Tectonic denudation of a Late Cretaceous–Tertiary collisional belt: regionally symmetric cooling patterns and their relation to extensional faults in the Anatolide belt of western Turkey

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Abstract – Thermochronological data reveal that the Late Cretaceous–Tertiary nappé pile of the Anatolide belt of western Turkey displays a two-stage cooling history. Three crustal segments differing in structure and cooling history have been identified. The Central Menderes metamorphic core complex represents an ‘inner’ axial segment of the Anatolide belt and exposes the lowest structural levels of the nappé pile, whereas the two ‘outer’ submassifs, the Gözdes submassif to the north and the Cine submassif to the south, represent higher levels of the nappé pile. A regionally significant phase of cooling in the Late Oligocene and Early Miocene affected the outer two submassifs and the upper structural levels of the Central Menderes metamorphic core complex. In the northern part of the Gözdes submassif, cooling was related to top-to-the-NNE movement on the Simav detachment, as the apatite fission-track ages show a northward-younging trend in the direction of movement on this detachment. In the Cine submassif, relatively rapid cooling in Late Oligocene and Early Miocene times may have been related to top-to-the-S extensional reactivation of the basal thrust of the overlying Lycaon nappes. The second phase of cooling in the Anatolide belt is related to Pliocene to Recent extension resulting in the formation of the Central Menderes metamorphic core complex in the inner part of the Anatolide belt. Core-complex development caused the formation of supra-detachment grabens, which document the ongoing separation of the Central Menderes metamorphic core complex from the outer submassifs.

Keywords: fission-track dating, exhumation, extension, Anatolides, Turkey.

1. Introduction

The Anatolide belt of western Turkey (also referred to as the Menderes Massif) and the adjacent Aegean Sea underwent regional extensional deformation from at least Oligocene times to the present (Lister, Banga & Fecesratr, 1984; Buick, 1991; Seyiğil, Scott & Rundle, 1992; Heltzel et al. 1995a; Gessner et al. 2001a). In the Aegean, exhumation is controlled by southward retreat of the subducting Hellenic slab, and displacement on major extensional detachments is of the order of 50 to more than 100 km (Lister & Fodor, 1996; Ring, Lavier & Reischmann, 2001; Ring, Brachert & Fassoulas, 2001). In contrast, no large-magnitude extensional detachments have been mapped so far in the Anatolide belt of western Turkey and exhumation is apparently not controlled by slab retreat (Heltzel et al. 1995b; Ring et al. 1999a). While both the Anatolide belt and the Aegean are characterized by southward propagation of crustal shortening, the geometric pattern of extensional faults and the overall magnitude of extension differ fundamentally in both areas.

In this article we report on the results of a reconnaissance fission-track study across the Anatolide belt of western Turkey. Fission-track thermochronology is a powerful tool for establishing regional variations in the cooling history of an orogen (Wagner & Reiner, 1972; Johnson, Harbury & Huford, 1997; Foster & John, 1999). Combining regional cooling patterns with structural analysis can deliver important constraints to the exhumation history of an orogen (Huford, Flisca & Jager, 1989; Wheeler & Butler, 1994; Brandon, Roden-Tice & Garver, 1998). We will show that the regional pattern of apatite fission-track ages is remarkably symmetric at the orogen scale, with the oldest (Late Oligocene to Early Miocene) cooling ages occurring in the southern and northern margins and younger (Late Miocene to Quaternary) cooling ages in the central part of the belt. When fission-track data are combined with 40Ar/39Ar-mica ages, two distinct periods of normal faulting are resolved, which have exhumed the Anatolide belt from mid-crustal depths since Late Oligocene times.
2. Tectonic setting

2.a. Hellenide-Anatolide orogen

The Late Cretaceous-Tertiary Hellenide-Anatolide orogen in the eastern Mediterranean shows significant along-strike variations in basement and structure (Fig. 1). Single-cumulate 207Pb/206Pb Pb-U-Pb Sm-Pb SHRIMP dating of magmatic zircons yielded Carboniferous protolith ages of ~300-320 Ma for large parts of the pre-Alpine basement of the Hellenide belt (S. Key, unpubl. Ph.D. thesis, Australian Nat. Univ., 1998; Reischmann, 1998; Engel & Reischmann, 1998; Ring, Laws & Bernet, 1999; Reischmann et al., 2001). In contrast, the pre-Alpine basement in the structurally deepest tektontomamorphic unit of the Anatolide belt in western Turkey, the Menderes nappes, yielded ages of ~530-560 Ma, and Carboniferous ages are conspicuously absent (Hetzel & Reischmann, 1996; C. Danna, unpubl. Ph.D. thesis, Univ. Mainz, 1997; Hetzel et al., 1998; Loos & Reischmann, 1999; Gesner et al., 2001). Therefore, Ring et al. (1999) and Gesner et al. (2001) regarded the Menderes nappes as part of the exotic Anatolide continent, or Menderes-Tauride block of Robertson & Dixon (1984), in the Hellenide-Anatolide orogen.

Late Cretaceous to Tertiary subduction and accretion processes in both regions were also different. In the Aegean, subduction, associated high-pressure metamorphism and the formation of southward-propagating magmatic arcs and back-arc basins took place from Late Cretaceous times until the present (Fytikas et al., 1984; Ring & Layter, 2003). In western Turkey, subduction and associated high-pressure metamorphism only occurred from Late Cretaceous to Eocene times (Gesner et al., 2001). It has been proposed that the coalescence of the Anatolian continent in the Eocene halted subduction and the development of magmatic arcs and back-arc basins in western Turkey (Hetzel et al., 1995b; Collins & Robertson, 1998; Ring et al., 1999a).

The Anatolide belt in western Turkey comprises three major tektontomamorphic units (Figs 1, 2), which are characterized by different lithologies, protolith ages, pre-collisional palaeogeographic and orogenic history (Ring et al., 1999a). This heterogeneous tektontic pile was assembled during Eocene collision-related crustal shortening. The Lycian nappes and the Izmit-Ankara suture zone represent the uppermost tektontomamorphic unit, which was deformed and, at least in part, metamorphosed under high-pressure conditions during Late Cretaceous times (before ~75 Ma) (Collins & Robertson, 1997; Sherlock et al., 1999; Oberhümel et al., 2001; Ring & Layter, 2003). The middle tektontomamorphic unit is correlative with the Cycladic blueschist unit in the Aegean (Candan et al., 1997; Ring et al., 1999a). In the Eocene, the Cycladic blueschist unit was emplaced onto the Menderes nappes along the Cyclades-Menderes thrust (Fig. 1) during the greenhouse-facies D3 event. Because the D3 emplacement of Late Cretaceous-Tertiary high-pressure units onto the Menderes nappes did not cause a high-pressure overprint in the latter, the upper and middle units must have been already exhumed to mid-crustal levels by Eocene times (Ring et al., 1999a; Gesner et al., 2001c).

2.b. The Menderes nappes

Samples for this study were collected from the Menderes nappes, which form the structurally lowest tektontomamorphic unit of the Anatolide belt. The Menderes nappes comprise from top to bottom: (1) the Selimiye nappe, (2) the Çine nappe, (3) the Bozdag nappes, and (4) the Bayındır nappe. The Çine and Bozdag nappes have a polyorogenic history, which ended back into the Neoproterozoic-Cambrian (Fessner et al., 2001b; Candan et al., 2001; Ring, Wiffner & Lackmann, 2001).

