

Kinematics of the Amanos Fault, southern Turkey, from Ar/Ar dating of offset Pleistocene basalt flows: transpression between the African and Arabian plates

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Abstract: We report four new Ar/Ar dates and 18 new geochemical analyses of Pleistocene basalts from the Karasu Valley of southern Turkey. These rocks have become offset left-laterally by slip on the N20°E-striking Amanos Fault. The geochemical analyses help to correlate some of the less-obvious offset fragments of basalt flows, and thus to measure amounts of slip; the dates enable slip rates to be calculated. On the basis of four individual slip-rate determinations, obtained in this manner, we estimate a weighted mean slip rate for this fault of 2.89 ± 0.05 mm/a ($\pm 2\sigma$). We have also obtained a slip rate of 2.68 ± 0.54 mm/a ($\pm 2\sigma$) for the subparallel East Hatay Fault farther east. Summing these values gives 5.57 ± 0.54 mm/a ($\pm 2\sigma$) as the overall left-lateral slip rate across the Dead Sea fault zone (DSFZ) in the Karasu Valley. These slip-rate estimates and other evidence from farther south on the DSFZ are consistent with a preferred Euler vector for the relative rotation of the Arabian and African plates of $0.434 \pm 0.012^\circ \text{ Ma}^{-1}$ about 31.1°N , 26.7°E . The Amanos Fault is misaligned to the tangential direction to this pole by 52° in the transpressive sense. Its geometry thus requires significant fault-normal distributed crustal shortening, taken up by crustal thickening and folding, in the adjacent Amanos Mountains. The vertical component of slip on the Amanos Fault is estimated as c. 0.15 mm/a. This minor component contributes to the uplift of the Amanos Mountains, which reaches rates of c. 0.2–0.4 mm/a. These slip rate estimates are considered representative of time since 3.73 ± 0.05 Ma, when the modern geometry of strike-slip faulting developed in this region; an estimated 11 km of slip on the Amanos Fault and c. 10 km of slip on the East Hatay Fault have occurred since then. It is inferred that both these faults came into being, and the associated deformation in the Amanos Mountains began, at that time. Prior to that, the northern part of the Africa–Arabia plate boundary was located further east.

Horizontal crustal motions in the eastern Mediterranean region relate to interactions between the African, Arabian, Turkish and Eurasian plates (AF, AR, TR and EU, respectively). The Dead

Sea fault zone (DSFZ) forms the AF–AR boundary (Fig. 1); it is well established that its southern part, between Israel and Jordan, is a left-lateral transform-fault zone (e.g. Freund *et al.* 1970;

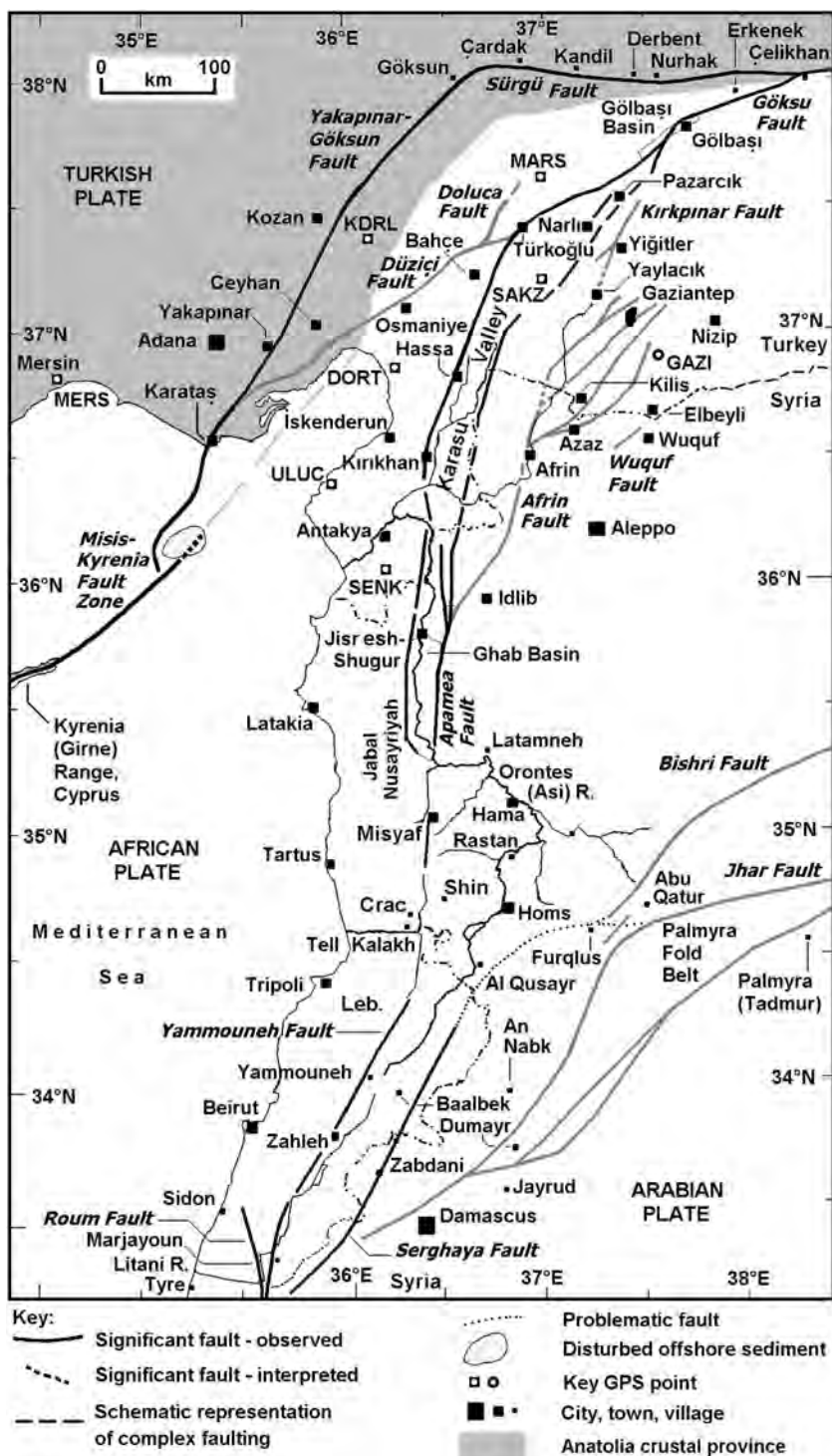


Fig. 1. Continued.

Garfunkel 1981). It has recently become clear (Westaway 2003, 2004; Gomez *et al.* 2006) that the northern DSFZ, in western Syria and southern Turkey, is transpressive; moving northward, it steps progressively to the right, away from its Euler pole. As currently defined (Westaway 2003, 2004), the northernmost DSFZ segment, consisting of multiple en échelon fault strands, runs along the Karasu Valley in the Hatay region of southern Turkey. The mainly left-lateral Amanos Fault, bounding the Amanos Mountains at the western margin of the Karasu Valley, between Kırıkhan, Hassa and Türkoğlu (Figs 1 & 2), is the subject of this study.

A geological map and a DEM-based topographic map of the area of Figure 1 were recently published by Gomez *et al.* (2006), and so are not repeated here. More detailed geological and topographic maps of the Karasu Valley and its immediate surroundings were recently published by Westaway *et al.* (2006a) and so, likewise, are not repeated here. Such illustrations enable the reader to visualize the topography, and its relationship to geological structure, along the DSFZ, supporting the much more detailed descriptions of key localities (Fig. 3) in the present study. Westaway *et al.* (2006a) also showed a map indicating the relationship between the DSFZ and other plate-boundary fault zones in the Middle East and eastern Mediterranean regions; such an illustration is thus not repeated here either.

This study region forms the NW margin of the Arabian Platform, which collided with the Anatolian continental fragment to the north (Fig. 1) during the Cenozoic, following the closure of the Southern Neotethys Ocean. During the Late Cretaceous (Maastrichtian; *c.* 70 Ma; e.g. Dilek & Delaioye 1992), the Hatay ophiolite was obducted on to the Arabian Platform margin. Strike-slip faulting along the AF–AR plate boundary initiated in the

early Middle Miocene (e.g. Garfunkel 1981), but the locations of, and slip senses on, active faults within this boundary have subsequently varied over time (e.g. Westaway 2003, 2004). The modern plate-boundary geometry is thought to have existed since *c.* 4 Ma (e.g. Westaway 2003, 2004), with the latest estimate of its initiation being 3.73 ± 0.05 Ma (Westaway *et al.* 2006a).

An understanding of the active kinematics of this region is important for several reasons. First, knowledge of slip rates on active faults is necessary to assess the recurrence intervals of large earthquakes, in order to quantify the regional seismic hazard (cf. Meghraoui *et al.* 2003). The most recent large earthquake in this region occurred on 13 August 1822. It had a macroseismic epicentre at 36.7°N , 36.5°E , near Aktepe (Fig. 3); a macroseismic magnitude of 7.5; and involved *c.* 200 km of fault rupture (Ambraseys 1989; Ambraseys & Jackson 1998). However, despite historical records of damage, there is no indication of which fault(s) ruptured during this event. Second, such knowledge is also important for testing regional kinematic models: in this case for the linkage between the AF–AR, AF–TR, and TR–AR plate boundaries (e.g. Westaway 2004). This includes comparison between short-timescale kinematic models from GPS satellite geodesy, such as that by McClusky *et al.* (2000), and longer-timescale geological evidence. Third, the recognition of the transpressive geometry of the northern DSFZ means that it provides scope for testing quantitative models for the distributed deformation associated with transpression, such as that by Westaway (1995). This aspect also relates to the development of the topography and structure of the Amanos Mountains adjoining the northern DSFZ (Figs 2 & 4). It has long been accepted that this mountain range developed as a consequence of the Cretaceous ophiolite obduction (e.g. Schwan 1971, 1972; Tolun

Fig. 1. Map (adapted from Fig. 2 of Westaway 2004, which lists original sources of information) of fault segments of the boundaries between the African, Arabian, and Turkish plates in southern Turkey, western Syria, and Lebanon, in relation to GPS points, from McClusky *et al.* (2000). Faults that are inferred to be significant during the present phase of deformation (post-*c.* 3.7 Ma) are black; those that appear to have been important earlier are grey. The Amanos Fault is shown running southward along the eastern margin of the Amanos Mountains for a distance of *c.* 110 km, from Türkoğlu, past İslahiye and Hassa, to Kırıkhan. K.V. denotes the Karasu Valley. The present study region comprises its *c.* 30-km-long segment between Ceylanlı, *c.* 5 km north of Kırıkhan, and Hassa. The subparallel East Hatay Fault passes through Tahtaköprü, *c.* 12 km farther east, bounding the western margin of the Kurd Daglı Range. The crust of the Anatolian continental fragment is shaded (except where offshore); its southern limit, at the northern margin of the Arabian Platform, marks the suture of the Southern Neotethys Ocean. As currently defined (e.g. Westaway *et al.* 2006a), the Göksu Fault, Sürgü Fault and Gölbaşı–Türkoğlu Fault (between Gölbaşı and Türkoğlu; illustrated) form the SW end of the East Anatolian fault zone, the boundary between the Turkish and Arabian plates. The ‘problematical fault’ is the inferred NE continuation of the Serghaya Fault across the Homs area of Syria (Walley 1998), which – if it exists – appears to have not been active since at least the Early Pliocene (Westaway 2004). The two symbols for GPS points represent points located within towns (which are named in Fig. 2), and point GAZI, which is named after Gaziantep but located some distance away from this city.

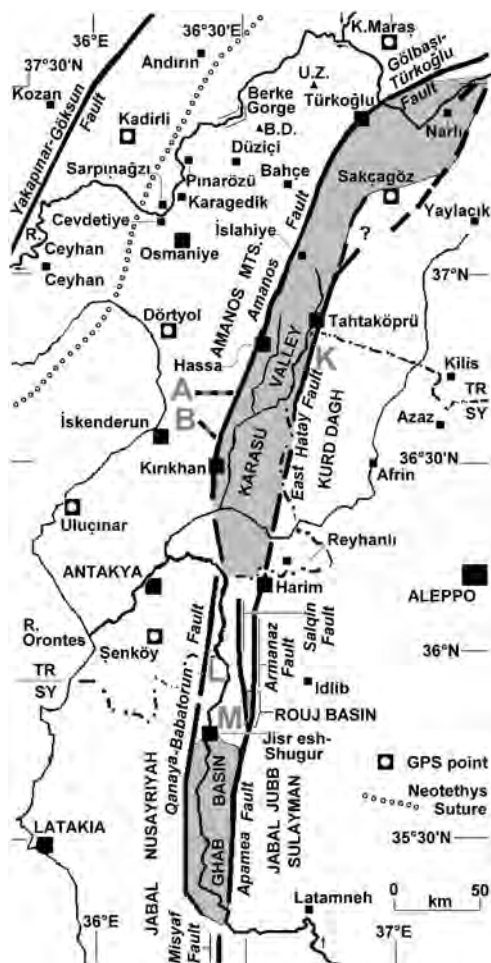


Fig. 2. More detailed map of the study region and its immediate surroundings. Grey shading schematically indicates lowlands forming linear valleys along the DSFZ. The southward continuation of the Karasu Valley, south of Kırıkhan, is called the Amik Basin, whereas its northward continuation is called the Narlı Plain or Aksu Plain, named after the River Aksu, which joins the Ceyhan west of Kahramanmaraş (abbreviated to K. Maraş). Only left-lateral faults considered active during the present phase of slip on the DSFZ are shown (see Fig. 1 for other faults). 'A' and 'B' denote the locations of Fig. 4a & b. 'K', 'L' and 'M' denote piercing points used to measure the total slip on DSFZ strands, as explained in the text. 'B.D.' and 'U.Z.' denote Beşikdülülü and Uluziyaret, two of the highest summits in the Amanos Mountains (2246 and 2268 m a.s.l., respectively).

& Pamir 1975; Brinkmann 1976, p. 89), its folding being thought to long predate the neotectonic phase. However, this mountain range is aligned along the northern DSFZ (Fig. 2) and contains anticline

axes that are also subparallel to this structural trend (Fig. 4). Such observations raise the question of whether some, at least, of this structure might instead relate to the active transpression.

Geological maps of the Karasu Valley have been published many times before (e.g. Çapan *et al.* 1987; Arger *et al.* 2000), and so will not be reproduced here. There are many localities where basalt flows have become offset left-laterally by slip on the Amanos Fault. Dating these basalts can thus indicate the vertical and horizontal slip rates on this fault, thereby addressing each of the above issues. Yurtmen *et al.* (2002) undertook this task using high-precision unspiked K–Ar dating, but found it more difficult than it might at first appear, due to the complexity of this volcanism (and thus the number of possible across-fault flow correlations); inconsistencies and site ambiguities affecting earlier low-precision spiked K–Ar dating; and the non-availability of large-scale topographic maps for much of the region, due to its proximity to the Syrian border (which limited location accuracy and thus the precision to which offset distances could be measured). Yurtmen *et al.* (2002) thus estimated the slip rate on the Amanos Fault as no more than *c.* 1.6 mm/a.

