain all major FT grain-age components exposed in the source area on a regional scale. Samples collected closer to the source reflect a more local signal, maybe only from a single drainage area, depending on sample location, which is not necessarily representative of the FT ages of the whole orogenic system (Bernet *et al.*, 2004a).

P2 lag-time trends are more variable and show more scatter than the P1 data, especially in the foreland basin samples (Fig. 7). The onset of relatively constant lag times for P2 in the hinterland was around 22 Ma, which is consistent with the P1 lag time. Sources areas for P2 zircons of <36 Ma can be located in the Central and Western Alps today (Fig. 2), and are not directly affected by exhumation due to normal faulting, but only by erosion. They have average long-term exhumation rates of ~ 0.4–$0.1\,\mathrm{km\,Myr^{-1}}$, based on the lag times presented in this study. These values correspond to the long-term exhumation rate estimates of Grasemann & Mancktelow (1993) for the Central Alps before the onset of the Simplon Fault and subsequently in the hanging wall of the fault. Kuhlemann *et al.* (2001) determined that 88% of the exhumation of the entire Alps since the Oligocene was controlled by erosion. Grains of the older P2 peaks (>50 Ma in Tables 1 and 2) are possibly derived from only partially reset cover units, similar to the partially or non-rest zircons that make up P3 and P4 in all the samples.

The fraction of grains with Alpine cooling ages (<36 Ma) increases over time (Fig. 5), which could mean that the area of exposed bedrock that contains zircon with Alpine cooling ages may have increased over time, at least relatively with respect to areas with pre-alpine cooling ages, because exhumation rates have remained fairly constant. With the available data, it is not evident how this relates to a possible change in size of the mountain belt as suggested by others. For example Schlunegger & Willett (1999) proposed growth of the Alps at the end of the Oligocene/early Miocene on the basis of thermochronologic data and dynamic modelling, and Cederbom *et al.* (2004) postulated a narrowing of the orogenic system from uplift and recycling of foreland basin sediments to the north of the Alps since the Pliocene.

Exhumation rates vs. sediment yield

Exhumation rate estimates for the fastest exhuming areas of 0.4–0.7 $\mathrm{km\,Myr^{-1}}$, and average erosion rate estimates of

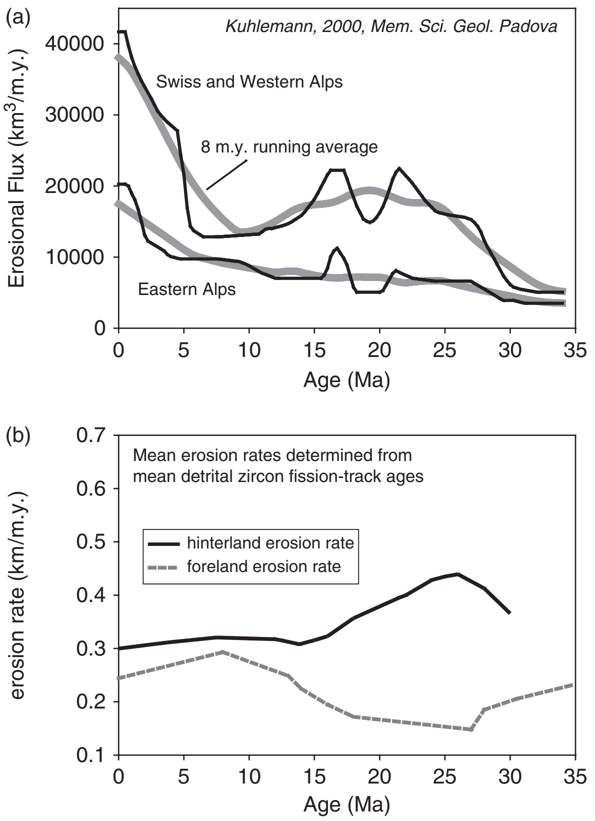


Fig. 10. Comparison of sediment yield data of the Alps (a) (after Kuhlemann *et al.*, 2001) with smoothed erosion rate estimates derived from central ages of detrital zircon FTGA distributions of the foreland and hinterland (b). Note that no increase in erosion rates is seen in the Pliocene, as would be expected by the proposed increase in sediment flux since that time.