The Selimiye nappe contains Palaeozoic metapelite, metabasite and marble (Schulzing, 1962; Çağlayan et al., 1960; Loos & Reischmann, 1999; Régnier et al., 2003). Most of the Çine nappe consists of deformed orthogneiss, largely undeformed metagranite and minor pelitic gneiss, eclogite and amphibolite. Protoliths of much of the orthogneiss/metagranite intruded at ~530-560 Ma (see previous Section). The underlying Bozdag nappes is made up of metapelitic with intercalated amphibolite, eclogite and marble lenses. Protoliths ages of all rock types of the Bozdag nappes are unknown, but geological constraints (Candan et al., 2001; Gesner et al., 2001b) suggest a Precambrian age for at least parts of these rocks. The Bozdag nappes was intruded by granodiorites at 230-240 Ma (C. Danna, unpubl. Ph.D. thesis, Univ. Mainz, 1997). The Bayındır nappes contains phyllite, quartzite, marble and granites of inferred Hercynian-Carboniferous and possibly Mesozoic age (O. Candan, pers. comm. 1999). The rocks were affected by a single Tertiary greenschist-facies metamorphism during the D3 event (Ring et al., 2001) and suggest an Eocene event for this age. D3 produced a variably spaced mylonitic shear-band foliation and a stretching lineation associated with top-to-the-S kinematic indicators. D3 shear zones define the boundaries between the Selimiye, Çine, Bozdag and Bayındır nappes, including the top-to-the-S Selimiye shear zone, which separates the Selimiye nappes from the underlying Çine nappes. Bauxart & Park (1994) interpreted the Selimiye shear zone as a Late Oligocene top-to-the-S extensional detachment. Hetzel & Reischmann (1996) showed that 40Ar/39Ar muscovite ages of 35-43 Ma constrain slow Eocene cooling during and after movement at the Selimiye shear zone. Subsequently, Ring et al. (1999a), Gesner et al. (2001c) and Régnier et al. (2003) supplied
Figure 1. Simplified tectonic map of Anatolide belt in western Turkey (this region is traditionally referred to as Menderes Massif); subdivision of region into three submassifs (Görges submassif, Keklik submassif, which occupies same area as the Central Menderes metamorphic core complex, and Cine submassif) follows Sevgili (1987). Note that the Görges and Cine submassifs are also referred to as outer submassifs whereas the Central Menderes metamorphic core complex represents inner submassif. Late Tertiary extension in study area is expressed by two symmetrically arranged detachment-fault systems: the Kuzey detachment in the north and Güney detachment in the south, both of which delimit the Central Menderes metamorphic core complex, also shown is Simav detachment (İşık & Tekeli, 2001) in northern part of Görges submassif and postulated Lycian detachment at southern tip of Cine submassif. Profile lines AA’, BB’, CC’, DD’ and EE’ refer to cross-section as Fig. 2. Abbreviations: BMG – Büyük Menderes graben, KG-M – Kıcık Menderes graben, GG – Gediz graben. Inset: Generalized map of the Mediterranean showing Hellenides (grey), Anatolides (black), Cyclades and Crete, location of main map and major detachment faults in Aegean: 1 – Creten detachment, 2 – Ios and Serifos detachments, 3 – Montium, Varsi and Mykonos detachments. Abbreviations: NAFFZ – North Anatolian fault zone; SAFFZ – South Anatolian fault zone.
Figure 2. (a) Cooling ages in relation to position along cross-section (FTT = fissure-track thermochronology); sample numbers correspond to Fig. 3 except those indicated by dagger and double daggers (samples T2, T20, 92T27, 92T54 from Hettel & Reischmann (1996); samples 93T18 and 93T55 from granite in footwall of Kaezy detachment from Hettel et al. (1995a); samples GD and BMD from Lips et al. (2001)). These 40Ar/39Ar muscovite ages from Çine nappe at southern end of profile (dagger, 93T12, 93T13) get younger with structural depth compatible with interpretation that ages are cooling ages. Note that in this diagram only samples from NNE-SSW profile (Fig. 1) are shown. (b) Schematic N-S cross-section showing architecture of nappe pile assembled during Eocene (D2,3) top-to-the-S nappe stacking (for profile lines refer to Fig. 1). Syncline in central part of Antioöide belt is due to large-scale folding during late phase of divergent detachment faulting (Gesner et al. 2001a). For geometric viability, section planes are oriented parallel to mean orientation of Umax, trace of foliation is projected into section plane and used to infer geometry of sub-surface structures.

evidence for the interpretation that the Selimiye shear zone originated as a thrust that fits into the regional picture of Eocene topo-to-the-S D2 thrusts (Collins & Robertson, 1998).

Tertiary metamorphic temperatures are difficult to constrain as Tertiary metamorphism is commonly retrogressive, overprinting pre-existing high-grade metamorphism and migmatization that has been dated at ~540-550 Ma (Damrat & Reischmann, 1999; T. Reischmann pers. comm. 1999; Gesner et al. 2001b). Metapelites at the southern rim of the Çine submassif, which has been assigned a Devonian to Carboniferous age, contains garnet, chloritoid, muscovite, rutile and graphite and records Tertiary metamorphic temperatures of 500–550°C (Whitney & Bozkurt, 2002; Régnier et al. 2003). Triassic granite in the Bozdag nappe south of Alaşehir also contains garnet, R. Hettel (unpub. Ph.D. thesis, Univ. Mainz, 1995) reported garnet and biotite from the Bayındır nappe; however, the growth of these minerals might be due to contact metamorphism related to Early Miocene intrusions of the Turtulu and Salihi granite bodies. Metapelites in the Çine and Bozdag nappes have no Tertiary garnet but Tertiary biotite (Ring, Willner & Lackmann, 2001). The occurrence of biotite in the Çine and Bozdag nappes suggests temperatures > ~400°C, while the absence of garnet suggests that Tertiary metamorphic temperatures in both nappes did not exceed ~450°C (Bucher & Frey, 1994). Overall, this summary suggests that Tertiary metamorphic temperatures in the Menderes nappes were between 400 and 550°C.

2c. Structures related to horizontal extension

In the northern part of the Gördes submassif, İşik & Tekeli (2001) identified the top-to-the-NNE Simav detachment (Fig. 1). The Simav detachment reactivated the Eocene Cyclades–Menderes thrust. Initial movement on the Simav detachment was synchronous with the intrusion of the Eğirgöz and Kayunoba granites, which are considerably stretched in the NNE direction. A SHRIMP-U/Pb-zircon age of 20.7 ± 0.5 Ma dates the intrusion of the Eğirgöz granite and thus, initial movement on the Simav detachment (Ring & Collins, 2003). The pervasive pre-existing foliation in the footwall of the Simav detachment is subhorizontal and parallels the detachment fault. The overall geometry across the Simav detachment resembles those across detachment faults in the nearby Aegean, such as the Mykonos detachment (Lee & Lister, 1992), the Vary detachment (Ring, Thomson & Brückner, 2003) or the Ios detachment (Lister & Forster, 1996), which are thought to have originated as low-angle structures.

Contractional D3 structures were overprinted in the Gördes submassif and in the northern Central Menderes metamorphic core complexes by ductile D2 shear bands.
These shear bands developed in the northern Central Menderes metamorphic core complex simultaneously with the intrusion of the Early Miocene (~20 Ma) Turgutlu and Sahibli granites during NNE-oriented horizontal extension (Hetzel et al. 1995c; Gessner et al. 2001c) (Fig. 3). The occurrence of anastilse in the contact aureole indicates shallow intrusion depths of ~12 km, which suggests that the temperature of the wall rock was <300-350 °C during intrusion.