Yurtmen *et al.* (2002) also identified additional localities where future dating could add to constraints on the Amanos Fault slip rate. In the meantime, Tatar *et al.* (2004) published three new spiked K–Ar dates and determined the magnetic polarity of the basalt at many sites in the Karasu Valley. The Yurtmen *et al.* (2002) study predated the realization that the northern DSFZ is transpressive, and so made no attempt to test any kinematic model incorporating transpression. Furthermore, as Westaway (2004) noted, the slip rates, after Yurtmen *et al.* (2002), summed across all documented active left-lateral faults in and around the Karasu Valley, have hitherto significantly underestimated the predicted rate of AF–AR relative motion. Evidently, either additional active faults have remained undocumented, or the existing evidence has led to underestimation of slip rates. The lack of large-scale maps has in the meantime been resolved, due to topographic imagery derived from the Shuttle Radar Topographic Mission (SRTM) and the use of hand-held GPS receivers for location in the field. Along with the importance of the subject matter, these factors justify revisiting the area studied by Yurtmen *et al.* (2002).

Prior to the Yurtmen *et al.* (2002) study, the kinematics of the left-lateral faulting along the Karasu Valley were poorly determined, due to problematic dating (using the spiked K–Ar method, which typically results in large margins of uncertainty for Middle or Late Pleistocene dates) and imprecise documentation of field localities. A great diversity

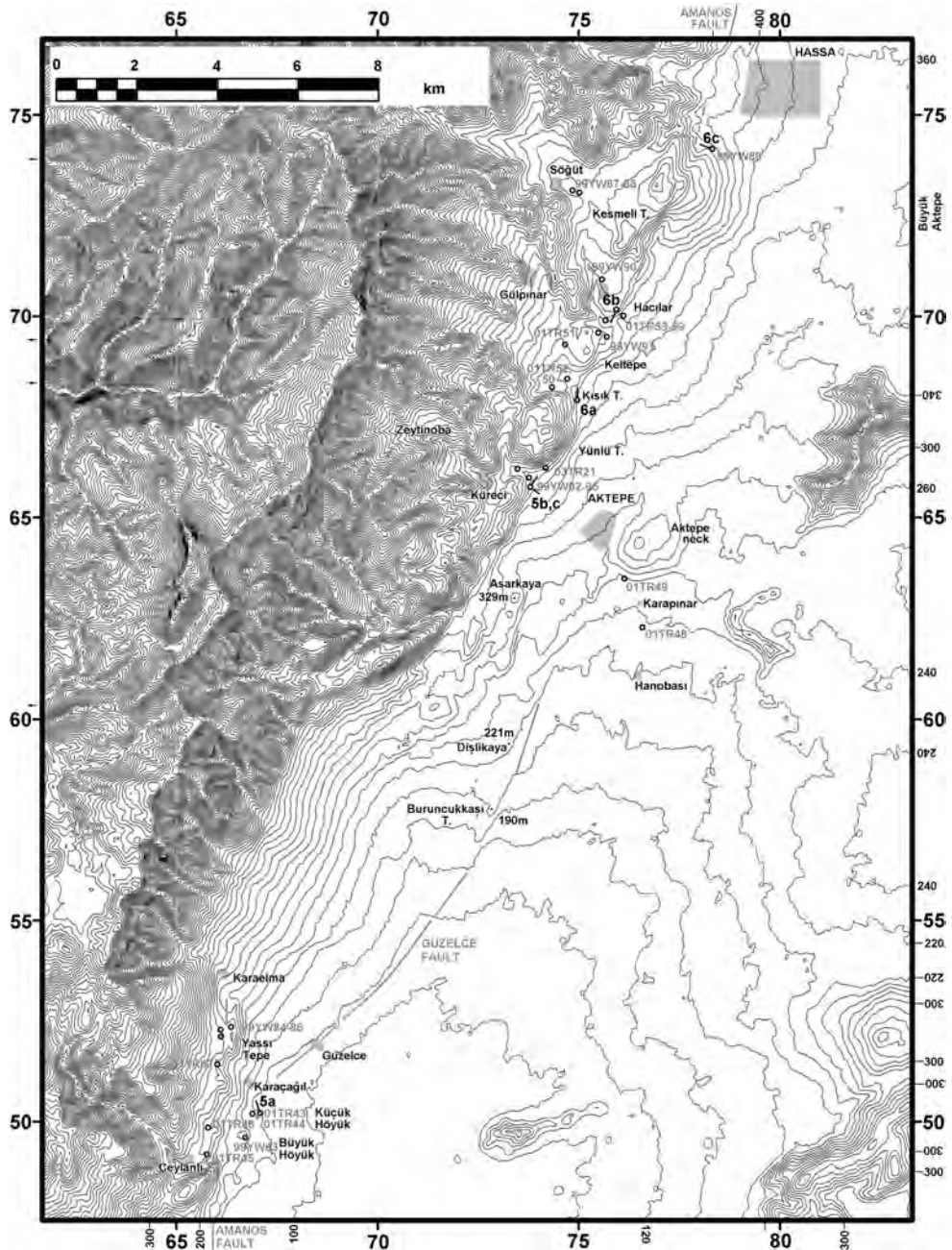


Fig. 3. Topographic map of the present study region, with a 20-m contour interval, based on SRTM data, showing the locations of basalt sampling sites from this study and Yurtmen *et al.* (2002), basaltic necks, and the positions of the Amanos and Güzelce faults. The position of the Amanos Fault is marked with ticks at the edges of the figure, in the Yassı Tepe and Hacilar areas, after Yurtmen *et al.* (2002). Another strand of this fault may run along the foot of the basalt bluff overlooking Karacagil, as illustrated (A–B) in Figure 5a. In many other localities the position of the fault is obvious from the range-bounding topography. Elsewhere, notably north of Karaelma, the line of the fault is unclear. Shaded areas next to place-names indicate the approximate extent of the named locality. Arrows indicate rivers draining westward, not eastward into the Karasu Valley. The positions of their headwaters indicate the close proximity of the drainage divide to the eastern margin of the Amanos Mountains. See Westaway *et al.* (2006a) for details of data sources and preparation technique.

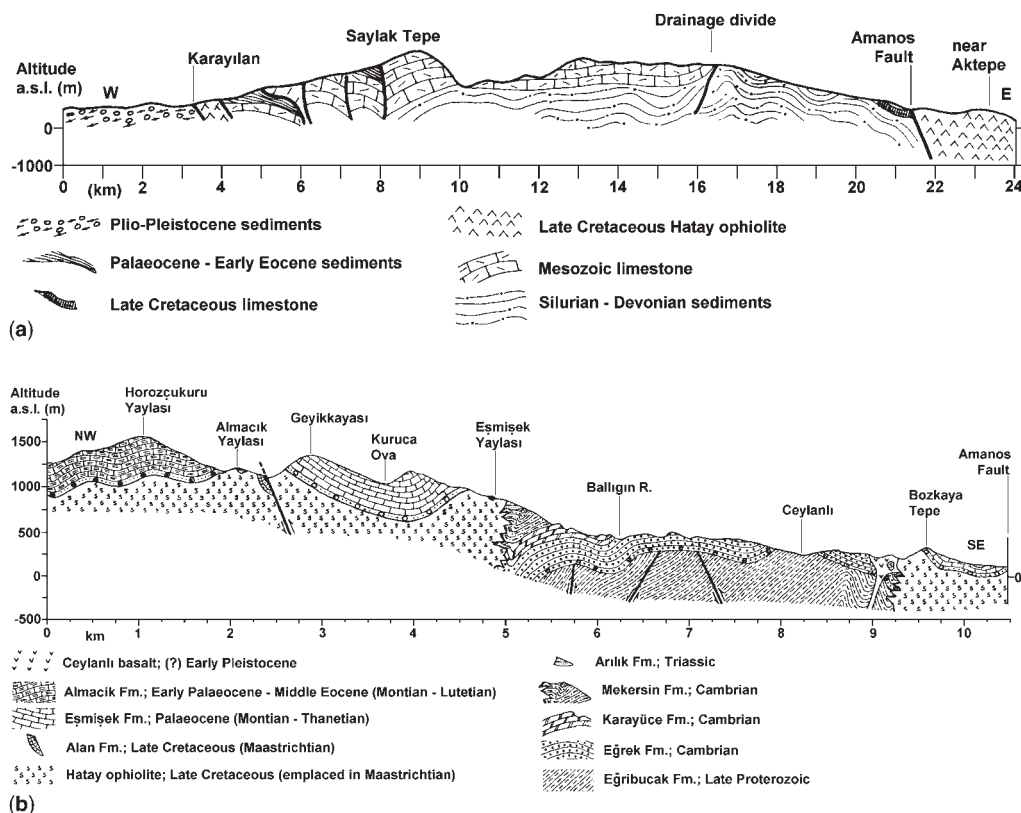


Fig. 4. Cross-sections illustrating the structure of the Amanos Mountains; see Figure 2 for locations.

(a) West-east profile along c. 36.7°N, adapted from cross-section II in plate II of Tolun & Pamir (1975). (b) NW-SE profile through Ceylanlı, adapted from cross-section VII in plate I of Atan (1969). The Maastrichtian-Eocene formations depicted are carbonate-dominated; typical lithologies of the older formations are summarized in Figure 12.

of views indeed existed, a popular view (recently restated by Mart *et al.* 2005) being that the Karasu Valley is a graben, bounded to the WNW and ESE by active normal faults (e.g. Schwan 1971). However, although there are obvious topographic escarpments along the margins of this valley (Figs 3 & 5), these are now thought to reflect minor components of vertical slip on predominantly left-lateral faults. Nonetheless, along most of the length of the valley there is little or no actual exposure of any fault surface; this is evidently because for thousands of years the landscape has been degraded by agriculture, and because no large earthquake has occurred for almost two centuries. Yurtmen *et al.* (2002) reviewed the diverse range of hypotheses that became superseded by their study, so such discussion is not repeated here. Tatar *et al.* (2004) have in the meantime proposed that the Karasu Valley is affected by pervasive distributed deformation between the bounding active left-lateral faults. We discount this possibility, as our re-analysis (discussed below)

reveals no requirement for any such complexity; the palaeomagnetic declinations that they have measured presumably reflect secular variation of the geomagnetic field rather than vertical-axis rotations.

The layout of this paper is as follows. First we discuss our fieldwork and laboratory analysis. Second, we use the information gained to estimate the slip rates on the Amanos Fault at the localities that have been studied. Finally, we compare this data-set with other evidence, in order to establish the extent of kinematic consistency for the DSFZ as a whole.

Field and laboratory procedures

Fieldwork

Four basalt samples were collected for Ar/Ar dating from critical localities previously identified by Yurtmen *et al.* (2002), illustrated in Figures 5a & b and 6a & b and located in Figure 3. We also

collected 14 other basalt samples for geochemical analysis, with the aim of using their geochemistry to correlate fragments of offset basalt flows (cf. Yurtmen *et al.* 2002). To facilitate accurate location, Universal Transverse Mercator (UTM) coordinates of all sample sites and other key localities (referenced to the WGS84 reference frame) were measured using a hand-held GPS receiver. UTM coordinates of some of the sample sites of Yurtmen *et al.* (2002) were likewise measured using GPS, in order to supplement the position fixes on the local maps used in that earlier study (fieldwork for which, in 1999, lacked the benefit of a working hand-held GPS receiver). This check was possible because one person (R.W.) participated in all fieldwork and, in 2001, he could precisely identify sites where he had worked in 1999. The UTM coordinates of each site determined in this manner were found to be typically *c.* 150 m south of the coordinates measured from the 1:25 000-scale maps, due to the different reference frame used in the production of these maps (cf. Westaway *et al.* 2004).

Ar/Ar dating

Initial screening involved inspection of hand specimens and petrographic thin sections. Samples were then crushed, washed in de-ionized water and dilute hydrochloric acid, sieved to a 60–80 μm size fraction, and phenocrysts and xenocrysts were removed by magnetic separation and hand picking. Following irradiation, argon isotopes were measured in the microcrystalline groundmass at the Laboratory for Noble Gas Geochronology, Massachusetts Institute of Technology, using procedures essentially the same as those described by Harford *et al.* (2002). Results are summarized in Table 1 and Figure 7.

All four Ar/Ar dates are tightly constrained, with $\pm 1\sigma$ errors ranging from $<1\%$ for the oldest sample (01TR47) to $<5\%$ for the youngest (01TR53). The MSWD, a measure of the internal consistency between the step-heating splits that contribute to each date, is <1 for three of the dates, and between 1 and 2 for the remaining date (Fig. 7). The extent of concordance between these new dating results and existing age control evidence, and its implications for estimation of the slip rate on the Amanos Fault, will be discussed below.

Geochemical analysis and interpretation

Geochemical analysis of whole-rock basalt samples, following the procedure used by Yurtmen *et al.* (2002), utilized the automated ARL 8420 X-ray fluorescence spectrometer at Keele University (Table 2). Figure 8 shows the classification of these basalts according to their alkali and silica content. As previously noted

(e.g. Çapan *et al.* 1987; Yurtmen *et al.* 2002), these basalts are compositionally diverse as a group, but most individual flow units show only limited variation. Most samples are basalt *sensu stricto*, but a few contain $>5\text{ wt}\%$ of alkali metal oxides and so classify as hawaiites. Sample 99YW83 of Yurtmen *et al.* (2002), from Büyük Höyük neck, in this scheme is classified as a mugearite.

No sample contains normative quartz; all samples contain normative olivine and some contain normative nepheline (Table 2). These basalts are thus classified, according to their normative composition, as olivine tholeiites and alkali olivine basalts (cf. Yurtmen *et al.* 2002). The most mafic flow unit (with the least SiO_2 and the most normative nepheline) is at Hacilar (cf. Yurtmen *et al.* 2002), represented in Figure 8 by samples 99YW90, 01TR53, 01TR54 and 01TR59. As Yurtmen *et al.* (2002) noted, as a result of the left-lateral slip on the Amanos Fault, this flow unit is juxtaposed against much less mafic basalts, which they thought originated from necks at Keltepe or Kısıc Tepe farther south (Fig. 3). Combined with dating evidence, this contrast in geochemistry is useful for constraining the kinematics of this fault (Yurtmen *et al.* 2002; see below). The next most mafic basalt seems to be that from the large flow unit in the central Karasu Valley, which originated from the Büyük Aktepe neck in the north (Fig. 3), flowing southward for $>20\text{ km}$ past the town of Aktepe (e.g. Çapan *et al.* 1987), and was studied in detail by Polat *et al.* (1997). We sampled this basalt SSE of Aktepe (sample 01TR48; Fig. 8), confirming its mafic characteristics. It can indeed be readily distinguished from other basalts in the vicinity, such as that from the nearby Aktepe neck, represented by sample 01TR49 (e.g. using the much lower SiO_2 and much higher MgO contents; Table 2).

In terms of trace-element geochemistry (Table 2), the Karasu Valley basalts have many characteristics of ocean island basalt (cf. Polat *et al.* 1997; Yurtmen *et al.* 2002), such as high abundances of the light rare-earth elements La (up to 36 ppm) and Ce (up to 80 ppm). They also contain high abundances of incompatible trace elements (e.g. up to 853 ppm Sr) and other characteristics (e.g. low Zr/Nb ratios of *c.* 4) considered indicative of small-degree partial melting of the asthenosphere (such as may occur in ocean islands due to the heating effect of a mantle plume). However, these basalts have clearly not formed in an ocean-island setting. Yurtmen *et al.* (2002) proposed instead that this volcanism has resulted from heating of the mantle lithosphere (due to the increase in Moho temperature that accompanies crustal thickening), causing the remelting of products of



Fig. 5. *Continued.*

earlier small-degree partial melting of the asthenosphere, which had frozen within the mantle lithosphere. Our new data are consistent with this explanation. The transpressive strike-slip faulting now evident in this region provides a mechanism for the required crustal thickening, and can thus be regarded as a potential cause for the volcanism.