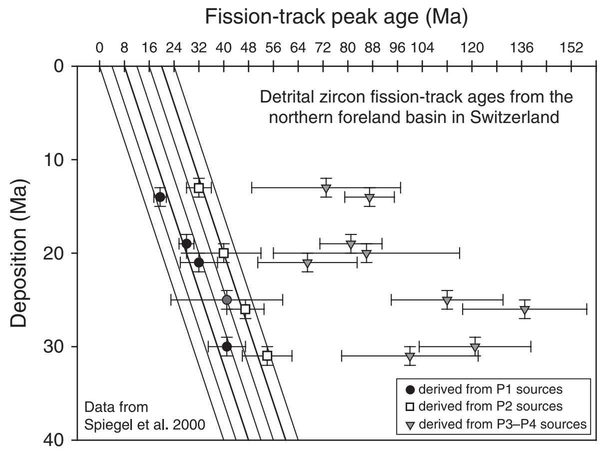


Fig. 11. Lag-time plot of detrital zircon FT data collected from alluvial fans in the Swiss Molasse basin (data from Spiegel *et al.*, 2000).

0.2–0.3 km Myr^{-1} derived from mean age lag times, are comparable with exhumation rate and overall erosion rate estimates from other studies (e.g. Clark & Jäger, 1969; Hinderer, 1999, 2001; Kuhlemann *et al.*, 2001; Schlunegger *et al.*, 2001). At first glance the constant lag times and constant exhumation rates seem to be in conflict with the considerable variation in sediment yield that Kuhlemann (2000) and Kuhlemann *et al.* (2001, 2002) determined by detailed sediment budget calculations for the Central, Western and Eastern Alps over the same period of time (Fig. 10). However, sediment yield can vary without actually changing exhumation rates, but because of changes in erodibility of exposed lithology (Schlunegger & Willett, 1999; Kühni & Pfiffner, 2001; Schlunegger *et al.*, 2001; Kuhlemann *et al.*, 2002) or changes in drainage size and/or a shift in the drainage divide. Kuhlemann (2000) and Kuhlemann *et al.* (2001, 2002) postulated a steep increase in erosion rates from increasing sediment yield starting from about 5 Ma onwards. Even though an increase in erosion rates is also reflected in bedrock apatite FT data form the Central and Western Alps (Vernon *et al.*, 2008), it is not detected in the detrital zircon FT data presented here. One explanation for this outcome is that the increasing sediment yield is not derived form the P1 and P2 source areas alone, but is coming also from recycling of foreland basin sediments. This at least would be expected for a reduction in the width of the Alps, which would cause uplift at the flanks of the range, as argued by Cederbom *et al.* (2004). These sediments would have a mix of detrital zircon ages and thus would not show well an increase in erosion rate. Therefore, the increase in sediment yield did not come only from the P1 areas given the steady lag times. If the entire Alps, including the P1 and P2 source areas, would be eroded much faster since the Pliocene (>1.4 km Myr^{-1}), because of climate change, as proposed by Cederbom *et al.* (2004), or for other reasons, then the Alps should have a much deeper relief today. In addition, it is surprising that zircons with Pliocene cooling ages are not detectable in modern river samples (Bernet *et al.*, 2004a, b). It seems that the zircon FT data are not sensitive enough to pick up this relatively recent increase in erosion rates and sediment yield. Therefore, while the Alps are being eroded more rapidly over the past 5 Myr, erosion was not efficient enough to remove sufficient rock to widely expose zircons with Pliocene or younger cooling ages.

Comparison with other detrital thermochronologic studies in the Alps

Spiegel *et al.* (2000, 2002, 2004) presented detrital zircon FT data from Oligocene to Miocene clastic alluvial fan deposits in the Swiss part of the northern foreland basin. Zircon FT data of Spiegel *et al.* (2000, 2002, 2004) from the Honegg-Napf fan (30–14 Ma depositional age) for example show a similar trend as our results presented here, with an average P1 lag time of $\sim$8.5 Myr between 21 and 14 Ma (Fig. 11). However, samples collected from alluvial fans farther to the east (Rigi-Höhrone fan, Hörnli fan, etc., see Fig. 1 in Spiegel *et al.*, 2000), have an average lag time of the youngest age component of $\sim$19.8 Myr between 31 and 13 Ma, which is much closer to the average P2 lag time of 16 Myr of this study. Spiegel *et al.* (2000, 2002) interpreted this as a result of episodic exhumation, but it is possible that the eastern alluvial fans just did not receive 'P1' zircons, and apparently only sampled 'P2' areas in the Central Alps. Given the proximity of the alluvial fans to their source area this is a feasible explanation. The closer the location is selected to the source, the more localized is the information and it is possible to miss the areas that experience the fastest exhumation. The local variation in exhumation of the Alps has been shown with a series of samples from modern rivers that drain the orogen to its foreland and hinterland today (Bernet *et al.*, 2004a, b), and it is more than likely that the same spatial variation in the exhumation signal existed in the past.

The detrital zircon FT data from the hinterland basin presented here confirm the results of a study on sediment from the Macingo formation in the northern Apennines by Dunkl *et al.* (2001). These authors found the same zircon FT cooling-age clusters of Palaeogene age (31–34 Ma), Late Cretaceous age (60–70 Ma) and Jurassic age (140–160 Ma). Dunkl *et al.* (2001) also found the same relation between euhedral and rounded grains as we found in our samples from the hinterland basin. For that reason, FT results of detrital zircon are very robust and results from different FT laboratories are compatible.