It has been suggested that the basal thrust of the Lycian nappes at the southern margin of the Çine submassif was reactivated in the Miocene as an extensional fault (Collins & Robertson, 1998, 2003; Göngör & Erdogan, 2001), the Lycian detachment (Figs. 1-3). Collins & Robertson (1998) proposed that reactivation was due to top-to-the-E extensional faulting. Rügner et al. (2003) showed that stretching lineations indicative of tectonic transport within a narrow zone underneath the base of the Lycian nappes were passively rotated from an N-S into an E-W direction suggesting that the top-to-the-E shear-sense indicators are rotated top-to-the-S structures. Below we will show that the cooling pattern in the southern Çine nappe is also not compatible with top-to-the-E extensional faulting. Another problem is that unequivocal evidence for extensional deformation at the southern margin of the Çine submassif is lacking. The geometry of the footwall of the Lycian detachment is a large-scale S-dipping monoclinc that characterizes the structure of the southern margin of the Çine submassif (Fig. 2).

In the Central Menderes metamorphic core complex, in the centre of the Anatolide belt, a later phase of bivergent brittle detachment faulting is associated with the development of the kuzyê and Güney detachments and the presently active Gediz and Büyük Menderes graben (Sengöl, 1987; Seyrignolu & Scott, 1996; Gessner et al. 2001a). Cohen et al. (1995) and subsequently Hoekstra (2000) showed that both graben formed in the Pliocene. The footwall of the top-to-the-N kuzyê detachment is a S-dipping monoclinc, whereas the footwall of the top-to-the-S Güney detachment is a N-dipping monoclinc. Both monoclins combine to form a large-scale syncline that characterizes the structure of the Central Menderes metamorphic core complex (Fig. 2). In contrast to the Sinan and Lycian detachments, the kuzyê and Güney detachments did not reactivate pre-existing structures but cut across them.

This summary suggests that at least two periods of bivergent normal faulting occurred in the Anatolide belt of western Turkey. Both periods are poorly dated. The first phase may have occurred during Miocene times and affected the upper structural levels of the Menderes nappes, which are now mainly represented by the two outer submassifs. The relationship of the Early Miocene extensional D4a shear bands in the northern Central Menderes metamorphic core complex to map-scale detachments is not known. The second phase of extensional deformation caused the development of the Central Menderes metamorphic core complex in the centre of the Anatolide belt and appears to be associated with the Pliocene graben.

3. Sampling strategy
For this study, orthogneiss, schist and granite samples were collected along a broadly NNE-SSW-trending transect from Sinan in the north to Milas in the south across the Anatalide belt (Fig. 3). We sampled at regular intervals of ~5-10 km to resolve the regional variation in cooling patterns across the main structural features. A causal link with denudation processes can be established by integrating the thermochronological data with additional independent geological information (Wheeler & Butler, 1994). Is the specific case of progressive unroofing of fault rocks below a crustal-scale normal fault, asymmetric cooling is expected, resulting in decreasing apparent ages along the direction of hanging-wall transport (Brandon, Roden-Tice & Garver, 1980; Foster & John, 1999; Ring et al. 1996b). Apatite fission-track ages can be correlated with distance in the direction of hanging-wall motion as inferred from asymmetric kinematic indicators (Bradbury & Nolen-Hoeksema, 1985; Foster, Miller & Miller, 1991; Foster et al. 1993; Johnson, 1997). This feature can be used to estimate slip rates for the causative fault and has also been used to reconstruct the pre-extensional geometry of the fault system (Foster et al. 1993; Foster & John, 1999; Gessner et al. 2001a; Ring, Thomson & Bröcker, 2003). For these reasons, the baseline samples for this study were collected parallel to the NNE-SSW tectonic transport direction of extensional deformation (Figs 2, 3).

4. Analytical procedures and data analysis
Standard heavy-liquid and zonagentic-separation methods were used to extract apatite, zircon and mica from the host rocks. Irradiation of the fission-track samples was carried out in the graphite reflector of the Risø reactor at Roskilde, Denmark, and monitored using Corning-glass dosimeters. Fissiontracks in apatite and zircon were analysed using the external-detector and c-calibration approach (Huford & Green, 1982; Huford, 1990). Ages were calculated using the central-age method of Galbraith & Laslett (1993), which allows for non-Poissonian variation within a population of single-grain ages belonging to an individual sample. Such variation in single-Eyivalent ages can occur in sedimentary samples as a consequence of differing apatite provenance. The relative error or dispersion on the central age is a measure of non-Poissonian variation. Where the relative error is greater than ~20%, the
Figure 1. Map showing sample localities for apatite and zircon fission-track and \(^{40}\)Ar/\(^{39}\)Ar-muscovite and biotite analysis. \(^{40}\)Ar/\(^{39}\)Argon data of samples 93T55 and 93T18 are from Hetzel et al. (1995a); \(^{40}\)Ar/\(^{39}\)Ar-muscovite data from samples 92T27, 92T54, T2 and T20 are from Hetzel & Reichmann (1996) and those from samples GD and BMD are from Lips et al. (2001). Also shown are extensional stretching trajectories of Late Oligocene/Early Mioce and latest Miocene/Recent phases of extensional faulting. Late Oligocene/Early Mioce top-to-the-NNE extension was ductile in northern Central Menderes metamorphic core complex and mainly brittle at Simav detachment; fault-plane solutions (lower-hemisphere focal projections, black indicates shortening quadrants), surface magnitude and slip vectors of Demirci and Alaplı earthquakes also shown. For abbreviations, see Fig. 1.
samples also likely to fail the standard $\chi^2$ test and the spread in single-grain ages is often attributable to the presence of different age populations.

Some fission-track ages from the Anatolide belt do not show single-grain spreads consistent with a single-age population when assessed using the relative error and the $\chi^2$ test. We think that the scatter in single ages is due to errors of commission and omission that resulted from the frequent presence of track-like crystal defects. Although possibly small in number, the misidentification of very short tracks and spurious etch pits has a significant effect because true spontaneous track numbers are low. The large number of crystals used and the likelihood that these two types of error are equally common means that they will tend to cancel out when the central age is calculated but produce a large single-grain-age spread. This conclusion is supported by the age occurrence of closely spaced samples, some of which fail the single-population tests while their neighbors do not.

Track-length data were obtained using a drawing tube and digitizing tablet. Typically, 100 track-length measurements were made for determining the mean track length and its standard deviation for track-length modelling and the construction of a thermal history. Fifteen of the 44 apatites analysed matched these criteria, while for 19 samples we obtained less than 100 track-length measurements. For 10 samples we did not obtain any track-length data.

The procedures for multi-grain fission-Track (ArAr) Ar analysis of samples was performed at the University of Manitoba in Winnipeg, Canada, and all samples were exposed to a thermal history of 1000°C. This has a significant effect because true spontaneous track numbers are low. The large number of crystals used and the likelihood that these two types of error are equally common means that they will tend to cancel out when the central age is calculated but produce a large single-grain-age spread. This conclusion is supported by the age occurrence of closely spaced samples, some of which fail the single-population tests while their neighbors do not.