In detail, as Yurtmen *et al.* (2002) noted, the Karasu Valley basalts have resulted from the combination of variable (but small) degrees of partial melting of the asthenosphere, followed by variable fractional crystallization and variable (but small) degrees of crustal contamination. Crustal contamination is evident in some samples, for instance from the Ba/La ratios of $\gg 10$ (Table 2). The interplay between effects of partial melting and subsequent fractional crystallization can be illustrated (after Pearce *et al.* 1990) by the variation of Cr against Y (Fig. 9). The compatible element Cr is taken up during early crystallization of basaltic melt (such as may occur at depth in a magma chamber) and so becomes depleted in the residual melt. Based on this criterion, the most 'primitive' magmas (i.e. with the least evidence of fractional crystallization) are those from the large Büyük Aktepe flow unit, represented by our sample 01TR48 and other samples from Polat *et al.* (1997). Conversely, the most evolved magmas are represented by the low-volume flow units at Büyük Höyük (sample 99YW83), Yassı Tepe (samples 99YW84–86), and Küreci (samples

99YW92–95 and 03TR21). This tendency for high-volume flows to have experienced much less fractional crystallization than low-volume flows is the opposite of what has been reported elsewhere (cf. Thompson *et al.* 1990).

The greatest local complexity in the basalt stratigraphy is evident around the left-laterally offset Hacilar river gorge (Figs 5b, 10 & 11). Our seven samples from this area (01TR53–9) reveal three chemically distinct flow units. Sample 01TR53 came from the uppermost basalt flow unit forming the top of the bluff (Fig. 10b) immediately west of the fault at the southern end of this offset reach. Sample 01TR54 came from what appears to be the top of a lower basalt flow within the same flow unit, directly underneath it. Sample 01TR59 was collected east of the fault, from the uppermost basalt flow there, which is observed to cascade over this bluff at an angle of *c.* 25° (Fig. 10b). These three samples are geochemically very similar to each other (Table 2 and Figs 8 & 9) and to sample 99YW90 of Yurtmen *et al.* (2002); they represent the Hacilar basalt.

Sample 01TR58 came from a flat-lying basalt flow, stratigraphically below the one which yielded sample 01TR59, just above river level east of the Amanos Fault (Fig. 10b). We have not observed any corresponding flat-lying basalt in the bluff west of the fault, suggesting that this flow is not of local provenance. This sample is most geochemically similar to sample 01TR50 (Fig. 8),

Fig. 5. (a) View NNW from the NE margin of the Küçük Höyük neck at [BA 67099 50157], looking past Karaçağıl to the line of the Amanos Fault at Yassı Tepe, *c.* 2 km away. Yurtmen *et al.* (2002) interpreted this fault, running northward from A, before following a gully on the far side of the Yassı Tepe basalt ridge (which was thus interpreted as a shutter ridge) behind the minaret of Karaçağıl mosque. North of this locality, this interpreted fault is not visible in this view. According to Yurtmen *et al.* (2002), the *c.* 428 ka Yassı Tepe basalt is offset left-laterally by an estimated *c.* 425 m by this fault. This Yassı Tepe basalt overlies more extensive older basalt, offset left-laterally by an estimated *c.* 1000 m, that Tatar *et al.* (2004) showed to be reverse-magnetized and we have dated to *c.* 1195 ka. The line A–B, projecting between Karaçağıl and the near face of Yassı Tepe, may mark a subparallel fault, with a component of downthrow to the east. The contact C–D, between basalt and Karasu Valley alluvium, typically marked in this area by a subdued bluff, is interpreted as the surface trace of Guzelce Fault. Note the concordant summits, locally at *c.* 1500 m a.s.l., of the Amanos mountain range in the background. (b) View northward along the Amanos Fault from [BA 73728 65812], on the southern side of the outlet of the Olucak river valley. The limestone mountain on the far side of the valley is Yünlü Tepe. Directly in front of it, near the left edge of the photo, is the downstream end of the basalt flow unit within this valley (P marks its top surface), which is truncated by the Amanos Fault at [BA 73872 66180] (Q). Basalt samples 99YW94 and 99YW95 were collected near river level below this basalt bluff. The line of this fault projects from this bluff into the distance along the escarpment bounding Yünlü Tepe (Q–R). To the right of this is a flat area formed of fluvial gravel, behind and to the right of which (R–S) is the outcrop of basalt, starting around [BA 74151 66226], from which sample 03TR21 was collected. On the right of the photo, at an azimuth of N30°E from the viewpoint, this basalt is observed to be truncated at a second bluff (S), suggesting that another strand of the Amanos Fault may be present. S–T and Q–U indicate the interpreted lines of these two faults as they approach the viewpoint. (c) View ESE across the Karasu Valley from the same place as (b). In the middle distance is the town of Aktepe. Behind it (with trees on the summit) is the Aktepe volcanic neck. Behind that is another hill formed by an inlier of the Hatay ophiolite protruding through the basalt and alluvium in the Karasu Valley floor. In the background, rising to *c.* 1000 m a.s.l. and *c.* 18 km away, is the Kurd Dag mountain range in NW Syria, also formed largely of the Hatay ophiolite and associated Mesozoic sediments. The international border runs along the foot of this mountain range, being demarcated by the Karasu River, which also marks the line of the East Hatay Fault (Y–Z).

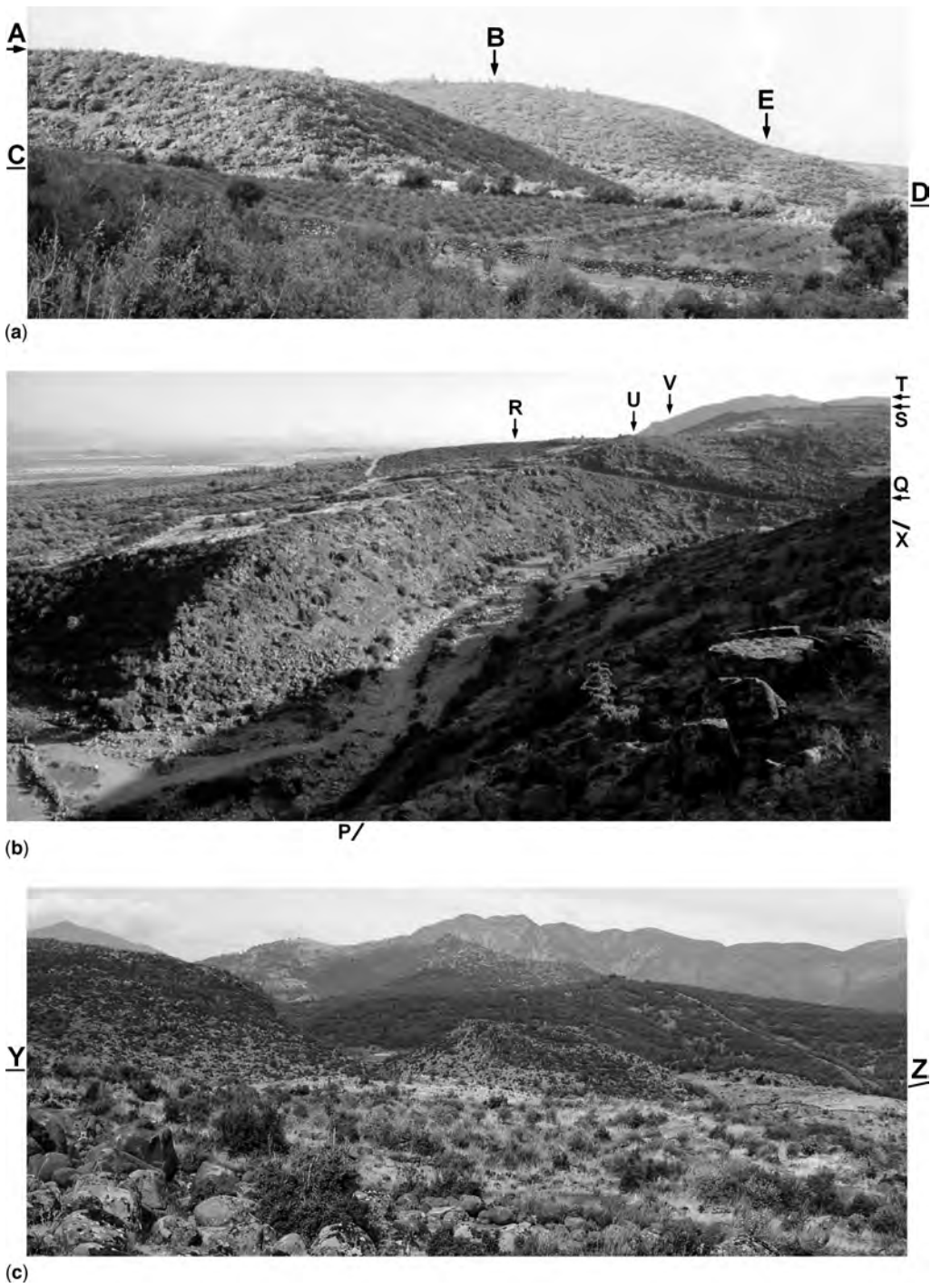


Fig. 6. *Continued.*

from Kısık Tepe, west of the Amanos Fault, c. 1.8 km farther south. We thus suggest that left-lateral slip has juxtaposed this (sample 01TR50) basalt to its present position.

The remaining three samples were collected from basalt flows just above river level, west of the Amanos Fault. Sample 01TR55 came from a point c. 25 m south of the stone bridge over the Kurtlan tributary of the Hacilar River, just upstream of the offset reach of the latter; the other two came from the places indicated in Figure 10, where they stratigraphically underlie the Hacilar basalt. The greatest overall geochemical similarity of these samples is with sample 99YW89 of Yurtmen *et al.* (2002), which was collected c. 4.2 km farther north, east of the Amanos Fault. Yurtmen *et al.* (2002) inferred that sample 99YW89 erupted from the Kesmeli Tepe neck north of Hacilar on the west side of the fault (Fig. 11). We presume that another part of the same flow unit flowed southward from its source to the fault line in the Hacilar area, and yielded our three samples (01TR55, 01TR56 and 01TR57), being later buried beneath the younger Hacilar basalt. At the points where samples 01TR56 and 01TR57 were collected, the basalt is observed to dip ESE at c. 40°, suggesting that it cascaded across a contemporaneous fault-line bluff. However, corresponding basalt is not found locally east of the fault; it has presumably been displaced farther NNE by the left-lateral slip (see below).

Constraints on horizontal crustal motions, from dating and basalt geochemistry

Hacilar

As noted above (Figs 6b, 10a & b), at Hacilar, the Hacilar river has incised into the young Hacilar

basalt (which flowed down this river valley, and is itself offset left-laterally; see above) on both sides of the Amanos Fault; thus, dating the basalt indicates the timescale on which this left-lateral offset has developed. According to Yurtmen *et al.* (2002) this basalt erupted from Camlı Tepe (Fig. 11), SE of Söğüt. This massive basalt shows minimal weathering, indicating a young age. As Yurtmen *et al.* (2002) noted, the c. 1.1 Ma whole-rock K–Ar date by Çapan *et al.* (1987) at a site somewhere around Hacilar indicates either inherited argon in their sample or that an older basalt flow was dated (or both), and thus has no bearing on estimation of the slip rate from the river offset.

Yurtmen *et al.* (2002) obtained an unspiked K–Ar date of 196 ± 12 ka ($\pm 2\sigma$) for the youngest basalt exposed in the river gorge section (their sample 99YW90), c. 1 km from the fault offset (Fig. 11). Rojay *et al.* (2001) obtained a spiked K–Ar date of 80 ± 120 ka ($\pm 2\sigma$) for the youngest basalt near this fault offset (their sample 08; Fig. 11), as the weighted mean of three splits, whose individual ages (all $\pm 2\sigma$) were reported as 130 ± 130 ka, 70 ± 72 ka, and 30 ± 60 ka. The Yurtmen *et al.* (2002) date is concordant with this overall date and with the oldest of these splits, suggesting that the same basalt flow has been dated at both sites. However, the high error margins of this spiked K–Ar dating limit its use in slip-rate determination.

Our new sample 01TR53, collected from the youngest basalt directly adjacent to the fault offset on its African side (Fig. 11), has yielded an Ar–Ar date of 159 ± 15 ka ($\pm 2\sigma$). This is concordant at the $\pm 3\sigma$ level with the Yurtmen *et al.* (2002) date, suggesting that the same basalt flow is present at both sites, consistent with the observation that no flow front can be seen between the sites. Yurtmen *et al.* (2002) estimated the offset of the

Fig. 6. (a) View north from [BA 74919 67956], looking at the escarpment that has formed along the eastern side of Kısık Tepe (A) by the component of downthrow to the east on the Amanos Fault. Behind Kısık Tepe, the southern part of the Keltepe basalt is also visible (B). The surface trace of the Amanos Fault is locally interpreted along the base of the escarpment (C–D). A less-pronounced break of slope within the Keltepe basalt (E) may mark a subsidiary structure. (b) View SSW from c. [BA 759 751], at the point where the road from Aktepe to Hacilar reaches the top of the Amanos Fault escarpment. The reach of the Hacilar River that has become offset left-laterally by slip on this fault (P–Q) is visible in the foreground. The basalt in the right foreground (in shadow) is part of the Hacilar flow unit that we have dated to 159 ± 15 ka ($\pm 2\sigma$). The basalt in the middle distance, beyond the offset river reach to the east of the fault (R), yielded sample 99YW91 dated to 476 ± 16 ka ($\pm 2\sigma$). Beyond it on the western side of the fault, the near skyline (S) is the outline of Keltepe neck, and the far skyline (T) is the north face of Yünlü Tepe (cf. Fig. 5b). South of the Hacilar River, the Amanos Fault can be identified from the alignment of scarps along the faces of Yünlü Tepe (U) and Keltepe (V), which project to the river at a similar break of slope along the line U–V–X, as Yurtmen *et al.* (2002) noted. (c) View WNW from [BA 78338 74184], near the point where basalt sample 99YW89 was collected, looking at the escarpment formed by the Amanos Fault. The basalt in the foreground, east of this fault, is inferred to have erupted from Kesmeli Tepe, farther south on the opposite side of the fault. This fault escarpment is formed in Upper Cretaceous limestone, with no basalt locally present on the far side, consistent with the development of significant left-lateral displacement since the basalt erupted. Some blocks of basalt have been recently dumped in the foreground, as a result of land clearance elsewhere, but *in situ* basalt is also present.