FT analysis is not the only method of detrital thermochronology applied to study provenance and exhumation in the Alps, but also $^{40}Ar/^{39}Ar$ analysis on detrital mica is widely applied (e.g. von Eynatten *et al.*, 1996, 1999; von Eynatten & Wijbrans, 2003; Morag *et al.*, 2008). An example of a regional episodic exhumation event in the Western Alps was clearly demonstrated by Carrapa *et al.* (2003) with a static peak at $\sim$38 Ma of $^{40}Ar/^{39}Ar$ ages of detrital mica collected from the Tertiary Piedmont basin (TPB) in northern Italy, to the east of the Western Alps. The static 38 Ma $^{40}Ar/^{39}Ar$ cooling age, which can be found throughout the stratigraphic record of the TPB, marks the end of rapid exhumation that caused thick layers of crustal rock to be quickly cooled below 350–450 °C, the closure temperature of the $^{40}Ar/^{39}Ar$ system for white mica (von Eynatten & Wijbrans, 2003), when these rocks reached mid-crustal levels. This exhumation event was driven by erosion in the southern part of the Western Alps and the Ligurian Alps (Carrapa *et al.*, 2004), coeval with the end of HP-metamorphism of the Dora Maira massif in the Western Alps (e.g. Avigad *et al.*, 1993). Zircon FT cooling ages of 20–30 Ma from bedrock exposed internal Western Alps today (see data repository of Bernet *et al.*, 2001; Schwartz *et al.*, 2007; Vernon *et al.*, 2008), indicate that cooling and exhumation considerably slowed down after the rocks had reached mid-crustal levels, and they were continuously exhumed at rates of 0.2–0.3 km Myr^{-1} ever since. Unfortunately, the Tertiary Piedmont basin does not retain the exhumational history of the Central Alps or the northern part of the Western Alps because its source lay exclusively in the southern part of the Western Alps and the Ligurian Alps, as shown by Carrapa *et al.* (2004). Therefore, fast continuous exhumation of zircon in the Central Alps since the Oligocene cannot be detected in the sedimentary record of the TPB. This exhumational signal by-passed the TPB and was preserved in the turbidite sequences of

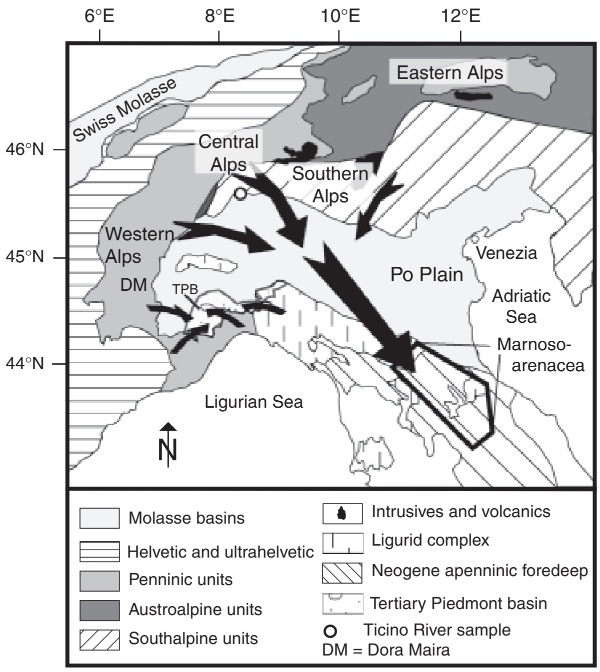


Fig. 12. Simplified geologic map of the Alps indicating the location of the Marnoso-arenacea Formation in the Northern Apennines (sample location of Bernet *et al.*, 2001 and this paper), and the Tertiary Piedmont basin (sample location of Carrapa *et al.*, 2003) Black arrows indicate sediment provenance of each basin.

that make up part of the Apennines today (Fig. 12, Garzanti & Malusà, 2008).