Thermochronological data have been interpreted in terms of a temperature history by using a normal ‘closure temperature’ in conjunction with the mineral age as a point on a time-temperature history plot (e.g. Hurford, 1986). The following closure temperatures have been used in this study: ~110 ± 10°C for fluorapatite (Gleadow & Duddy, 1981) and 240–280°C for zircon (Tagami & Dumitru, 1996). The isotopic closure temperatures for Ar in biotite and muscovite are lower (200°C); they are determined using the appropriate closure temperatures for Ar diffusion in biotite and muscovite.

Our choice does not affect the general conclusions of this study in any significant way. More detailed thermal history information can be obtained from apatite fission-track data using a quantitative approach, based on empirically calibrated algorithms (e.g. Laslett et al. (1987) and a search algorithm to examine a wide range of time-temperature space for acceptable thermal history models (e.g. Gallagher, 1995). We used the modelling method of Gallagher (1995) and the annealing model for Durogne apatite of Laslett et al. (1987) to obtain thermal histories. To define an acceptable thermal history, we used a likelihood function (Gallagher, 1995), which is a measure of the probability that a given thermal history would have produced the observed data. Acceptable models are found by a stochastic search, using a genetic algorithm, of a model time-temperature space. In this method, the sampling of thermal histories is progressively biased toward those histories which maximize the likelihood of obtaining the observed data. In this study, a typical modelling run considered ~1 000 thermal histories defined by four to ten time-temperature points for a given sample location. We varied the model time-temperature input parameters until the predicted age and mean length of the modelling run was within 5% of the measured age and track length.

Regional variations in the thermal histories of all samples are readily visualized through contouring the site-specific temperature estimates at given time intervals. One way of displaying the cooling history of an orogen is to convert thermal history information into maps for different time intervals (Gleadow et al. 1999) (Fig. 5). To construct such maps, we used the 15 apatites for which we obtained 100 or more track-length measurements, and those results were used to help interpret the 19 less well-constrained samples.

5. Thermochronological data
5a. Results
Forty-four apatite fission-track analyses were completed whose central ages vary between 27.9 ± 1.2 Ma and 1.8 ± 0.6 Ma. In addition to the apatite results, four zircon analyses have central ages between 27.9 ± 1.5 Ma and 5.2 ± 1.1 Ma (Table 1). Ages from all samples are much younger than the crystallization or (pre-metamorphic) sedimentation age of the rocks from which they were obtained. Therefore, the fission-track analyses (central-age and track-length data) are interpreted as yielding cooling history information. The $\text{Ar}/\text{Ar}$-muscovite-isochron ages range between 31.6 ± 1.2 Ma and 24.7 ± 1.2 Ma (Fig. 4). All sample localities and additional samples for $\text{Ar}/\text{Ar}$ dating reported in Hetzel et al. (1995a), Hetzel & Reichmann (1996) and Lips et al. (2001) are shown in Fig. 3.
5.4. Interpretation of apatite fission-track data

The results of thermal history modelling of the apatite fission-track data are presented as a series of contoured time-temperature slices, advancing in nine steps from Early Miocene times to the present (Fig. 6). Note that the contouring is strongly controlled by the distribution of sample points (Fig. 3) and the discussion below should be considered accordingly. Prior to ~30 Ma, temperatures for all samples are greater than ~130 °C.

The apatite fission-track data alone do not yield any information about the thermal evolution of the Anauide belt prior to this time. The main features of these maps are listed below.

26 Ma: The first recorded cooling in the belt to temperatures within the apatite partial-annealing zone (~110-60 °C) (Fitzgerald et al. 1995) occurred on the southern margin of the Güões submassif. Initial cooling is based entirely on the age of 27.9 ± 1.2 Ma of sample T68.
### Table 1. Fission-product (f) data for apatite and zircon sample localities

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location (Lat/N)</th>
<th>Elevation (m)</th>
<th>Mineral &amp; no. of crystals</th>
<th>Sponatious (α)</th>
<th>Induced (α)</th>
<th>Dosimetry (Nd)</th>
<th>Central FT age (Ma fission 14)</th>
<th>Relative error (%)</th>
<th>MCLT (μs) (μs)</th>
<th>S.D. (μs)</th>
</tr>
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<tbody>
<tr>
<td>T1</td>
<td>38.19317</td>
<td>150</td>
<td>Apatite (25)</td>
<td>0.050</td>
<td>0.050</td>
<td>0.050</td>
<td>22.15±0.26</td>
<td>3.17</td>
<td>0.01</td>
<td>0.01</td>
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<tr>
<td>T2</td>
<td>38.24981</td>
<td>700</td>
<td>Apatite (25)</td>
<td>0.121</td>
<td>0.110</td>
<td>0.113</td>
<td>100.05±0.57</td>
<td>0.75</td>
<td>0.12</td>
<td>0.12</td>
</tr>
<tr>
<td>T3</td>
<td>38.24981</td>
<td>21811</td>
<td>Zircon (20)</td>
<td>3.040</td>
<td>3.040</td>
<td>3.040</td>
<td>24.20±1.12</td>
<td>14.62</td>
<td>N.D.</td>
<td>0.12</td>
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<tr>
<td>T4</td>
<td>38.24541</td>
<td>420</td>
<td>Apatite (10)</td>
<td>0.318</td>
<td>0.318</td>
<td>0.318</td>
<td>10.64±0.64</td>
<td>13.1</td>
<td>0.43</td>
<td>0.43</td>
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<tr>
<td>T5</td>
<td>38.23575</td>
<td>380</td>
<td>Apatite (22)</td>
<td>0.153</td>
<td>0.153</td>
<td>0.153</td>
<td>8.92±0.65</td>
<td>0.00</td>
<td>0.12</td>
<td>0.12</td>
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<tr>
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<td>830</td>
<td>Zircon (20)</td>
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<td>0.235</td>
<td>17.03±1.08</td>
<td>0.01</td>
<td>0.11</td>
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<td>1180</td>
<td>Apatite (20)</td>
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<td>0.027</td>
<td>0.027</td>
<td>10.48±1.40</td>
<td>0.02</td>
<td>N.D.</td>
<td>0.12</td>
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<tr>
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<td>38.21062</td>
<td>1200</td>
<td>Apatite (19)</td>
<td>0.011</td>
<td>0.011</td>
<td>0.011</td>
<td>7.90±1.75</td>
<td>0.00</td>
<td>N.D.</td>
<td>0.12</td>
</tr>
<tr>
<td>T9</td>
<td>38.19179</td>
<td>2150</td>
<td>Apatite (9)</td>
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<td>0.025</td>
<td>0.025</td>
<td>3.73±0.69</td>
<td>0.01</td>
<td>0.13</td>
<td>0.13</td>
</tr>
<tr>
<td>T10</td>
<td>26.06113</td>
<td>900</td>
<td>Apatite (25)</td>
<td>1.037</td>
<td>1.037</td>
<td>1.037</td>
<td>6.08±0.83</td>
<td>0.63</td>
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<td>0.12</td>
</tr>
<tr>
<td>T11</td>
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<td>1920</td>
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<td>1.037</td>
<td>1.037</td>
<td>8.74±2.67</td>
<td>59.94</td>
<td>N.D.</td>
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<td>370</td>
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<td>0.080</td>
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<td>7.15±0.62</td>
<td>0.10</td>
<td>N.D.</td>
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<tr>
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<td>Apatite (10)</td>
<td>0.400</td>
<td>0.400</td>
<td>0.400</td>
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<td>0.111</td>
<td>0.111</td>
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<td>24.83</td>
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<tr>
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<td>0.567</td>
<td>0.567</td>
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<td>0.118</td>
<td>0.118</td>
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<td>0.267</td>
<td>0.267</td>
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<td>0.544</td>
<td>0.544</td>
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<td>2.83</td>
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<td>10.48±1.28</td>
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<td>860</td>
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<td>0.517</td>
<td>0.517</td>
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<td>0.517</td>
<td>0.517</td>
<td>19.05±1.10</td>
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<td>0.517</td>
<td>0.517</td>
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<td>0.517</td>
<td>0.517</td>
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<td>Apatite (20)</td>
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<td>0.517</td>
<td>0.517</td>
<td>18.58±0.85</td>
<td>0.33</td>
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<td>0.517</td>
<td>0.517</td>
<td>18.58±0.85</td>
<td>0.33</td>
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<td>0.517</td>
<td>0.517</td>
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<td>0.33</td>
<td>0.13</td>
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<td>220</td>
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<td>0.517</td>
<td>0.517</td>
<td>27.85±1.22</td>
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<td>Zircon (9)</td>
<td>0.517</td>
<td>0.517</td>
<td>0.517</td>
<td>1.87±0.36</td>
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<td>Zircon (9)</td>
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<td>0.517</td>
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<td>1.87±0.36</td>
<td>2.40</td>
<td>N.D.</td>
<td>0.12</td>
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</table>
25 Ma Cooling into the uppermost aperite partial-annealing zone occurred in the northern Čine submassif. At the same time cooling at temperatures in the lower aperite partial-annealing zone occurred in the southern Čine submassif.