Table 1. Summary of Ar–Ar dating results for Karasu Valley basalts

Sample	Locality	UTM coordinates	Laboratory no.	$^{40}\text{Ar}(r)/^{39}\text{Ar}(k) (\pm 2\sigma)$	Age (ka) $(\pm 2\sigma)$	MSWD	$^{39}\text{Ar}(k) (\%)$	K/Ca	n/N
01TR53	Hacılar	BA 75632 69900	B20 RW5F041P	$0.2906 \pm 0.0266 (\pm 9.16\%)$	$158.7 \pm 14.6 (\pm 9.18\%)$	0.71	79.37	0.357 ± 0.005	7/10
03TR21	Küreçli	BA 74151 66226	B17 RW5F041M	$0.6796 \pm 0.0224 (\pm 3.29\%)$	$385.2 \pm 12.9 (\pm 3.35\%)$	0.24	81.02	0.285 ± 0.004	7/10
01TR51	Kısıklı Tepe	BA 74511 69310	B19 RW5F041O	$0.7446 \pm 0.0163 (\pm 2.20\%)$	$411.8 \pm 9.4 (\pm 2.28\%)$	0.24	92.80	0.530 ± 0.007	7/10
01TR47	Karaçagıl	BA 66092 51365	B18 RW5F041N	$2.1355 \pm 0.0252 (\pm 1.18\%)$	$1195.1 \pm 15.8 (\pm 1.32\%)$	1.64	90.74	0.453 ± 0.006	6/10

All samples were of basaltic groundmass, separated as described in the text. $^{40}\text{Ar}(r)/^{39}\text{Ar}(k)$ is the measured ratio of the number of ^{40}Ar atoms produced by radioactive decay of the potassium in the sample to the number of ^{39}Ar atoms produced by irradiation of the potassium in the sample. MSWD is the mean squared weighted deviation, a measure of the scatter between individual age estimates for different heating steps. $^{39}\text{Ar}(k) (\%)$ is the percentage of the ^{39}Ar in the irradiated sample that is determined to have been produced by irradiation of potassium. K/Ca is the measured ratio of potassium to calcium atoms in the sample, which is used to correct the ^{39}Ar measurements for ^{39}Ar produced by irradiation of calcium. N is the total of heating steps measured; n is the number of these steps that were used for each age determination. All ages have been calibrated relative to the US Geological Survey Taylor Creek Rhyolite tr-2a sanidine standard, with an age of 28.34 Ma (after Renne *et al.* 1998). Age calculations assume the decay and isotopic abundance constants from Steiger & Jäger (1977).

Hacılar River by the Amanos Fault (Figs 6b & 10a) as 325 ± 25 m. Dividing this offset by our new age for this basalt gives an estimate for the local slip rate of 2.05 ± 0.37 mm/a $(\pm 2\sigma)$. However, rather than being aligned with its present course on the upstream side of the Amanos Fault, it is possible that immediately after the basalt eruption the Hacılar River flowed along what is now the extreme southern margin of its valley, as such a position would have coincided with the southern margin of the basalt flow. If so, the present offset of this river underestimates its true offset, which we estimate as 450 ± 25 m (Fig. 11). With such a slip measurement, the estimated slip rate on the Amanos Fault adjusts to 2.84 ± 0.41 mm/a $(\pm 2\sigma)$.

Hassa

As Yurtmen *et al.* (2002) noted, study of the basalt south of Hassa has been difficult, due to the unavailability of the local 1:25000-scale topographic map sheet. The SRTM coverage now available (Fig. 11) makes much more precise analysis possible.

Yurtmen *et al.* (2002) obtained an unspiked K–Ar date of 393 ± 22 ka $(\pm 2\sigma)$ for a sample (99YW89) collected near the northern margin of the basalt outcrop south of Hassa, just east of the Amanos Fault (Figs 6c & 11). We established its coordinates as [BA 78338 74184]; the margin of the basalt being an estimated *c.* 100 m farther north.

The SRTM imagery (Fig. 11) now reveals that this basalt forms a substantial edifice, resembling half a cone, with a radius of *c.* 1.5–2 km. In its centre, the basalt surface (reaching *c.* 605 m a.s.l.) is now juxtaposed against a hill, Kalecik Tepe (summit *c.* 665 m a.s.l., *c.* 300 m west of the fault), formed in Late Cretaceous limestone, such that by chance there is locally no east-facing scarp across the Amanos Fault. West of the Amanos Fault, the highest point with a basalt outcrop is Kesmeli Tepe (*c.* 615 m a.s.l. from both the local map and the SRTM data in Fig. 11), *c.* 2 km NNE of Hacılar. Yurtmen *et al.* (2002) tentatively interpreted this as the neck from which the basalt erupted. However, allowing for subsequent downthrow to the east on the Amanos Fault, it is possible that the neck was east of the fault (at the point now east of Kalecik Tepe) from which basalt flowed westward on to what is now Kesmeli Tepe. Basalt seems to have also flowed WNW and northward from the Kesmeli Tepe area, ponding the Hayıtlı River valley to create the alluvial plain evident around the village of Söğüt. This Kesmeli Tepe basalt also forms the land surface westward for *c.* 300 m and southward for *c.* 700 m from Kesmeli Tepe on the African side of the Amanos Fault; beyond these points it has been truncated by

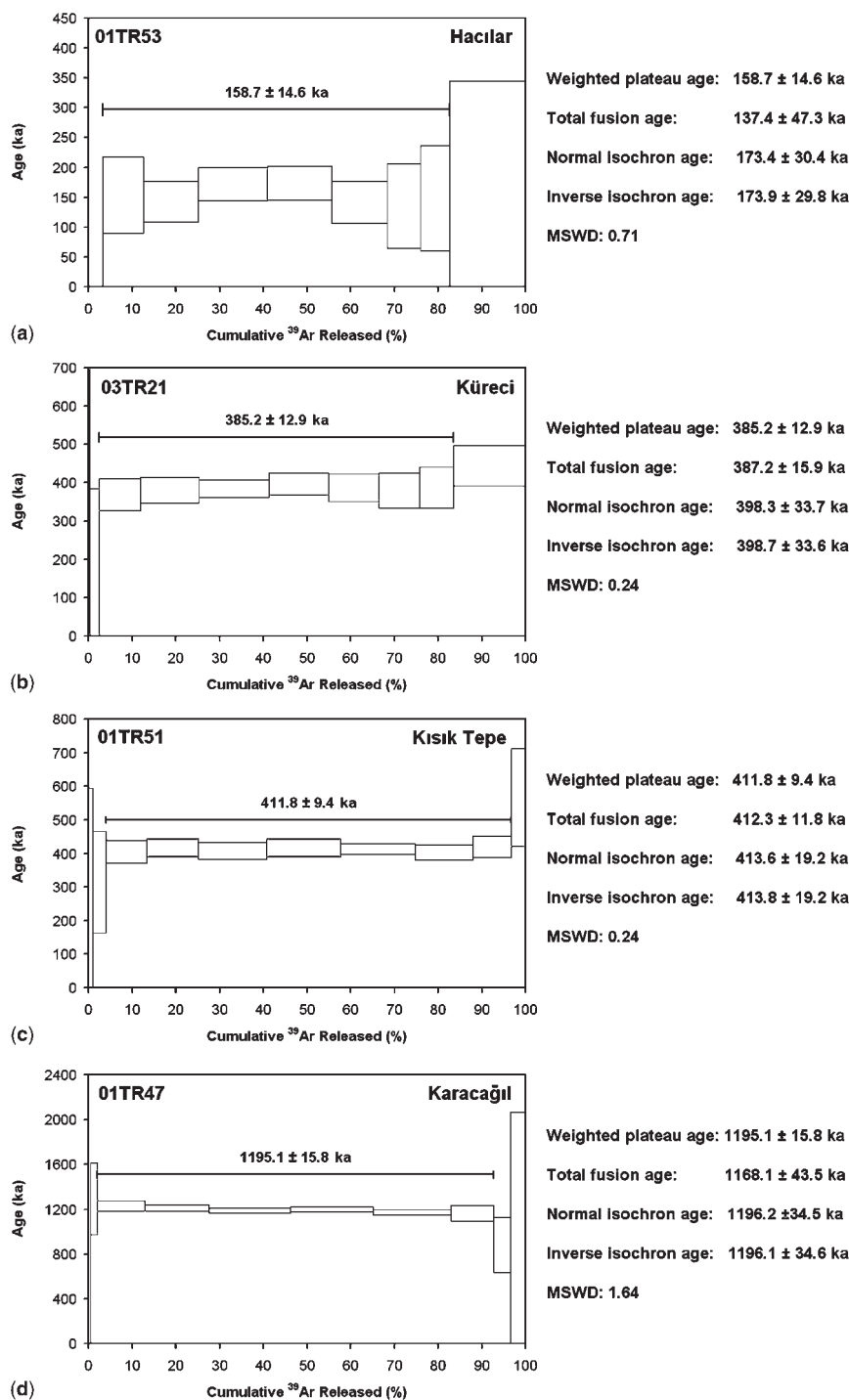


Fig. 7. Ar/Ar age spectra derived from step heating of our four samples. The vertical bars indicate the uncertainty in age derived from each heating step for each sample. The horizontal bars indicate the range of heating steps that contribute to each date. To the right of each diagram the age determinations for the sample are listed, using four different methods, along with the MSWD values (Table 1). The extent of concordance between these different age determinations provides a test of the reliability of each date. For full details see the online supplement at <http://www.geolsoc.org.uk/sup18274>. A hard copy can be obtained from the Geological Society Library.

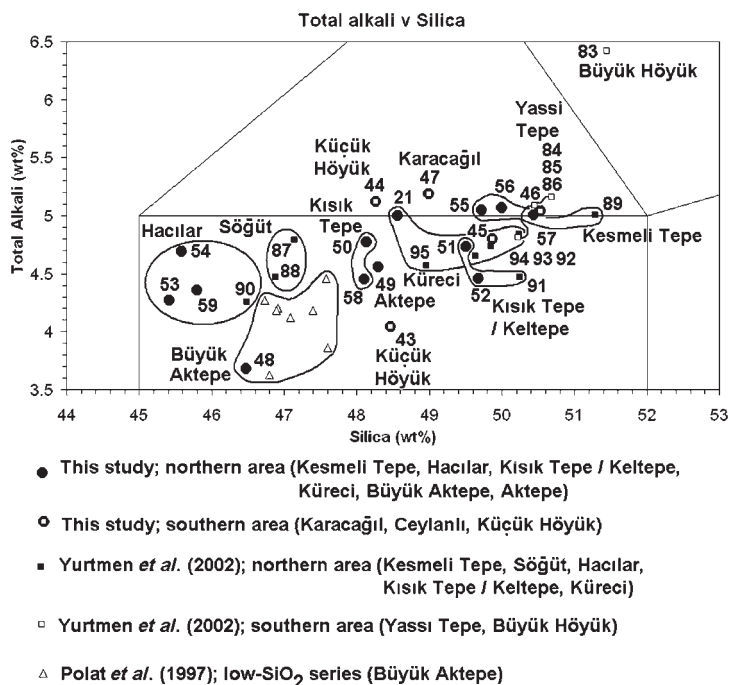


Fig. 8. Total alkali (Na₂O + K₂O) v. SiO₂ diagram for Karasu Valley basalts (after Le Bas *et al.* 1986). Samples from Ceylanlı are unlabelled. Samples are labelled in this figure and Fig. 9 using the last two digits of their sample numbers.

fluvial incision and then inset by the younger Hacilar basalt. However, our interpretation of samples 01TR55-7 (see above) suggests that this Kesmeli Tepe basalt persists beneath the Hacilar basalt, west of the Amanos Fault, to a distance of c. 2.5 km south of Kesmeli Tepe.

Restoring the highest point of the basalt cone east of the Amanos Fault (c. [BA 77455 78080]; L in Fig. 11) against Kesmeli Tepe (c. [BA 76355 76910]; 'M' in Fig. 11) indicates 1600 ± 25 m of subsequent left-lateral slip on this fault. The large volume of basalt produced suggests that this neck was probably active for many tens of thousands of years. Sample 99YW89 came from one of the most distal parts of this flow unit, and in the field it can be seen to almost directly overlie bedrock. However, it cannot be presumed to indicate the start of eruption from this neck; it may simply mark the greatest eruption rate, when basalt flowed farthest from its source. The data-set presented by Yurtmen *et al.* (2002) indicated a concentration of dates around 400–500 ka. The oldest of this group of dates, from Küreci (c. 5 km SSW of Kesmeli Tepe), was 553 ± 20 ka ($\pm 2\sigma$; see below). If this is taken as also marking the start of the Kesmeli Tepe eruption, then the subsequent time-averaged slip rate on the Amanos Fault can be estimated as 2.89 ± 0.14 mm/a ($\pm 2\sigma$).

Küreci

Our new Ar/Ar date for sample 03TR21, of 385 ± 13 ka ($\pm 2\sigma$), supplements the data already available from the Küreci area (Fig. 3). As Yurtmen *et al.* (2002) discussed, basalt erupted from a neck near Zeytinoba and flowed eastward down the Küreci river valley, past Küreci village and into the Karasu Valley, crossing the Amanos Fault (Fig. 12). Çapan *et al.* (1987) obtained a spiked K–Ar date of 600 ± 200 ka ($\pm 2\sigma$) for this basalt just east of Küreci (their sample 18). Rojay *et al.* (2001) added another such date of 190 ± 100 ka ($\pm 2\sigma$) (their sample 07) for the top of the basalt just west of the Amanos Fault. Yurtmen *et al.* (2002) obtained three unspiked K–Ar dates: 553 ± 20 ka ($\pm 2\sigma$) for the base of the c. 30-m-thick flow unit east of Küreci (sample 99YW92), 325 ± 14 ka ($\pm 2\sigma$) for the top of this flow unit, also east of Küreci (sample 99YW93), and 253 ± 14 ka ($\pm 2\sigma$) for basalt in the modern valley floor adjacent to the Amanos Fault (sample 99YW94). Tatar *et al.* (2004) added another spiked K–Ar date, of 620 ± 260 ka ($\pm 2\sigma$) for the proximal part of this basalt near Zeytinoba (their sample 33); they also established that six samples were normally magnetized, evidently all dating from the Brunhes chron.

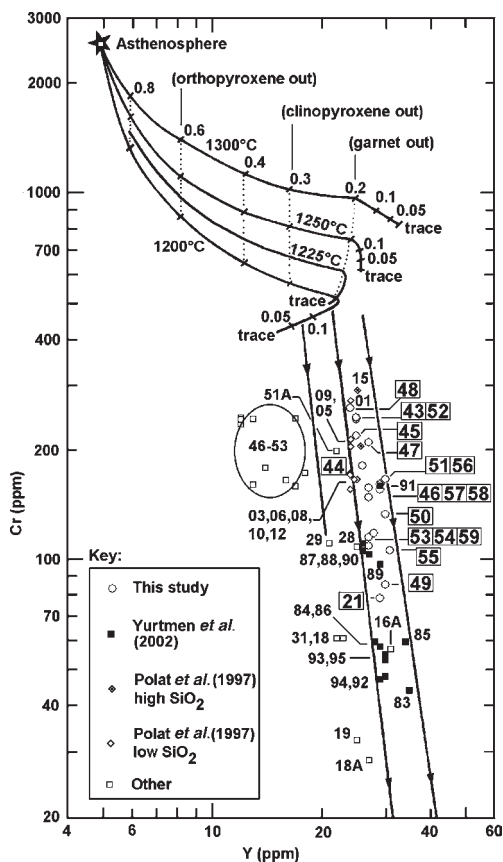


Fig. 9. Chromium v. yttrium diagram highlighting the roles of partial melting and fractional crystallization in the formation of the basalts in the study region. Solid lines are partial melting trends, calculated by Pearce *et al.* (1990) for garnet lherzolite (55% olivine, 20% orthopyroxene, 12.5% clinopyroxene, and 12.5% garnet) at 1200 °C, 1250 °C, and 1300 °C, with each solid phase disappearing at the indicated degree of partial melting. See Pearce *et al.* (1990) for the values of the partition coefficients used and for sources of data. The line for 1225 °C was estimated by Yurtmen *et al.* (2002) by interpolation. Barbed lines indicate mafic crystallization trends, also from Pearce *et al.* (1990).