Dynamic modelling, GPS and seismic studies

Dynamic numeric modelling of the Central Alps by Schlunegger & Willett (1999) and suggested that exhumation of the metamorphic core of the Central Alps (Lepontine dome, Fig. 1) was variable, with higher rates before 20 Ma. Although their study is consistent with very rapid tectonic exhumation of the Lepontine dome around 20 Ma, they cannot rule out higher rates of erosional exhumation over a longer period of time. Therefore, Schlunegger & Willett (1999) concluded that total exhumation was driven by a combination of extensional tectonics and erosion, which is consistent with the view presented in this study. Schlunegger & Willett (1999) also proposed a balance between erosion and convergence mass flux for the early stages of Alpine evolution, but postulated that erosion rates and overall exhumation rates may have decreased since the Early Miocene, maybe because of a change in climate. These authors argued that erosion failed to keep up with mass accretion by convergence, causing broadening of the width of the orogen to accommodate surplus material. The conclusions drawn by Schlunegger & Willett (1999) state that the area of fast exhumation may have been reduced, but is still large enough to produce detectable amounts of relatively fast cooled zircons in the stratigraphic record. However, the data of this study show that the contribution of zircon with synorogenic cooling ages has increased over time (Fig. 5), whereas the exhumation rate remained constant. A newer study by Willett *et al.* (2006) argued that erosion rates in the Alps have significantly increased since the Messinian Salinity Crisis ($\sim$5.6 Ma) as a result of baselevel drop in the hinterland and erosion of the Southern Alps, and extensive erosion of the Northern Foreland basin at the same time. Both, erosion of the Southern Alps, which still expose mainly rocks with Jurassic cooling ages or older, and recycling of the Northern Foreland basin will contribute to the increase in sediment yield that has been postulated by Kuhlemann (2000), but it does not mean that long-term average exhumation rates have changed so far in the Central Alps or that therefore the zircons with young Alpine cooling ages in the metamorphic core of the Central Alps would be exhumed differently than before. Kühni & Pfiffner (2001) used numeric modelling to predict the topographic evolution of the Central Alps and to detect shifts in drainage divides. Their findings are not in contradiction with our results, confirming that the Alps have been a fairly symmetric mountain belt with equal amounts of orographic precipitation and erosion on both sides of the main drainage divide since at least the middle Miocene.

Naylor & Sinclair (2007) and Hoth *et al.* (2007) highlighted the punctuated deformation of thrust sheets during frontal accretion in doubly vergent orogenic wedges such as the European Alps, using numeric and analogue modelling respectively. Frontal accretion and thrust deformation clearly influence localized surface uplift and exhumation which in turn has an impact on the cooling history of the crustal rocks that will eventually be exposed at the surface. Their models predict that rates of surface uplift, frontal accretion and exhumation are punctuated on times scales of 0.1–5 Myr. This time scale of punctuation is at least 30% shorter than the long-term average lag time of the P1 zircons in the Alpine orogenic sediments. Therefore, such punctuations may not easily be detected in the detrital zircon FT record. Obviously further research is needed to better integrate detrital zircon FT results into dynamic modelling and to find a consensus in the interpretation of the two different approaches. In addition, the data presented in this study cannot prove or disprove faster exhumation rates during the constructional phase of the Alps throughout the Eocene and most of the Oligocene, because the non-reset cover units were removed from large parts of the Western and Central Alps during these times, providing only old FT cooling ages to the sedimentary record. Such zircons do not confer much information on Alpine exhumation rates except that they indicate the removal of partially or non-reset cover units.

Current GPS (Sue *et al.*, 2000; Calais *et al.*, 2002) or seismic studies in the Western Alps (e.g. Eva *et al.*, 1997; Sue *et al.*, 1999) provide a mixed picture on the activity of the Alps today. The Western Alps may be inactive by now, which should lead to an increase of lag time in the long run and draw down of the orogen as long as erosion rates remain relatively high. Selverstone (2005) argued that the Western Alps may be on the verge of collapsing but that the same is not true for the Central and Eastern Alps. Overall a decay of the Alps is not seen in the detrital zircon FT record yet (also see Bernet *et al.*, 2004a, b).

CONCLUSIONS

Detrital zircon FT results of 24 sandstone samples and previously published data from stratigraphically controlled sections of the Alpine foreland and hinterland were used to determine sediment provenance and long-term average exhumation rates of the European Alps. Individual peak ages can be related to sources in the Central, Western, Eastern and Southern Alps. Relatively constant lag times of the two youngest peaks in each sample from the foreland and hinterland indicate continuous exhumation of zircon with young cooling ages in the fastest exhuming parts of the Central and Western Alps by a combination of erosion and normal faulting since the Late Oligocene/Early Miocene. It is remarkable that the long-term exhumational signals preserved in the western and northern foreland basins and the hinterland basin are similar through time.

The data presented in this paper are in good agreement with other detrital zircon FT studies in the Alps. However, so far neither an increase in erosion rates since the Pliocene is detected, nor a decrease of exhumation due to the lack of convergence, as suggested by current GPS and seismic data. The detrital zircon FT data presented here do not provide any indication of major long-term changes of tectonic or climatic forcing during the evolution of the Alps since continental collision, and they seem to be insensitive to fluctuations on time scales of <5 Myr. Future work is needed to find a better consensus between detrital thermochronologic data, dynamic modelling, and sediment budget estimations.

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