26 Ma: Cooling started in a down-faulted block in the Kućnik Menderes graben in the central part of the Central Menderes metamorphic core complex. Further cooling occurred in the Čine submassif; the locus of cooling appears to be in the southeastern sector of the submassif. Note that in the southern Čine submassif the cooling pattern is still controlled by sample T6; all other samples are still above ~130°C.

20 Ma: In the Čine submassif, cooling started to progress northward. The samples in the Čine submassif are not evenly distributed but instead lie on one single NNE–SSW transect, and the limited E–W distribution of sample points restricts the interpretation of the data in this orientation. However, it seems reasonable to speculate that cooling in the entire Čine submassif propagated northward. Cooling in the Čine submassif continued and, in general, it seems that cooling progressed in a more southerly direction.

18 Ma: By this time much of the southern Čine and the northern Čine submassifs are predicted to have been at temperatures in the lower aperite partial-annealing zone or below. With the exception of the basement block in the Kućnik Menderes graben, the Central Menderes metamorphic core complex remains at temperatures >130°C. The contoured temperature distribution in the northern Čine submassif is a little misleading because sample points are not evenly distributed but lie on two NNE–SSW transects in the western and eastern part of this submassif. Therefore, the relatively warm area in the central part of the northern Čine submassif is an artefact of the contouring procedure.

14 Ma: Cooling below temperatures of the aperite partial-annealing zone in the northern part of the Čine submassif and the southern part of the Čine

Table 1 (cont.)

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location (Lat°/Long° E)</th>
<th>Elevation (m)</th>
<th>Mineral and no. of crystals</th>
<th>Spontaneous (\chi_0) (N)</th>
<th>Isotropic (\alpha_0) (N)</th>
<th>Anisotropic (\alpha_{10} \alpha_{20}) (N)</th>
<th>Dominator (\phi_i) (N)</th>
<th>Central FT age (Ma ± 1σ)</th>
<th>Relative error %</th>
<th>MCTL (μm) and no. of tracks</th>
<th>S.D. 1σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>T71</td>
<td>28.47938</td>
<td>90</td>
<td>Aperature (20)</td>
<td>0.25</td>
<td>0.12</td>
<td>0.13</td>
<td>0.17</td>
<td>22.57 ± 0.11</td>
<td>9.47</td>
<td>14.00 ± 0.12</td>
<td>1.24</td>
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<td>28.47209</td>
<td>110</td>
<td>Aperature (15)</td>
<td>0.39</td>
<td>0.17</td>
<td>0.17</td>
<td>0.17</td>
<td>21.66 ± 0.16</td>
<td>15.88</td>
<td>14.42 ± 0.08</td>
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<td>0.15</td>
<td>0.17</td>
<td>0.17</td>
<td>22.86 ± 0.13</td>
<td>0.20</td>
<td>14.87 ± 0.18</td>
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<td>0.17</td>
<td>0.17</td>
<td>0.17</td>
<td>21.04 ± 0.21</td>
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<td>14.52 ± 0.19</td>
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<td>0.17</td>
<td>21.31 ± 0.88</td>
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<td>0.15</td>
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<td>24.69 ± 0.12</td>
<td>10.54</td>
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<td>N.A.</td>
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<tr>
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<td>0.15</td>
<td>0.15</td>
<td>22.52 ± 0.18</td>
<td>10.54</td>
<td>N.D.</td>
<td>N.A.</td>
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<td>N.A.</td>
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<td>0.15</td>
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<td>15.12 ± 1.09</td>
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<td>0.15</td>
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<td>14.75 ± 0.14</td>
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<td>0.15</td>
<td>18.20 ± 1.15</td>
<td>14.61</td>
<td>14.34 ± 0.10</td>
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<tr>
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<td>18.23 ± 1.06</td>
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<tr>
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<td>Aperature (20)</td>
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<td>0.15</td>
<td>0.15</td>
<td>18.23 ± 1.06</td>
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<td>N.A.</td>
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<tr>
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<td>0.15</td>
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<td>18.23 ± 1.06</td>
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<td>0.15</td>
<td>0.15</td>
<td>18.23 ± 1.06</td>
<td>10.54</td>
<td>N.D.</td>
<td>N.A.</td>
</tr>
</tbody>
</table>

Dosimeter track densities (d100’s) are 10^6 cm^-2; other track densities (d10’s, d20’s) are 10^5 cm^-2; except aperature where (d10’s) = 10^6 cm^-2; number of tracks counted shown in brackets.

Analyses are by external-detector method using 0.5 for 4σ/2σ geometry-correction factor.

Zircon ages calculated using dosimeter glass CN2 with C100 = 137 ± 2.14; aperature ages calculated using dosimeter glass CNS with C100 = 361 ± 23.

S.D. = standard deviation; MCTL = mean counted track length; N.D. = not determined; N.A. = not applicable.

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U. RING AND OTHERS
submassif was completed. Temperatures throughout the Gördes and Çine submassifs were at surface values at this time, except a small warm spot in the southwestern Çine submassif, which is controlled by sample T84. During the 18–14 Ma period, cooling took place in the southeastern part of the Central Menderes metamorphic core complex. The contoured temperature distribution in the footwall of the Güney detachment is probably slightly misleading since the relatively cool samples in the Kılıçlık Menderes graben in the centre of the Central Menderes metamorphic core complex (samples T36, T40, T42, T44) control the map pattern. Sample
Figure 6. Time slices from 28 Ma to 2 Ma imposed on general outline of Menderes nappe with major detachments (Fig. 1); map for 28 Ma shows locations of 34 samples used to construct maps; contours were drawn by computer program but were subsequently manipulated to include geological information, for example, hanging wall and footwall of detachments were not considered by computer-controlled contouring procedure, but manipulation ensured that cooling of footwall samples did not affect those of the hanging wall. Cooling below ~ 130 °C commenced in southern Gördes subnappes at ~ 28 Ma but did not progress northward until 22–20 Ma. Cooling in northern Gürne subnappes started at 25 Ma and progressed southward towards base of Lycaen nappe. Down-faulted block in the central Central Menderes metamorphic core complex started to cool ~ 22 Ma. Footwall of Kusey detachment remained at elevated temperatures above ~ 130 °C until 5 Ma; footwall of Güzey detachment started to cool below ~ 130 °C at 14 Ma but final cooling only occurred after 2 Ma.
T35 from the immediate foothill of the central Güney detachment is still above 130°C. This temperature is more likely to be typical of the footwall of the Güney detachment.