Sample 03TR21 came from [BA 74151 66226], at the southern limit of basalt outcrop, just east of the Amanos Fault (point D in Fig. 12). Basalt is not present locally across the fault, where outcrop is instead of Late Cretaceous limestone (the Karadağ Limestone of Atan 1969). We measured the northern edge of the basalt in the Küreci valley, just west of the fault, at [BA 73872 66180] (point C in Fig. 12), near where Rojay *et al.* (2001) collected their dating sample (reported by them at [BA 73835 66305]).

Atan (1969) mapped a small outlier of basalt adjoining the Amanos Fault on its eastern side, NE of Yünlü Tepe (Fig. 12). From this map, the northern margin of this outcrop is offset by 1725 ± 50 m from its counterpart west of the Amanos Fault in the Küreci Valley (A–B in Fig. 12). Assuming that this outlier marks the oldest part of the Küreci flow unit, with the same age as our sample 99YW92, the subsequent time-averaged slip rate on the Amanos Fault can be estimated as 3.12 ± 0.21 mm/a ($\pm 2\sigma$). As Figure 12 indicates, this basalt outlier abuts a ridge of ophiolite on its southern side. Once the basalt outlier was displaced away from the Küreci Valley by left-lateral slip on the Amanos Fault, this ophiolite ridge would have been juxtaposed across the mouth of this river valley. It would have thus acted as a shutter ridge, inhibiting later-erupting basalt from flowing out into the Karasu Valley. Ponding of the Küreci basalt by this shutter ridge is presumably the reason why this basalt has attained such a substantial thickness west of the Amanos Fault and yet is virtually absent east of this fault. As noted above, our own fieldwork suggests that the piercing point where the Amanos Fault intersects the northern margin of the Küreci Valley is 'C', not 'B', in Figure 12, c. 100 m farther west. If so, then the fault offset adjusts to A–C or c. 1800 ± 50 m and the associated slip rate (making the same age assumption as before) adjusts to 3.25 ± 0.22 mm/a ($\pm 2\sigma$).

Dating the end of this basalt eruption is more problematic. At a relatively late stage, left-lateral slip displaced this ophiolite ridge north of the Küreci Valley, so basalt was again able to flow eastward into the Karasu Valley. However, the limited volume of basalt east of the Amanos Fault suggests that volcanism had more or less ended by this time. Currently, four dates (all $\pm 2\sigma$) potentially constrain the end of this volcanism: 190 ± 100 ka from Rojay *et al.* (2001); 325 ± 14 ka and 253 ± 14 ka from Yurtmen *et al.* (2002); and our new date of 385 ± 13 ka. These dates are not concordant; Yurtmen *et al.* (2002) discussed this issue but could offer no satisfactory explanation. Since all four dates would appear to reflect the youngest volcanism in this locality, it nonetheless seems appropriate to determine their weighted mean: 323 ± 8 ka ($\pm 2\sigma$).

We estimate that the site of our sample 03TR21 is 700 ± 50 m from the point where the Amanos Fault intersects the southern margin of the Küreci Valley ('G' in Fig. 12). This figure thus indicates the left-lateral slip on the fault strand between points D and G since this basalt erupted. Taking the age of this basalt as our weighted mean value of 323 ± 8 ka ($\pm 2\sigma$), we estimate a slip rate on the Amanos Fault of 2.17 ± 0.31 mm/a ($\pm 2\sigma$). As illustrated in Figure 12, a second active fault

strand appears to pass east of point D, forming the bluff on the skyline in the extreme right of Figure 5b. The slip rate on this fault can be tentatively estimated as *c.* 1 mm/a, i.e. the difference between the *c.* 3 mm/a slip rate derived from offset A–C (which represents the overall slip rate on the Amanos Fault north of the Küreci Valley) and the *c.* 2 mm/a slip rate derived from offset D–G (which represents the local slip rate on the western strand of the Amanos Fault).

Kısık Tepe

Yurtmen *et al.* (2002) drew attention to two basalt-capped hills, thought to represent necks, west of the Amanos Fault between Hacilar and Küreci, called Keltepe and Kısık Tepe (Figs 3 & 11). They also noted basalt east of the Amanos Fault south of Hacilar which was much more weathered than the nearby Hacilar basalt, and which yielded a much older unspiked K–Ar date of 476 ± 16 ka ($\pm 2\sigma$) (sample 99YW91). This basalt, which Yurtmen *et al.* (2002) also showed to be chemically distinct from the Hacilar basalt, reaches its maximum thickness east of the Amanos Fault *c.* 250 m south of the 99YW91 site, at Kocaören (*c.* [BA 75565 69485]; ‘H’ in Fig. 11). Yurtmen *et al.* (2002) suggested that the basalt in this area erupted from Keltepe (*c.* [BA 75255 69125]; ‘J’ in Fig. 11), from which it is now offset by *c.* 475 m. From the age of sample 99YW91 and this measured offset, they tentatively deduced that this segment of the Amanos Fault has a slip rate of 1.00 ± 0.14 mm/a ($\pm 2\sigma$).

To check this possibility, we collected three samples of basalt from west of the fault around Kısık Tepe and west of Keltepe. All three are geochemically similar to Yurtmen *et al.*’s (2002) sample 99YW91 (Table 2). Sample 01TR51, collected west of Keltepe, was also Ar/Ar dated to 412 ± 9 ka ($\pm 2\sigma$).

Local geological mapping (e.g. Atan 1969) indicates that the eruption of the Kısık Tepe and Keltepe basalts ponded the valleys of the Kızılyar and Kurtlan rivers, which flow eastward towards the Karasu Valley. It follows that both eruptions must have been roughly synchronous, otherwise no such ponding would have been possible. The Kurtlan River has subsequently incised through Cretaceous limestone bedrock, around the northern margin of the Keltepe basalt, to form a new course, joining the Hacilar River just west of the Amanos Fault (see Fig. 11, inset). The Kızılyar has instead cut a narrow gorge through the Kısık Tepe basalt adjacent to sample site 01TR52.

We consider it probable that the basalt forming our sample 01TR51 originated from Keltepe, rather than Kısık Tepe. However, the geomorphological argument above suggests that both necks

erupted around the same time, so this date can also estimate the timing of eruption of the Kısık Tepe basalt. In view of our earlier estimates for the slip rate on the Amanos Fault, we suggest that the basalt east of the Amanos Fault at Kocaören and at sample sites 99YW91 and 01TR58 (discussed above) is equivalent to that at Kısık Tepe, not Keltepe. Given the greater thickness of basalt and surface altitude (*c.* 465 m at Kocaören; *c.* 435 m at Kısık Tepe; *c.* 440 m west of Kısık Tepe around sample site 01TR50) it is indeed possible that this basalt erupted from a neck at Kocaören, from which it flowed westward across the Amanos Fault into the Kısık Tepe area. Since Kısık Tepe (*c.* [BA 74930 68450]; ‘K’ in Fig. 11) is *c.* 750 m SSW of Keltepe (Fig. 3), the amount of subsequent left-lateral offset can be estimated as 1225 ± 25 m.

Although the ages of samples 99YW91 and 01TR51 are not concordant, as for Küreci we form their weighted mean as the ‘best estimate’ of the timing of local volcanism: 427 ± 8 ka ($\pm 2\sigma$). Dividing this by the estimated amount of subsequent left-lateral slip (H–K; Fig. 11) gives a slip rate of 2.87 ± 0.05 mm/a ($\pm 2\sigma$).

Karacağıl

Located *c.* 15 km SSW of Küreci, the Karacağıl area is the only other part of the Karasu Valley where we (and Yurtmen *et al.* 2002) have dated any basalts. In a similar way to farther north, the Amanos Fault runs near the front of the Amanos Mountains. North of Karacağıl and south of Karaelma, at Yassı Tepe, this fault (which locally trends north–south) transects basalt that Yurtmen *et al.* (2002) dated to 428 ± 14 ka ($\pm 2\sigma$) (their sample 99YW84). Yurtmen *et al.* (2002) estimated from the local geomorphology that this Yassı Tepe basalt has become offset left-laterally by 425 ± 25 m, giving a slip rate of 0.99 ± 0.14 mm/a ($\pm 2\sigma$).

Near Karacağıl, just east of the Amanos Fault, Parlak *et al.* (1998) obtained a spiked K–Ar date (their sample 51) of 1050 ± 600 ka ($\pm 2\sigma$). Tatar *et al.* (2004) showed that this basalt is reverse-magnetized, supporting eruption during the Matuyama chron. From the geomorphology, Yurtmen *et al.* (2002) tentatively estimated that this older Karacağıl basalt is offset across the Amanos Fault by 1000 ± 50 m, suggesting a slip rate of 0.95 ± 0.55 mm/a ($\pm 2\sigma$). To improve upon this slip-rate estimate, by redating this basalt, sample 01TR47 was thus collected west of the Amanos Fault in the dry valley of the Çınarlı River, and yielded an Ar/Ar date of 1195 ± 16 ka ($\pm 2\sigma$). The slip rate thus adjusts to 0.84 ± 0.08 mm/a ($\pm 2\sigma$), consistent with the estimate from the Yassı Tepe basalt.

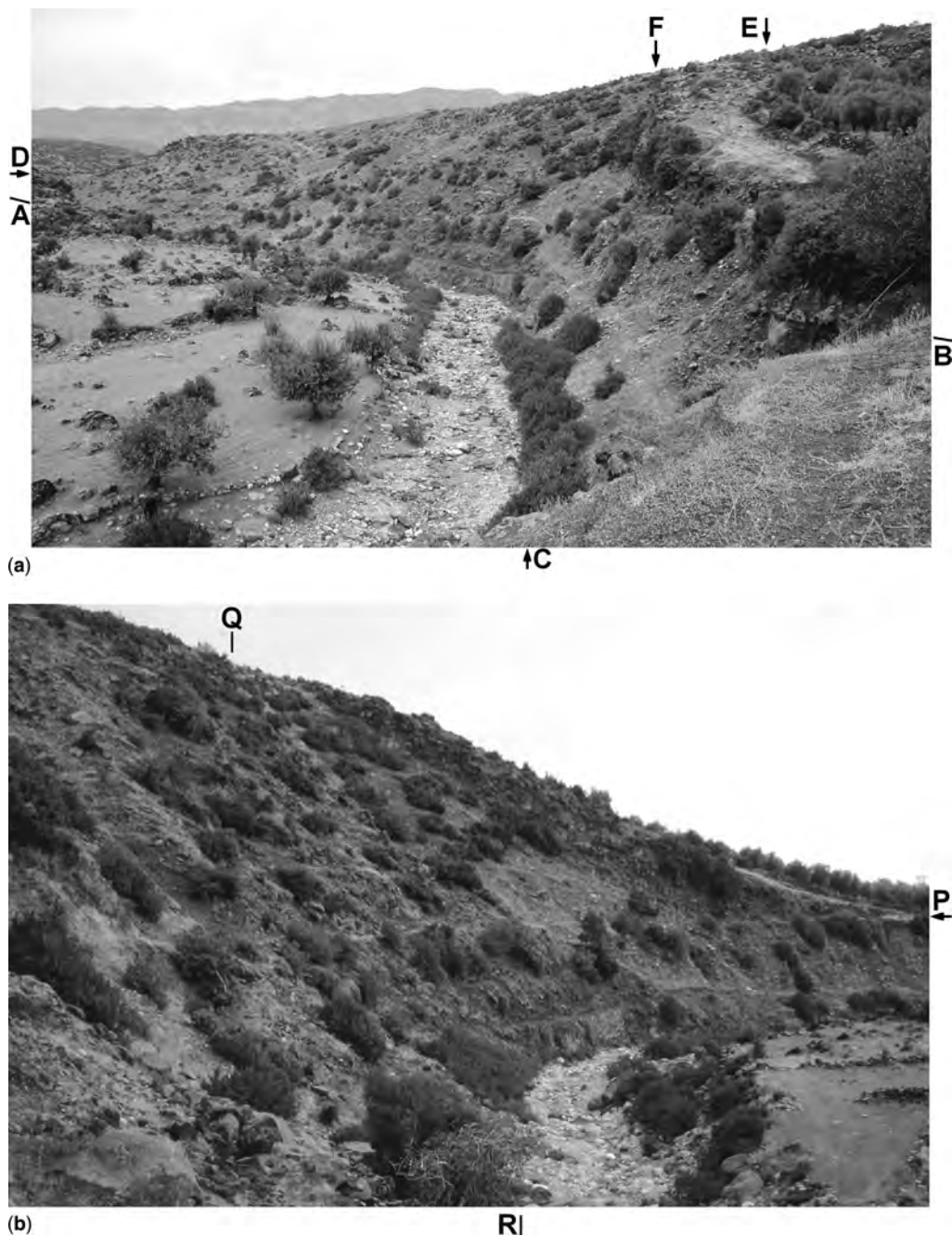


Fig. 10. (a) View SSW looking upstream along the offset reach of the Hacilar River from [BA 75979 69965], where sample 01TR59 was collected. The Amanos Fault (which projects along the line A–B) follows the c. 80 m high bluff in the left bank of this river, leaving the river at the point where this bank is supported by a retaining wall near the centre of the view (in line with C). To the left of this wall, at the base of the bluff, two basalt flows are visible: a lower one, which is vesicular, and an upper massive one with a rubbly base. This upper flow, which yielded samples 01TR56 and 01TR57 (the latter being collected c. 20 m north of the former), dips at c. 40° towards S60°E, indicating that the basalt cascaded over a bluff at the fault line. The point where the Hacilar river joins the Amanos Fault

Yurtmen *et al.* (2002) reported a second significant left-lateral fault, which they named the Guzelce Fault (Fig. 3). This fault, recognizable as a bluff where the land surface is downthrown to the SE by up to several tens of metres, trends NE, obliquely away from the Amanos Mountains; its intersection with the Amanos Fault can be projected roughly halfway between Karacağıl and Ceylanlı (Fig. 3).

South of this fault intersection, basalt is present on both sides of the Amanos Fault. However, as there is no continuity of outcrop, no direct estimate of the local left-lateral slip rate is possible. West of the Amanos Fault at Ceylanlı, Çapan *et al.* (1987) and Rojay *et al.* (2001) obtained spiked K–Ar dates of 1730 ± 200 ka and 1570 ± 160 ka (both $\pm 2\sigma$), respectively (samples 31 and 02). Parlak *et al.* (1998) reported dates from this area of 2200 ± 1400 ka, 790 ± 600 ka, and 400 ± 400 ka (all $\pm 2\sigma$). Tatar *et al.* (2004) established that the local basalt is reverse-magnetized, consistent with eruption early in the Matuyama chron and with the weighted mean of these five dates, which is 1500 ± 15 ka ($\pm 2\sigma$). Rojay *et al.* (2001) pointed out that this Ceylanlı basalt outcrop is truncated by the Amanos Fault, which has evidently taken up significant downthrow to the east as well as left-lateral slip. However, there is no outcrop in the Karasu Valley interior that can be matched to this Ceylanlı basalt; its offset counterpart is presumably beneath the alluvium in the valley interior and displaced northward relative to Ceylanlı by an unknown distance.

The only outcrop of basalt east of the Amanos Fault in this area forms the upper parts of the Büyük Höyük and Küçük Höyük necks, which protrude above the surrounding alluvium. Büyük Höyük was dated to 600 ± 600 ka (spiked K–Ar; $\pm 2\sigma$) by Parlak *et al.* (1998) and to 828 ± 24 ka (unspiked K–Ar; $\pm 2\sigma$) by Yurtmen *et al.* (2002).

The weighted mean of these dates is 828 ± 24 ka ($\pm 2\sigma$), consistent with the reversed geomagnetic polarity measured by Tatar *et al.* (2004). Küçük Höyük was dated to 660 ± 80 ka (spiked K–Ar; $\pm 2\sigma$) by Rojay *et al.* (2001), but has not been investigated magnetostratigraphically. As noted above, the geochemistry of the basalts differs significantly both between these necks and relative to the basalts west of the Amanos Fault in the Ceylanlı area. No basis thus exists for correlating the basalts across the Amanos Fault in this area.

The Guzelce Fault

The Guzelce Fault can be traced northeast then NNE of its intersection with the Amanos Fault for c. 20 km to near Aktepe, as illustrated in Figure 3. For most of its length, basalt crops out on its western side, with alluvium cropping out below the scarp on its eastern side. This basalt appears to have erupted from the Aktepe neck (Fig. 3), and to have flowed SSW down the Karasu Valley. Rojay *et al.* (2001) dated the basalt at Aktepe neck to 260 ± 80 ka ($\pm 2\sigma$) (their sample 06). Water-supply boreholes (DSİ 1975; Rojay *et al.* 2001) indicate that basalt (presumably from the same flow unit) is also present in the subsurface east of the Guzelce Fault. In the past quarter of a million years, since the basalt eruption, this fault has thus evidently taken up tens of metres of downthrow to the east, plus a presumably much greater amount of left-lateral slip.

We tentatively reconcile our estimates of the left-lateral slip rate on the Amanos Fault, of c. 3 mm/a at the sites north of Aktepe (Küreci, Kısık Tepe, Hacılar, Hassa) and c. 1 mm/a in the vicinity of Karacağıl, by presuming that the 'missing' c. 2 mm/a is taken up south of Aktepe on the Guzelce Fault. We thus presume that south of the point where the Guzelce Fault splays from

Fig. 10. (Continued) (Q in Fig. 6b) is obscured behind D, the northern end of the Kocaören basalt (R in Fig. 6b). The basalt visible in the right foreground, between B and C, on the east side of the Amanos Fault, slopes towards the viewpoint and yielded sample 01TR59, indicating that it is part of the Hacılar basalt. The near skyline to the left of F is formed by the main outcrop of the Hacılar basalt, on the west side of the Amanos Fault. (b) View NNE from a point c. 50 m SSW of the retaining wall and adjacent basalt flows mentioned above (which are also shown here and link the two photos), looking downstream along the offset reach of the Hacılar River. The viewpoint in (a) is on the skyline at P. The estimated line of the Amanos Fault is marked both on the skyline (Q, adjacent to F in (a)) and in the dry river bed (R); as noted above, the fault leaves the river where its bank is supported by the retaining wall, near the middle of this field of view. Between the skyline and the point where it intersects the river the Amanos Fault can be seen to dip eastward at c. 50°. On its eastern (Arabian) side, a flat-lying flow unit of ropy basalt with a vesicular top is just visible in the left bank of the river after it has left the fault line, behind the low river terrace in the foreground. This yielded sample 01TR58. This basalt flow is overlain by very coarse fluvial gravel, mostly comprising limestone clasts, forming a higher river terrace reaching c. 15 m above river level. This fluvial gravel is overlain by a wedge of colluvium, then by thin basalt that plunges down the bluff at c. 25°, from which sample 01TR59 was collected. The upper end of this *in situ* dipping basalt is separated from its counterpart west of the fault by a c. 20-m-wide gap (to the left of Q), where only basalt rubble is present, providing an indication of the heave on the Amanos Fault since the eruption of this basalt occurred.

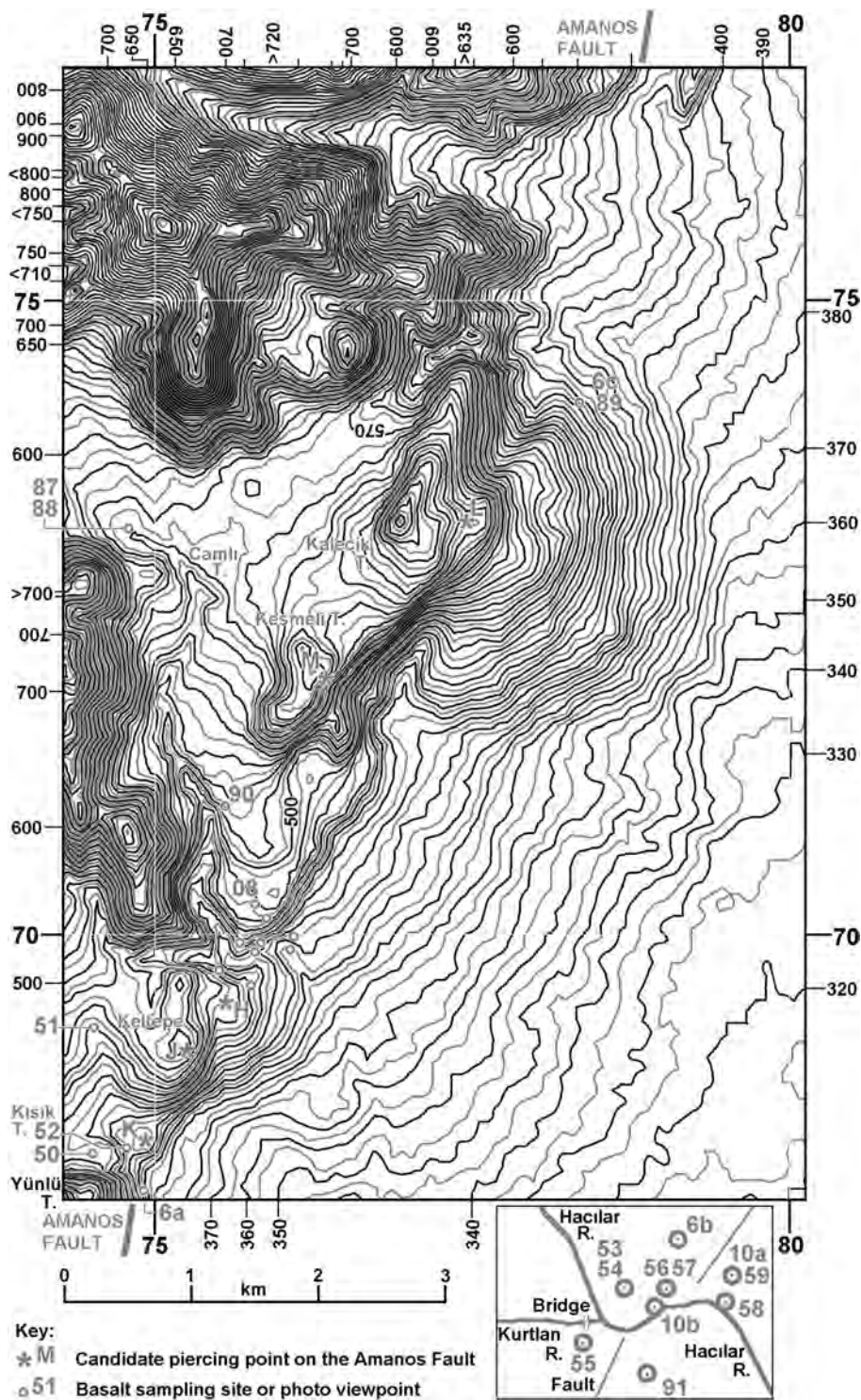


Fig. 11. More detailed SRTM topographic map of the Amanos Fault and its surroundings in the Hassa–Kesmeli Tepe–Hacilar–Keltepe–Kısık Tepe area, with 5-m contours, showing sample locations, photo viewpoints, and candidate piercing points used for slip and slip rate calculation. See Westaway *et al.* (2006a) for details of data sources and preparation technique. The inset shows the area of the offset Hacilar river gorge (c. [BA 76 70]) at double the scale.

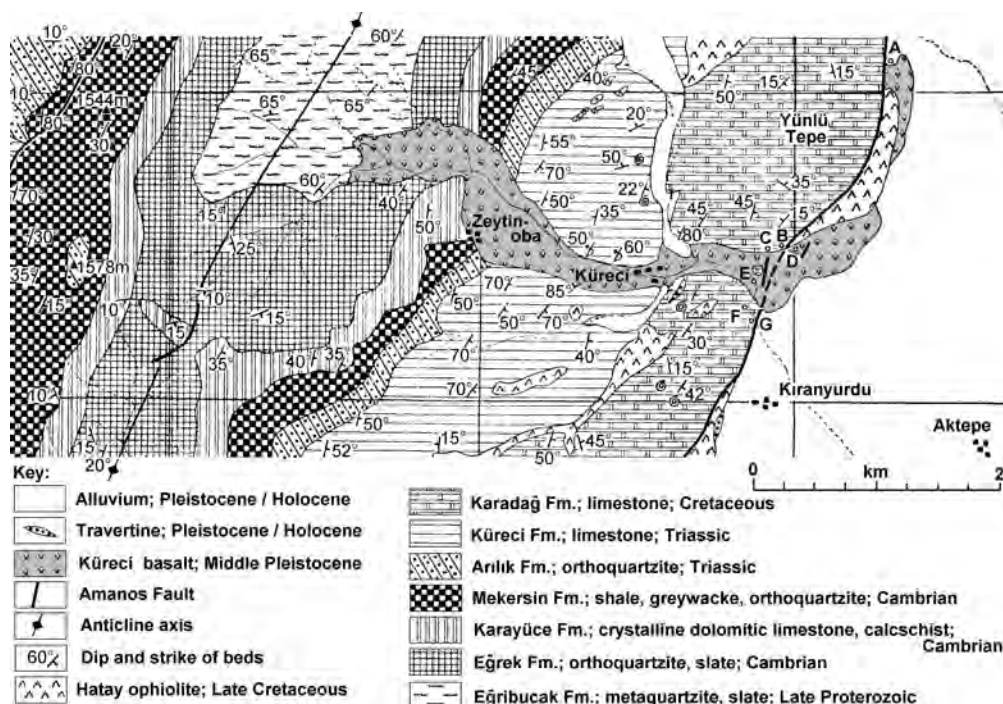


Fig. 12. Geological map of the Küreci basalt flow unit, adapted from part of plate II of Atan (1969), showing his stratigraphic terminology for the area. The meaning of points A, B, C, D and G, illustrated, is discussed in the text. The apparent absence of basalt directly south of point D, not recognized by Atan (1969) but evident from our own fieldwork, is illustrated schematically. Line segments linking C–G and east of D indicate positions of strands of the Amanos Fault deduced from our own fieldwork. Dashed line B–G illustrates Atan's (1969) tentative alternative interpretation. Point E is the location where basalt sample 99YW95 was collected, determined by GPS as [BA 73857 65985]. This site can be reached along a track that leaves the Kiranyurdu–Küreci road at [BA 73703 66045]. Point F is the viewpoint where the photographs in Figure 5b & c were taken. Note the anticline axis suparallel to the Amanos Fault, c. 4 km farther west. The Eğribucak inlier of Late Proterozoic basement and the overlying Palaeozoic succession crop out locally, forming the core of the Amanos mountain range. Note that Atan's (1969) formation names are unofficial and that other terminologies also exist. Later studies such as Schwan (1971) and Dean & Monod (1985) discuss how these different nomenclatures correlate.

the Amanos Fault, the slip rate on the latter is once again c. 3 mm/a. However, we do not understand how displacement 'transfers' between the northern end of the Guzelce Fault and the Amanos Fault (whether by localized slip on an unidentified fault linking the two, in the area west of Aktepe, or by local distributed deformation); further investigation of this point is beyond the scope of this study.

Basalt east of the Guzelce Fault

Çapan *et al.* (1987) first proposed that much of the basalt in the central-southern Karasu Valley originated from the large Büyük Aktepe neck, located c. 10 km east of Hassa, just west of the Syrian border. In their interpretation, basalt flowed SSW from this source for >20 km, passing east of Aktepe town and the adjoining neck. The southern

limit of exposure of this flow unit is c. [BA 750 570], c. 8 km south of Aktepe, although it may continue in the subsurface beneath the young alluvium of the Karasu Valley interior.

At a locality that they called Korogha Geri, c. 4 km ESE of Aktepe town, Çapan *et al.* (1987) obtained a spiked K–Ar date of 2100 ± 400 ka ($\pm 2\sigma$). Basalts from the Karasu Valley were subsequently analyzed geochemically by Polat *et al.* (1997). Although they published no location information, and their full data-set cannot now be traced, enquiries during the preparation of the Yurtmen *et al.* (2002) paper indicated that their 'low SiO₂' set of basalts (the majority that they studied) came from this flow unit. Tatar *et al.* (2004) showed that this flow unit, both in the area near Aktepe (they sampled it at their sites 8 and 9, c. 1.5 and c. 3 km SSE of Aktepe at Karapınar and Hanobası Fig. 3) and in the area around Büyük

Aktepe neck, is normally magnetized. They also obtained two spiked K–Ar dates for this flow unit, 1420 ± 240 ka ($\pm 2\sigma$) at site 43 c. 4 km west of the Büyük Aktepe neck and 490 ± 600 ka ($\pm 2\sigma$) at site 42 adjacent to this neck, from what may be a late eruption unrelated to the wider flow unit. The weighted mean of the dates from Çapan *et al.* (1987) and site 43 of Tatar *et al.* (2004) is 1600 ± 206 ka ($\pm 2\sigma$). This error margin overlaps with the age span of the normal-polarity Olduvai subchron. This flow unit would thus seem to be the oldest known in the study region, predating the reverse-magnetized basalts in the Ceylanlı area (see above). We have not found any sites where this basalt is suitable for dating, but geochemical analysis of sample 01TR48 of it, collected just east of the bridge over the Kargılı River, c. 2 km SSE of Aktepe, confirms the mafic and primitive magmatic characteristics observed by Polat *et al.* (1997) (Fig. 8).

We inferred earlier that, south of Aktepe, the left-lateral slip seems to be partitioned with c. 0.8 mm/a on the Amanos Fault and c. 2.2 mm/a on the Güzelce Fault. Taking a nominal age for the Büyük Aktepe neck flow unit of 1.8 Ma, restoring SW–NE-oriented left-lateral slip on the Güzelce Fault would juxtapose localities east of this fault southwestward by c. 4 km, placing the southern end of this basalt outcrop c. [BA 722 542]. Such a restoration would roughly align the southern end of this basalt with the northern end of other basalt present east of the Amanos Fault around Karaelma (Fig. 3).

According to Atan (1969), the northern limit of this Karaelma basalt east of the Amanos Fault is c. [BA 665 535]. This basalt covers a substantial area east and south of this locality, and may also be present farther north and east in the subsurface, beneath young alluvial-fan deposits shed from the Amanos Mountains. Tatar *et al.* (2004) determined normal geomagnetic polarity in the basalt at Karaelma. Notwithstanding the vague location information that they provided, it appears that they studied the Karaelma basalt, rather than the stratigraphically younger Yassı Tepe basalt that crops out south of Karaelma (see above). Their magnetostratigraphic results from this area may thus again indicate the Olduvai subchron, raising the possibility that the Karaelma basalt forms the offset SW part of the larger Büyük Aktepe flow unit, as is tentatively implied by our above slip restoration. These basalts thus offer scope for future refinement of the Amanos Fault kinematics, but further discussion is beyond the scope of this study.