10 Ma: Cooling to temperatures below the apatite partial-annealing zone at the margins of the Kıuçek Menderes graben occurred. Temperatures in the southern segment of the Central Menderes metamorphic core complex remained at values of ~80–90°C. In the central part of the Central Menderes metamorphic core complex cooling into the apatite partial-annealing zone has occurred.

5 Ma: The northern flank of the Central Menderes metamorphic core complex below the Kuzey detachment remained at temperatures above 130°C. The southern flank of the Central Menderes metamorphic core complex below the Güney detachment is also thought to be at temperatures >80°C; however, this inference is entirely based on sample T35.

2 Ma: The northern and southern flanks of the Central Menderes metamorphic core complex below the Güney and Kuzey detachments cool to similar temperatures of ~60–80°C.

The final cooling to surface temperatures occurred during the interval 2–4 Ma in the footwalls of the Kuzey and Güney detachments.

5.5c. Summary of apatite fission-track data

The onset of cooling from >130 to around 60°C recorded by the apatite fission-track data occurred during Late Oligocene times in the southern Gördes submassif, but this cooling is restricted to sample T86. Regional cooling in the Gördes submassif commenced at ~20 Ma and was completed by ~14 Ma. In the Çine submassif regional cooling occurred during the same time span. The cooling pattern as derived from the apatite fission-track data supports top-to-the-S/SW extension, not top-to-the-E extension and therefore supplies independent support for the conclusion by Régnier et al. (2003) that transport lineations were passively rotated within a relatively narrow zone at the base and directly underneath the Lycian nappes after tectonic movement (see Section 2.2c). The central part of the Central Menderes metamorphic core complex also began cooling at this time. Nonetheless, with the exception of the down-faulted basement block in the Kıuçek Menderes graben, the present outcrop of the Central Menderes metamorphic core complex remained at temperatures greater than 130°C during this period. Cooling in the Gördes submassif appears to have migrated northwards towards the Simav detachment and the İmroz–Ankara zone, and southwards in the Çine submassif towards the base of the Lycian nappes.

From Late Miocene time onwards, a second phase of cooling from >130 to around 60°C affected much of the Central Menderes metamorphic core complex, Cooling across the inner submassif was asymmetric with early cooling in the south. However, it appears that pronounced cooling in the immedia occulsion of the Kuzey and Güney detachments did not start until the Pliocene. Final cooling to surface temperatures of footwall rocks to the Güney and Kuzey detachments occurred during the Quaternary.

Although the cooling patterns appear to be locally asymmetric towards the major extensional detachments bounding the various submassifs, the overall finite pattern across the western Anatolian belt is remarkably symmetric.

5.d. Regional cooling histories

By combining the thermal histories inferred from apatite fission-track data with the zircon fission-track data (Table 1), the 40Ar/39Ar-muscovite ages shown in Fig. 4, and published 40Ar/39Ar-muscovite and biotite ages (Hetzet & Reischmann, 1996; Lipts et al. 2001), we illustrate the regional variations in the temperature history of the different submassifs (Fig. 7). The oldest cooling phase we can constrain with the data occurred in the southern Çine submassif (Fig. 7a). Initial cooling between ~40 Ma and ~25 Ma was slow (<10°C Myr−1). The distribution of the 40Ar/39Ar-muscovite ages (Figs 3, 7a) suggest that cooling from maximum metamorphic temperatures of ~500–550°C (Régnier et al. 2003) at the southern margin apparently progressed northward and structurally downward in the Çine submassif, indicating that the muscovite ages are cooling ages. More rapid cooling at rates greater than 30°C Myr−1 occurred between ~25 and 18 Ma. In contrast to the northward younging 40Ar/39Ar-muscovite ages, the apatite fission-track data show that shallow-level cooling progressed southward and structurally upward (Figs 2, 6).

The footwall of the Simav detachment cooled rapidly at rates greater than 50°C Myr−1 from after 25 to 20 Ma (Fig. 7b). This inference is based on one single 40Ar/39Ar-muscovite age and seven apatite fission-track ages.

The hanging wall of the Kuzey detachment (the southernmost Gördes submassif), the central Central Menderes metamorphic core complex and the hanging wall of the Güney detachment cooled rapidly at the Oligocene/Miocene boundary at rates greater than 50°C Myr−1 (Fig. 7c–e). The presently exposed rocks of the Gördes and Çine submassifs and those in the central part of the Central Menderes metamorphic core complex were close to the surface in the Miocene (Fig. 7a–e).

In contrast, the footwall rocks of the Kuzey and Güney detachments remained at relatively high temperatures well after cooling to near surface temperatures had occurred in the two outer submassifs. After intrusion of the Saltili granite at ~20 Ma, the rocks
Figure 7. (a-e) Temperature-time diagrams for five areas along NNE-SSW cross-section (Figs 1, 2) combining 40Ar/39Ar-amphibole, muscovite and biotite, and apatite and zircon fission-track data; assumed closure temperature for Ar in muscovite is ~400–450 °C; ~350 °C for Ar in biotite, 240–280 °C for fission tracks in biotite and ~110 ± 10 °C for felspar; intrusion age of Salkil granite and 40Ar/39Ar-biotite data from Salkil granite are from Hetzel et al. (1995d); 40Ar/39Ar-muscovite data from samples 92T27, 92T54, T2 and T20 are from Hetzel & Reischmann (1996) and those from samples GD and BMD are from Lips et al. (2001). Sample numbers in grey in (c) and (e) represent samples from hanging wall. T-t histories for apatite were obtained from apatite fission-track data using Montenegro program of Gallagher (1995) which is based on apatite-annealing model of Laslett et al. (1987) and takes track-length distributions into account (note that for temperatures above ~110 °C and below ~60 °C, apatite fission-track data provide no effective constraint on model evolution); whenever zircon fission-track or 40Ar/39Ar-mica data were available, cooling curves were further refined by including those data. Çine (a) and Göñes submassifs (b and c) and also the central part of the Central Menderes metamorphic core complex (d) cooled in Early and Middle Miocene times to near surface temperatures, therefore Çine and Göñes submassifs can be viewed as being "pinned" to Earth’s surface, providing fixed framework in time and space to consider the subsequent emergence of the Central Menderes metamorphic core complex. Final cooling in footwall of the Kuzey (c) and Günyøy detachment (e) occurred in Plio-Pleistocene times; early cooling of footwall of the Günyøy detachment is poorly constrained, and dashed and non-dashed paths indicate alternative early cooling paths after metamorphic climax at 36–38 Ma (sample BMD). Note that tracks above and below the Günyøy detachment have a relatively similar cooling history in Late Oligocene and Early Miocene times, that is, before the Günyøy detachment formed, and that the poorly resolved Early/Middle Miocene cooling event in footwall of the Günyøy detachment coincides in time with D1d extension during intrusion of Salkil and Turgutlu granitoids in the footwall of the Kuzey detachment.
in the footwall of the Kuzey detachment did not cool to any significant degree until ~5 Ma when they started to cool rapidly at rates greater than 50 °C Myr⁻¹ (Fig. 7c). Cooling in the footwall of the Giney detachment was more complex and involved a first poorly resolved phase in the Early/Middle Miocene followed by final cooling from shallow depth after ~5 Ma (Fig. 7e). The dashed and the non-dashed lines in Fig. 7e show two possible cooling paths for early cooling of the Giney footwall and highlight the problem in defining the early time-temperature history of the footwall of the Giney detachment.

6. Discussion

6.a. Denudation of the outer submassifs

In the Górdes submassif, the apatite fission-track ages decrease from 28 Ma at the southern margin to 16 Ma near the Simav graben in the north. However, sample T68 probably does not underlie the Simav detachment and therefore its cooling history, which is distinctly different from that of the other samples from the Górdes submassif, appears unrelated to movement on the Simav detachment. In the central part of the Górdes submassif, the apatite fission-track ages are virtually similar at 19-21 Ma (Fig. 2). These cooling ages are the same as or slightly younger than those for the syn-detachment intrusion of the Eğrigöz granite at 20.7 ± 0.5 Ma (Ring & Collins, 2003) and demonstrate fast Early Miocene cooling (Fig. 7b). Therefore, and because the apatite fission-track ages young northward in the direction of tectonic transport on the Simav detachment, cooling was related to movement on the Simav detachment.

A slip rate of ~10 km Myr⁻¹ can be estimated from the inverse slope of apatite fission-track ages with distance in the slip direction (Fig. 2). Offset of geological markers suggests a minimum displacement of ~50 km (Figs 1-3). Barometric estimates show that the total amount of exhumation since the Eocene peak of metamorphism in the Menderes nappes is of the order of 20-25 km (Ring et al., 1999a). Gesner et al. (2001a) showed that movement on the Kuzey and Giney detachments accounts for the final 10-12 km of this exhumation. Assuming that all of the earlier exhumation was due to detachment faulting, we can calculate a dip angle of the Simav detachment, which, if displacement was assumed to be 50 km, would be between 10° and 20°. If a smaller displacement of 25 km was assumed, the dip angle would be of the order of 25-30°. If erosion and/or vertical ductile thinning acted as additional exhumation processes or displacement was greater than 50 km, then the dip angle for the Simav detachment would be less. Although these estimates are not very accurate, the conclusion that the Simav detachment had an initially low dip angle, that is, less than 30°, appears to be robust.

In the Cine submassif, the apatite fission-track ages young southward from ~22 Ma in the north to 18 Ma in the south (Fig. 2) and reveal a regional pattern strikingly similar to that in the footwall of the Simav detachment in the Górdes submassif. As the apatite fission-track cooling patterns are so similar, the cause of cooling may also be similar. This assumption has far-reaching implications because it demands the existence of an Early Miocene extensional detachment at the southern margin of the Cine submassif. However, as noted above, there is to well-defined extensional fault known from this area. Another problem is the geometry of the footwall of the demanded Lycian detachment. If the monocline in the footwall of the detachment represents a refolded system (Hart & Bartley, 1995), a top-to-the-N slip faulting would be implied for which there is no evidence.

The cooling curve at the southern margin of the Cine submassif (Fig. 7a) indicates an initially slow cooling rate of ~10°C Myr⁻¹ during Late Eocene and Oligocene times. The only major structure known from the Cine submassif that was active at this time is the Eocene Selimiye shear zone (Hetzl & Reuschmann, 1996). The northward- and structurally downward-younging ⁴⁰Ar/³⁹Ar-white-mica ages are compatible with top-to-the-S thrusting and are another strong indication that the Selimiye shear zone is a thrust and not an extensional fault as advocated by Bockgart & Park (1994). We relate slow cooling primarily to erosion after movement on the Selimiye shear zone. Erosion rates in this case would be of the order of 0.2-0.5 km Myr⁻¹. Acceleration of cooling to values greater than 30 °C Myr⁻¹ coincides with the proposed Late Oligocene-Miocene extensional reactivation of the basal thrust of the Lycian nappes (Collins & Robertson, 1998, 2003; Gümüş & Erdoğan, 2001). Our data indicate a Late Oligocene age for the onset of extension. The southward and structurally upward younging of the apatite fission-track data fits well with the interpretation that this phase of movement of the Lycian nappes was extensional and implies a slip rate of ~4-5 km Myr⁻¹. Our general interpretation of cooling in the Cine submassif is that the fission-track ages are related to extensional faulting whereas the ⁴⁰Ar/³⁹Ar ages are not. This interpretation implies that the extensional fault roots at a relatively high crustal level, most probably at the brittle/ductile transition.

Cooling to near-surface temperatures in the southern part of the Górdes submassif and the northern part of the Cine submassif occurred in the Early to Middle Miocene and both outer submassifs were deposited on basement, which has a regionally subhorizontal Eocene foliation. Afterwards, a Late Miocene possibly Early Pliocene erosion surface formed on the flat-lying sediments on these plateau
6.3. Central Menderes metamorphic core complex

The asymmetry of the cooling patterns across the Kuzey and Gűney detachments and their associated graben indicates that the primary control on the unroofing of the Central Menderes metamorphic core complex was tectonic rather than erosional. Both the Kuzey and Gűney detachments and the present-day Gediz and Büyük Ménderes graben root below the corresponding Gűrides and Cine submassifs (Cohen et al. 1995). This structural arrangement and the kinematic indicators during extensional deformation (Fig. 3), showing northerly hanging-wall transport in the northern Central Menderes metamorphic core complex (Hetzel et al. 1995a) and southerly directed hanging-wall transport in the southern Central Menderes metamorphic core complex (Emre & Stöblitiz, 1997), together with the temperature history information, support the contention that the Central Menderes metamorphic core complex underwent tectonic denudation that culminated in its exposure at the surface in the Pliocene and Quaternary.

Gessner et al. (2001a) showed that the gross synclinal structure and the brittle detachment systems in the central Central Menderes metamorphic core complex are related structures. Tectonic unloading along the detachments induced upward flexure of the upper crust. This rotated the presently exposed detachment surfaces into their present low angle of dip during progressive exhumation. The model of Gessner et al. (2001a) implies that the steep faults, which bound the Gediz and Büyük Menderes grabens, formed as the two detachments were rotated into shallower orientations, that is, detachment and graben faults are related and manifest different stages of faulting and exhumation. Thermal modelling suggests that cooling related to formation of the Central Menderes metamorphic core complex started in the latest Miocene to Pliocene and seismic activity suggests that this process is still operating, that is, the Central Menderes metamorphic core complex is a nascent core complex. The estimated displacement of ~10-12 km infers that the detachments operated at average slip rates of ~2 km Myr⁻¹. The initial dip angle was of the order of ~60° for the Kuzey detachment and ~40° for the Gűney detachment (Gessner et al. 2001a). Hence, the slip rates, the amount of displacement, the dip angles and the geometry of the footwall for the two detachments that bound the Central Ménderes metamorphic core complex are significantly different from those of the two outer detachments.

While the cooling history of the footwall of the Kuzey detachment is straightforward, the cooling history for the Gűney detachment is more complex. This complexity might be an artefact caused by the poor data coverage, or by the fact that at the onset of cooling the southern margin of the Central Menderes metamorphic core complex had a lower geothermal gradient than that of the northern margin, which had been intruded by granitic magmas. In this case, the footwall of the Gűney detachment may have had less potential to record a high-resolution cooling history. Another possibility might be that the poorly resolved early cooling history in the Miocene was related to movement on the proposed Lycian detachment. However, this scenario would demand a displacement of up to 100 km on the Lycian detachment. Hayward (1984) proposed that the Lycian nappe moved ~70 km in Miocene times.