Vertical crustal motions

As already noted, the principal active faults within the Karasu Valley are not pure left-lateral faults;

there are also minor components of normal slip, with downthrow to the east on the Amanos and Güzelce faults and to the west on the East Hatay Fault. Overall, the Karasu Valley therefore resembles a graben, it being thus sometimes (rather loosely) called the ‘Karasu Rift’ (e.g. by Tatar *et al.* 2004; cf. Mart *et al.* 2005). However, recent kinematic models (e.g. Westaway 2004) indicate that any extension on these faults is more than compensated by the distributed shortening within the surrounding mountain ranges (the Amanos Mountains to the west, and the Kurd Dag or Jabal al-Akrad to the east) that is occurring to accommodate the transpression along the northern DSFZ.

At several localities where basalt is offset (e.g. Küreci, Fig. 5b; Kısık Tepe, Fig. 6a, and Hacılar, Fig. 10), scarps tens of metres high have developed along the Amanos Fault over hundreds of thousands of years, implying vertical slip rates of c. 0.1 mm/a, only a few percent of the c. 3 mm/a left-lateral slip rate. As indicated in Figure 10b and its caption, at Hacılar an estimated c. 20 m of heave has developed on this c. 50°-dipping fault since the basalt eruption at c. 160 ka. The horizontal slip rate can thus be estimated as c. 0.13 mm/a, and the vertical slip rate as c. $0.13 \text{ mm/a} \times \tan(50^\circ)$ or c. 0.15 mm/a. This is less than the vertical slip rate estimate by Yurtmen *et al.* (2002), who inferred that the bluff along the Amanos Fault at Hacilar developed entirely since the youngest basalt eruption. However, our subsequent fieldwork (summarized above and in the caption to Fig. 10) establishes that part of this bluff already existed at c. 160 ka, since the basalt cascaded down it. Moreover, a bluff already existed along the line of this fault in this locality when the older basalt beneath the Hacilar basalt (which we correlated above with the Kesmeli Tepe basalt, with an inferred age of c. 400–500 ka), since this basalt likewise has a steep local dip to the ESE. It is thus evident that here, and by analogy elsewhere on the Amanos Fault, the vertical component of slip is constantly ‘trying’ to add to the local relief, but other local processes, such as erosion, deposition by alluvial fans, and the intermittent eruption of basalts, act to progressively remove this relief.

Geological and topographic maps of the Amanos Mountains have been published by Westaway *et al.* (2006a); for cross-sections illustrating their structure see Figure 4. These mountains can be regarded as an asymmetrical anticline oriented SSW–NNE (parallel to the Amanos Fault), with a gentle western limb and a steeper eastern limb. This asymmetry means that, typically, the oldest rocks present (of Late Proterozoic and Palaeozoic ages) crop out near the eastern margin of the range. For instance, inliers of the Late Proterozoic Eğribucak Formation occur near Zeytinoba in the north of the present study region (Fig. 12) and

near Ceylanlı in the south (Fig. 4b). However, in detail, this mountain range is more complex, due to the combined effects of folding into many smaller-scale anticlines and localized reverse faulting (Figs 4b & 12).

As already noted, the folded structure of the Amanos Mountains was intensively researched in the 1960s (e.g. Atan 1969; Schwan 1971, 1972); this work led to the conclusion that this folding is ancient, probably related to the Late Cretaceous ophiolite obduction, and thus unrelated to the active crustal deformation. Moreover, at the time when this fieldwork was done, mainly in 1966–1968, the Karasu Valley was thought to be a graben; the component of left-lateral slip across it had not yet been recognized. This component was first recognized across the southern DSFZ by Freund *et al.* (1970); the subsequent history of development of ideas regarding the northern DSFZ (which has included setbacks due to the publication of mistaken hypotheses) has been discussed at length by Yurtmen *et al.* (2002) and Westaway (2003, 2004), and so is not repeated here. Furthermore, when the work on the Amanos Mountains was carried out, the EAFZ had not yet been recognized; its ‘discovery’ is generally attributed to Arpat & Şaroğlu (1972) and the widespread acceptance of its role as the TR–AR plate boundary to McKenzie (1976). Thus, at the time when the structure of the Amanos Mountains was intensively researched, there was not yet even any suggestion that these mountains are located within the linkage of left-lateral faults forming the AF–AR and TR–AR plate boundaries. Indeed, prior to the Westaway (2003, 2004) studies, the possibility of active transpression associated with this left-lateral faulting had not occurred to anyone. The structure of these mountains was instead considered to be a problem that had already been solved, the solution being of no interest to any consideration of the active tectonics.

Given that the Amanos Mountains can now be seen to form part of a zone of active transpression along the northern DSFZ (cf. Westaway 2003, 2004), it is important to consider the possibility that part (as opposed to none) of their structural development relates to this active deformation. This situation contrasts with that farther south along the Lebanon stepover (Fig. 1). As discussed by Walley (1998), it was formerly thought that the structural development of that area relates entirely to transpression along the DSFZ (cf. Westaway 1995), but more recent analysis has concluded that part (as opposed to all) of the structural development relates to this active deformation, and part of it relates to processes in the Late Cretaceous or at other ancient times (cf. Gomez *et al.* 2006). Thus, the Lebanon and Amanos mountain ranges may have experienced similar histories of

deformation, with the most recent phase of both relating to the active transpression. As Westaway (2003, 2004) also noted, the Jebel Nusayriyah or Syrian Coastal Range in between (Fig. 1) can now be seen to be another transpressive stepover along the northern DSFZ, but with less-dramatic deformation than the others because it is less-strongly misaligned to the tangential direction to the DSFZ Euler pole (cf. Gomez *et al.* 2006).

The geomorphology of the Amanos Mountains closely reflects their structure. Along most of their length the drainage divide is near the eastern margin of the range, often only *c.* 2–4 km west of the Karasu Valley. For instance, the few roads that cross these mountains tend to cross cols no more than *c.* 2–4 km from the Karasu Valley. A notable example is the motorway linking Adana to Gaziantep, which crosses this mountain range near Bahçe (Fig. 2). Its summit is in the *c.* 1.2 km long Aslanlı tunnel, from which this motorway emerges overlooking the Karasu Valley. This typically asymmetrical drainage divide suggests that local rates of Late Cenozoic surface uplift have been greatest close to the eastern margin of the mountain range, mimicking their overall structure.

The overall topography of the Amanos Mountains is more symmetrical, with the highest summits typically midway between the western and eastern range fronts. In the south, the mountain range is barely 15 km wide and rises no higher than *c.* 1800–1900 m a.s.l. between İskenderun and Kırıkhan; the highest summit altitude decreases even lower, to 1427 m a.s.l., WSW of Kırıkhan beside the southern end of the Amanos Fault. North of Hassa, the mountain range widens to *c.* 30–40 km, with the highest summits >2200 m a.s.l. The highest topography of all is found in the widest part of the Amanos Mountains in the north, NE of Düziçi and SW of Kahramanmaraş, for instance, the mountains Uluziyaret (2268 m a.s.l.) and Beşikdülü (2246 m a.s.l.) (Fig. 2). It can thus be presumed that erosion along the eastern margin of the mountain range has been rapid enough to limit the local growth of topography; hence, the exhumation of the oldest rocks in the region. For comparison, the Kurd Dag range, east of the East Hatay Fault in NW Syria (Fig. 2), typically rises no higher than *c.* 1100 m a.s.l. (Fig. 5b), although in the north, on the Turkish border, it reaches 1275 m a.s.l.

The only river to transect the Amanos Mountains is the Ceyhan, which enters this range west of Kahramanmaraş and leaves it at Cevdetiye (Fig. 2). Its Berke Gorge passes through the uplands north of Beşikdülü mountain and is *c.* 1200-m-deep (river *c.* 300 m a.s.l.; land surface adjoining the gorge *c.* 1500 m a.s.l.), one of the most dramatic river gorges in Turkey. It can thus be inferred that much of the surface uplift that

formed the Amanos Mountains occurred at a late stage, after this river had become established (following the Middle Miocene emergence of the region above sea-level); presumably, the Ceyhan had sufficient erosional power to incise in pace with the subsequent surface uplift.

Estimates of the uplift rate of the part of the Amanos Mountains that is transected by the River Ceyhan have recently been obtained by us (and will be published in full elsewhere; see Seyrek *et al.* in press) from Ar/Ar dating of basalt flows that cap fluvial terraces. The basalts in this area form the eastern part of the Ceyhan–Osmaniye volcanic field (cf. Bilgin & Ercan 1981; Arger *et al.* 2000; Yurtmen *et al.* 2000; Gürsoy *et al.* 2003). We have obtained six concordant Ar–Ar dates for basalt abutting the Ceyhan Gorge in the western part of the Amanos Mountains, which yield a weighted mean age of 278 ± 7 ka ($\pm 2\sigma$), suggesting that a single, brief, episode of volcanism occurred in this area. At Pınarözü, beside the Aslan-ış Dam (c. 25 km north of Osmaniye; Fig. 2), the basalt reaches no lower than c. 195 m a.s.l.; the river was locally c. 85 m a.s.l. before the dam flooded it, indicating c. 110 m of incision, at a time-averaged rate of 0.39 mm/a. About 10 km farther downstream, at Karagedik, the basalt reaches no lower than c. 145 m a.s.l.; the river is locally c. 70 m a.s.l., indicating c. 80 m of incision, at a time-averaged rate of 0.27 mm/a. Roughly 5–10 km farther downstream, at the western margin of the Amanos Mountains around Sarpınagzı and Cevdetiye (Fig. 2), a well-developed Ceyhan terrace is observed c. 35 m above river level. Assuming that these terrace deposits accumulated in MIS 6 (c. 140 ka), an uplift rate of 0.25 mm/a is indicated. Upstream extrapolation of this variation in uplift rates would imply a rate of at least c. 0.4 mm/a along the Berke Gorge in the core of the Amanos Mountains. If extrapolated back in time, this would predict that the incision of this gorge began around c. 3 Ma, roughly when the modern geometry of strike-slip faulting in this region is thought to have developed (cf. Westaway *et al.* 2006a).

Quaternary regional surface uplift is widely observed elsewhere in Turkey, away from active fault zones (e.g. Demir *et al.* 2004). However, in the Arabian Platform and along the Mediterranean coastline, its rate seldom exceeds c. 0.1 mm/a; the much faster uplift now evident in the Amanos Mountains requires a different explanation. We suggest that this vertical crustal motion, as evidenced by the vertical slip on the Amanos Fault and the fluvial geomorphology (including the dating of Ceyhan terraces, above), is being caused primarily by the component of crustal thickening required to balance the crustal shortening that is required by the transpressive geometry of this

fault. Supporting evidence for this view is provided by the orientation of fold axes within the Amanos Mountains, which (as in Fig. 12) are typically subparallel to the Amanos Fault, making them optimally oriented to accommodate fault-normal crustal shortening (cf. Westaway 1995). The asymmetry of the Amanos Mountains, revealed by the asymmetrical drainage and folding (see above) thus implies that the strain rates for distributed crustal shortening and thickening typically increase eastward to maxima close to the Amanos Fault. However, the characteristic eastward downwarping of the structure in close proximity to this fault, evident in Figure 4, means that vertical slip rates on the Amanos Fault will underestimate the peak uplift rates in the interior of the Amanos Mountains. Thus (for instance) the peak uplift rate in the Amanos Mountains west of Hacılar may be several times greater than the local c. 0.15 mm/a vertical slip rate on this fault.

Estimation of the total Late Cenozoic surface uplift in the Amanos Mountains is difficult, due to the typical lack of suitable outcrop evidence; as Figs 4 & 12 show, most outcrop is far too old to be used in such a calculation. Probably the best evidence is provided by outcrop, flanking the eastern margin of the southern Amanos Mountains (c. 7 km WNW of Kırıkhan; c. [BA 565 448]) of Middle Miocene (c. 15 Ma) reefal limestone, of the Kepez Formation, which is now c. 950 m a.s.l. (cf. Yurtmen *et al.* 2002). This requires a time-averaged uplift rate of c. 0.06 mm/a; but, if the local uplift is assumed to have been concentrated after the initiation of the modern geometry of strike-slip faulting (c. 3.7 Ma; Westaway *et al.* 2006a), then the subsequent local rate of surface uplift may have reached 0.25 mm/a, comparable with the evidence from the Ceyhan terraces.

As Westaway (1995) noted, the spatial average strain rate E_s for the distributed shortening along a transpressive stepover (with deformation partitioned symmetrically on both sides) is:

$$E_s = \frac{V \sin(\theta)}{2H} \quad (1)$$

where V is the local relative plate velocity and H is the width of the mountain range forming the stepover. Using equation (1), with $V = 8$ mm/a (Fig. 13), $H = 30$ km and $\theta = 52^\circ$, we estimate the spatial average strain rate for crustal shortening in the Amanos Mountains as c. 0.1 Ma^{-1} . The shortening factor, if deformation at this rate persists for time t , is $\exp(-ht)$ and is thus c. 1.5 if this deformation has persisted since c. 3.7 Ma. A predicted shortening strain of this magnitude seems roughly consistent with the evidence (Figs 4 & 12).

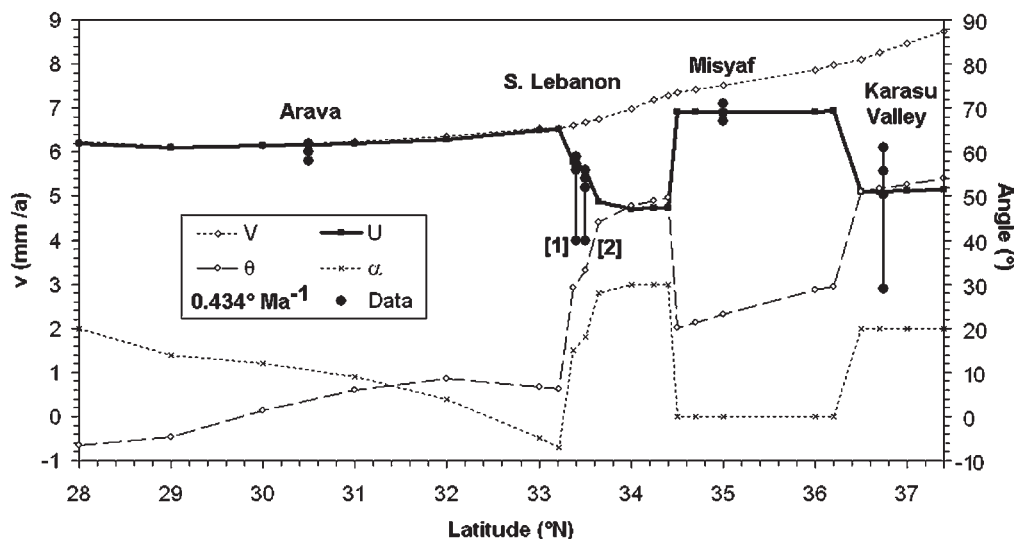


Fig. 13. Summary of the geometry of the Dead Sea fault zone, between the Tiran Strait and Türkoğlu at the northern end of the Amanos Fault. At each point, V is the rate of relative motion between the African and Arabian plates, calculated using spherical trigonometry (including the effect of Earth ellipticity) from the position of the point relative to the assumed Euler pole at 31.1°N , 26.7°E , given the preferred $0.434^{\circ}/\text{Ma}$ rate of relative rotation. α is the observed strike of the principal DSFZ strand, measured clockwise from north. θ is the difference in angle between α and the tangential direction to the Euler pole. U is the maximum possible rate of left-lateral slip on this fault strand, calculated as $V \times \cos(\theta)$, where V is the local rate of relative plate motion. The data used for comparison with predictions of this kinematic model are discussed in the text. For southern Lebanon, analysis 1 uses the overall time-averaged slip rate for the Serghaya Fault, estimated in the text, whereas analysis 2 uses the Holocene slip rate from Gomez *et al.* (2003). These two estimates have been shown at separate localities, to avoid clutter, although in reality they apply to the same part of the Serghaya Fault.