6c. Proposed denudation history and the symmetric cooling patterns

A working hypothesis for the unroofing of the Anatolide belt is summarized in Fig. 8. Crustal shortening and suturing of the İzmir-Ankara zone with the rocks of the future Anatolide belt and southerly directed thrusting of the Lycian nappe and the Cycladic blueschist unit occurred from Late Cretaceous to Eocene times (Collins & Robertson, 1998; Gessner et al. 2001c). A first phase of exhumation commenced in the Late Oligocene and Early Miocene and appears to be symmetric across the Anatolide belt. Relative displacement between the upper/middle and lower units and the Anatolide belt in the north was accommodated by the Sinus detachment. We also envisage extension in the southern part of the belt was accommodated by top-to-the-S extensional movement at the base of the Lycian nappe. The early extensional phase resulted in fragmentation and thinning of the upper and middle units promoting cooling in the Gűrides and Cine submassifs.

During Pliocene times, extension was transferred towards deeper structural levels of the Anatolide belt and the Kuzey and Gűney detachments formed. Both detachments deformed the Late Miocene/Pliocene erosion surface, facilitated tectonic denudation of the Central Menderes metamorphic core complex until the present and also appear to have a symmetric regional arrangement.

Although the general Late Oligocene/Early Miocene cooling pattern is symmetric on a regional scale, we note that syn-extensional granites are only exposed in the northern half of the Anatolide belt, that is, the Turgutlu and Sahili granites south of the Gediz graben and the Demirci, Koyunoba, Efigöz and minor
formed at high angles and cut across pre-existing structures.

Finally, it should be noted that the regional cooling and extension pattern in the Menderes nappes of the Anatolide belt in western Turkey must not be the only extensional structures. We showed that extensional faulting and associated cooling progressed structurally downward towards the centre of the Anatolide belt, suggesting that older normal faults should be searched for in the upper tectonic units. Geisser et al. (2001a) discussed the possibility of an Eocen top-to-the-NE ductile normal shear zone within the Cycladic blueschist unit.

6.d. Comparisons with the pattern of extensional faults in the central and southern Aegean Sea

The pattern of top-to-the-N and top-to-the-S extensional faults in the Anatolide belt of western Turkey is more or less symmetric at the orogen scale. Accord-
ingly, the pattern of apatite fission-track ages also has a symmetric arrangement at this scale. In contrast, the nearby Aegean is characterized by more uniform top-to-
the-N extensional faults, with top-to-the-E extension on Samos, the top-to-the-S los detachment and the little-known top-to-the-S detachment on Serifos Island being the only known exceptions (Lee & Lister, 1992; Buick, 1991; Lister & Forster, 1996; Forster & Lister, 1999; Thomson, Stöckhert & Brix, 1999; Ring, Laws & Bernet, 1999; Grasemann et al. 2002; Ring & Reischmann, 2002). The apatite fission-track ages reported by Alhertest et al. (1982), Thomson, Stöckhert & Brix (1999) and Ring, Thomson & Bröcker (2003) from the Aegean appear to young northward from ~13–20 Ma on the island of Crete to ~8–12 Ma on the islands in the Cyclades. This northward-younging trend agrees with the admittedly poorly constrained ages for the extensional detachments that primarily caused the cooling. The Cretan detachment, which probably underlies much of the southern Aegean (Ring & Reischmann, 2002) operated at ~20 Ma (Thomson, Stöckhert & Brix, 1999), the los detachment at ~12 Ma (Henjes-Kunst & Kreuzer, 1982) and the Montsouria detachment on Naxos and the Varia detachment on Syros and Tinos at ~12 Ma (Lister & Forster, 1996; Ring, Thomson & Bröcker, 2003) (Fig. 1, inset). The differences in the pattern of detachment faults and associated cooling between the Aegean and the western Anatolide belt underscores that there are major along-
strike differences in the Hellenide–Anatolide orogen and that the cause for extension in both areas was different.

In the Aegean the southward retreating Hellenic subducting slab is widely envisioned to be the cause of detachment faulting (Lister, Banga & Foastra, 1984; Buick, 1991; Thomson, Stöckhert & Brix, 1999; Ring & Loyer, 2003). Hetzel et al. (1995b) proposed that the collision of the exogenic Anatolide microcontinent in Eocene times halted subduction in the Anatolide belt. Hetzel et al. (1995b) proposed that differences in potential energy associated with thermal relaxation caused Late Oligocene/Early Miocene extension in the Anatolide belt, which is consistent with extensional faulting in the ‘inner’ submassif and the formation of the Central Menderes metamorphic core complex. The conclusion that detachment faulting from Pliocene times to the present controlled the cooling history of this central section of the Anatolide belt. Extensional faulting also controlled earlier cooling in the Gördes and Çine submassifs. A hypothesis has been put forward, which explains cooling in these two submassifs as a consequence of Late Oligocene to Early Miocene normal faulting and thinning of a crustal-scale hanging wall consisting of the Cycladic blueschist unit, the Lycian nappes and the İzmir–Ankara zone.

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References

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related granites in the northern Géodes massif. An additional problem with the Turgutlu and Salihi granities is that they are syn-tectonic with respect to the dextral \( D_{2a} \) extensional event at \( \approx 18-20 \text{ Ma} \) (Hettstedt et al., 1995a), but this extensional event did not cause any significant cooling of the granites (Fig. 7c). The timing of the \( D_{2a} \) extensional event in the northern Central Mesoliria metamorphic core complex fits well with the timing for the Sinvar detachment and it might be feasible to assume that a small-scale extensional shear zone formed in the footwall of the Sinvar detachment. However, this interpretation would demand a displacement of more than 100 km on the Sinvar detachment. It was accepted that cooling of sample T58 from the southernmost Géodes subsurface was not related to movement on the Sinvar detachment, then the Sinvar detachment was not directly associated with intrusion of the Turgutlu and Salihi granites. Another deviation from the overall symmetry of the extensional faulting is that the Sinvar detachment in the north is associated with extensive basin development, whereas in the south, Miocene basins are more limited.

Was the detachment history of the western Anatolide belt the product of two distinct events, or did it arise from a single extensional event that progressed more or less continuously from higher towards deeper structural levels? We argue that the cooling curves suggest two distinct events that were separated in time by \( \approx 15 \text{ Myr} \). There is growing evidence that detachments are short-lived (up to \( \approx 5 \text{ Myr} \)) and relatively rapidly operating structures triggered by tectonic pulses (e.g., Lister & Fossey, 1996; Ring, Thompson & Bröcker, 2003). The different tectonic characteristics of the two outer detachment systems as compared to the detachments bounding the Central Mesoliria metamorphic core complex are in line with our inference that they were caused by two separate events.

The reactivation of the Cyclades–Mesoliria thrust and the basal thrust of the Lycian nappes during the first extension event appears to be very important for the formation of the detachment as low-angle faults. Low-angle detachments in the Aegean also reactivated former thrust planes (e.g., the Cretan, the low and the Vrhi detachments). The Lycian and Géodes detachments
large-scale out-of-sequence thrusting during Eocene collision in western Turkey. *Journal of the Geological Society* 152, 769–84.


Tectonic denudation of a Late Cretaceous