Neglecting any effects of advection of the thermal boundary at the base of the brittle layer relative to the rock column, or of other thermal effects such as those resulting from erosion, the rate of thickening of the brittle upper crust can be crudely estimated by multiplying the strain rate by the thickness of this brittle layer, which can be estimated as *c.* 15 km. One thus obtains *c.* 1.5 mm/a.

The Amanos mountain range is so narrow that (like the Lebanon Mountains; cf. Khair *et al.* 1993) it is unlikely to be isostatically supported by downward deflection of the underlying mantle lithosphere. The local isostatic balance is probably determined instead by the extent to which local processes (erosion and crustal thickening) can maintain the base of the brittle layer at a different depth beneath this mountain range than beneath its surroundings, given the tendency for such variations to be dynamically removed by diffusion of heat within the crust. Experience of numerical modelling of such effects elsewhere (e.g. Westaway 2002; Westaway *et al.* 2004, 2006b) indicates that the surface uplift rate will be only a small proportion of the associated thickening rate of the brittle layer; thus, the observed uplift rates in the Amanos Mountains of a few tenths of 1 mm/a

can be tentatively explained. However, formal numerical modelling of this effect is beyond the scope of this study, but will be discussed elsewhere; see Seyrek *et al.*, in press. In the meantime, our preliminary analysis suggests the strong possibility that the observed surface uplift, topography and much of the shortening strain expressed in the structure of the Amanos Mountains and surroundings, have developed during the present phase of strike-slip faulting on the DSFZ, as consequences of the local transpression. We thus see no basis to infer that any left-lateral faulting occurred in the vicinity of the Karasu Valley before *c.* 3.7 Ma; before this time, the northern DSFZ was presumably located farther east (e.g. on the Afrin Fault; Fig. 1), and led into the associated array of faults depicted in the Gaziantep region of SE Turkey in Figure 1.

Discussion

Recurrence of large earthquakes in the Karasu Valley

The magnitude estimate of 7.5 for the 1822 earthquake, from Ambraseys & Jackson (1998),

translates to a seismic moment of $c. 2 \times 10^{20}$ N m (after Kanamori & Anderson 1975). Taking the length of faulting as 200 km (after Ambraseys & Jackson 1998), the thickness of the brittle upper crust as 15 km, and its shear modulus as 30 GPa, $c. 2.2$ m of coseismic slip can be estimated. This earthquake can be presumed to have de-stressed the whole region, so will not recur until sufficient strain has accumulated between the adjoining plates to match this coseismic slip. Dividing the 2.2 m of coseismic slip by our 5.57 ± 0.54 mm/a ($\pm 2\sigma$) overall slip rate gives an estimated recurrence interval of 395 ± 38 ($\pm 2\sigma$) years. Thus, even though we are now predicting significantly faster overall left-lateral slip across the Karasu Valley than before, no such large earthquake is expected to recur on this part of the DSFZ until the late 22nd century, at the earliest, although (of course) a smaller event may occur sooner.

Comparison with other transpressive DSFZ stepovers

Our analyses of vertical and horizontal crustal motions in and around the Karasu Valley support the suggestion (by Westaway 2003, 2004) that this DSFZ segment is a transpressive stepover. The region thus warrants comparison with the two other significant transpressive stepovers on the DSFZ, formed by the Lebanon/Anti-Lebanon Mountains and the Jabal Nusayriyah (Fig. 1).

Westaway (2003, 2004) determined the DSFZ Euler pole at 31.1° N, 26.7° E. The tangential direction to it, measured at Hacilar (36.75° N, 36.45° E), is $N32^\circ$ W. The Amanos Fault strikes $N20^\circ$ E; thus, it is misaligned by an angle, θ , of 52° in the transpressive sense. Westaway (2004) likewise determined θ for the Lebanon stepover as 48 – 50° , given the local $N30^\circ$ E fault strike and the $N18^\circ$ W– $N20^\circ$ W tangential direction to the pole. For the Jabal Nusayriyah stepover (between Tell Kalakh and Jisr esh-Shugur; Fig. 1), θ increases northward from 20 to 30° , given the north–south strike of the faulting and the $N20^\circ$ W– $N30^\circ$ W tangential direction to the pole.

The Lebanon and Amanos stepovers thus have similar geometries, with similar values of θ . Both also consist of a central valley flanked by mountain ranges, with the principal active left-lateral faults running along the valley margins. Thus, the Bekaa Valley is analogous to the Karasu Valley, the Lebanon Mountains to the Amanos Mountains, and the Anti-Lebanon Mountains to the Kurd Dag. The Yammounneh Fault, which typically follows the west side of the Bekaa Valley, is thus analogous to the Amanos Fault; and the Serghaya Fault, which for much of its length is similarly

located relative to the eastern margin of this valley, is thus analogous to the East Hatay Fault.

Structural cross-sections through both the Amanos and Lebanon mountain ranges look similar, as can be seen by comparing our Figure 4 with, for instance, Figure 4 of Westaway (2004) or Figure 4 of Gomez *et al.* (2006). In terms of equation (1), H is roughly the same for the Amanos and Lebanon stepovers, and V is less for the latter, because it is nearer the DSFZ Euler pole. One thus expects somewhat higher E_s in the Amanos Mountains than in the Lebanon Mountains. However, the Lebanon Mountains rise much higher than the Amanos Mountains: their highest summit (Qurnat as-Sawda) reaches 3087 m a.s.l.; the highest point in the Anti-Lebanon Mountains (Jabal ash-Shaykh or Mount Hermon) reaches 2814 m a.s.l. These high summits are near the range fronts bounding the major strike-slip faults. They are thus not analogous to the highest summits in the Amanos Mountains; as noted above, erosion has prevented any really high topography from developing along the eastern margin of the Amanos Mountains adjacent to the Amanos Fault. The lesser degree of erosion in the Lebanon Mountains is indicated by the absence of outcrops older than Mesozoic, there being no Palaeozoic or Precambrian inliers (the limited Late Cenozoic erosion of these mountains was also noted by Walley 1998).

In comparison, the Jabal Nusayriyah stepover is a more subdued edifice, reflecting the much lower range of θ . The topography increases northward along it, reflecting the northward increase in θ and E_s . The highest topography is indeed at its northern end, rising to 939 m a.s.l. east of the DSFZ in the Jabal Jubb Sulayman, and to 1562 m west of the DSFZ in the Jabal Nusayriyah (Fig. 1). Like the Amanos stepover, the Jabal Nusayriyah stepover is thus asymmetrical, with higher mountains on its western side. However, only its northern part contains a broad linear valley bounded by left-lateral faults on both sides, the Ghab Basin; in the south, through Misyaf (Fig. 1) the DSFZ consists of only a single active fault, delineating a narrow linear valley. This difference in morphology may relate to the northward increase in E_s that is required by its geometry.

Consistency between local slip rates and regional kinematics

Setting aside localities with multiple en échelon fault strands (e.g. where the Guzelce Fault is present), we have obtained four estimates (all $\pm 2\sigma$) of the slip rate on the Amanos Fault from offset Pleistocene basalt flows: 2.89 ± 0.14 mm/a

at Kesmeli Tepe/Hassa; 2.84 ± 0.41 mm/a at Hacilar; 2.87 ± 0.05 mm/a at Kısık Tepe/Kocaören; and 3.25 ± 0.22 mm/a at Küreci. The weighted mean is 2.87 ± 0.05 mm/a for the first three, which are tightly grouped, and 2.89 ± 0.05 mm/a ($\pm 2\sigma$) for all four, indicating total slip since 3.73 ± 0.05 Ma of 10.8 ± 0.2 km ($\pm 2\sigma$). Westaway (2004) estimated the total slip on the East Hatay Fault as c. 10 km from piercing points derived from offset outcrops of the Hatay ophiolite ('K' in Fig. 2). These offsets were measured from maps, since it has proved impossible to carry out fieldwork on the Turkey–Syria border where this fault and its piercing points are located, and so cannot be checked directly. We thus assign a nominal margin of uncertainty of 1 km, making this slip estimate 10 ± 1 km, giving a time-averaged slip rate since 3.73 ± 0.05 Ma of 2.68 ± 0.54 mm/a ($\pm 2\sigma$). Summing these slip rate estimates for the Amanos and East Hatay faults gives 5.57 ± 0.54 mm/a ($\pm 2\sigma$).

We now combine this data-set with three others from farther south to assess the overall kinematics of the DSFZ (cf. Westaway 2004). First, Ginat *et al.* (1998) estimated c. 15 km of left-lateral slip (which we take as 15 ± 1 km) since c. 2.5 Ma in the Arava Valley in southern Israel (c. 30.5° N), indicating a time-averaged slip rate of 6.0 ± 0.2 mm/a. Second, we combine Westaway's (2004) analysis of the offset of the Litani River in southern Lebanon (c. 33.4° N) by the Yammouneh Fault, which yielded a c. 4.0 mm/a slip rate estimate for the Middle–Late Pleistocene, with the analysis by Gomez *et al.* (2003) of the Serghaya Fault at c. 33.6° N that indicated a Holocene slip rate of 1.4 ± 0.2 mm/a. We thus estimate a 5.4 ± 0.2 mm/a overall slip rate for both faults at c. 33.4° N. Gomez *et al.* (2006) have subsequently reported the total slip on the part of the Serghaya Fault between c. 33.5 and c. 33.9° N as 6–7 km from drainage offsets in the Anti-Lebanon Mountains. They estimated that this slip post-dates the development of an erosion surface of estimated Messinian age (c. 7–5.5 Ma), indicating a time-averaged slip rate of c. 1 mm/a. However, if the present geometry of faulting in this area developed at c. 3.7 Ma, as is now suggested for the northern DSFZ, this estimate of 6.5 ± 0.5 km of slip indicates a time-averaged rate of 1.74 ± 0.14 mm/a. Finally, Meghraoui *et al.* (2003) estimated the slip rate of the DSFZ strand at Misyaf in western Syria (c. 35.0° N), from trenching and radiocarbon dating of the slip in great earthquakes over the past c. 2 ka, as 6.9 ± 0.2 mm/a ($\pm 2\sigma$).

Theory by Westaway (1995) shows that the slip rate U for a strike-slip fault within a stepover can be no greater than $V \times \cos(\theta)$. We assume in the following analysis that all faults investigated are

slipping at this maximum rate permitted by their geometry, given their misalignment relative to the tangential direction to the AF–AR Euler pole.

As Westaway (2004) noted, the Meghraoui *et al.* (2003) data-set is consistent with a rate of relative rotation of $0.434 \pm 0.012^\circ$ /Ma about the preferred DSFZ Euler pole. As Figure 13 shows, this Euler vector is also consistent with the data from Israel and Lebanon, and with the lower bound to our estimate for the overall left-lateral slip rate across the Karasu Valley. We thus conclude that there is now no basis to suppose that the northern DSFZ, at the latitude of the Karasu Valley, consists of any other faults with significant slip rates (cf. Westaway 2004); we thus infer (for instance) that the Afrin Fault, farther east (Fig. 1), represents an earlier phase of DSFZ activity, preceding c. 3.7 Ma.

However, in detail, our Karasu Valley data-set predicts faster slip than is expected from the regional kinematic model (i.e. the opposite problem to the Westaway 2004 analysis). Greater consistency would result if the total slip on the East Hatay Fault were adjusted to 9 ± 1 km, making its slip rate 2.41 ± 0.54 mm/a ($\pm 2\sigma$) and the overall slip rate 5.30 ± 0.54 mm/a ($\pm 2\sigma$) (Fig. 13). Although further work is thus clearly needed to tie down the precise slip rates, the basis for overall kinematic consistency is now established.

Evidence pertaining to the history and rate of slip on the northern DSFZ is also provided by the left-lateral offset of the Homs basalt in western Syria (cf. Westaway 2003, 2004). Eruption of this basalt began at c. 8 Ma but was concentrated during c. 6.5–5 Ma (see Westaway 2003, 2004, for detailed maps and syntheses of the dating evidence). Estimates of the left-lateral slip since this basalt eruption have ranged from <10 km to c. 20 km (see compilation by Westaway 2003), depending on choices of piercing points. The situation is complicated because, as Westaway (2004) noted, the linear valley along the DSFZ already existed at the time of this volcanism, with basalts flowing into it from both sides, making it impossible to match individual flows across the fault zone. Chorowicz *et al.* (2005) and Gomez *et al.* (2006) have argued that the western end of the main volcanic edifice, east of the DSFZ north of Shin (Fig. 1), can be uniquely matched against the basalt west of the DSFZ in the vicinity of the crusader castle of Crac des Chevaliers (Qalat al-Hisn), giving a left-lateral offset of 20 km. Dividing this by a nominal 6 Ma age for the basalt, Chorowicz *et al.* (2005) estimated a subsequent slip rate of 3.3 mm/a, much less than the 6.9 mm/a prediction for this locality (c. 34.7° N) from the present kinematic model (Fig. 13).

Chorowicz *et al.* (2005) did not support their interpretation with proper evidence, such as matching the dates or geochemistry of the basalts at their candidate piercing points; their interpretation instead matched the topography. However, since the transpression in this area is asymmetrical (see above, also Westaway 2003, 2004), being stronger west of the DSFZ, much of the present topography has developed since the basalt eruption, at different rates on opposite sides of the DSFZ; this present topography should thus not be assumed to match that which existed during the basalt eruption. This proposal for piercing points seems unlikely, as available maps (e.g. those by Westaway 2003, 2004) and simple field inspection indicate that the basalt around Crac des Chevaliers is much thinner than that in its supposed counterpart east of the DSFZ.

Westaway (2003) noted that the basalt west of the DSFZ appears thickest near the Syria–Lebanon border, c. 15 km south of Crac des Chevaliers. If this point is matched against the Chorowicz *et al.* (2005) piercing point east of the DSFZ, then the left-lateral slip since the basalt eruption has been c. 35 km, much higher than any previous estimate. Our kinematic model (Fig. 13) predicts c. 26 km of left-lateral slip in this area since the present slip phase began at c. 3.7 Ma (assuming a constant slip rate). If the remaining c. 9 km of post-basalt slip occurred between c. 5 Ma and c. 3.7 Ma, then the contemporaneous rate would have been c. 6.9 mm/a, as at present. However, if this slip occurred between c. 6.5 Ma and c. 3.7 Ma, the rate would have been only c. 3.2 mm/a, consistent with the view (Westaway 2003, 2004) that prior to the present slip phase much of the AF–AR relative motion was taken up on structures east of the modern northern DSFZ (Fig. 1).

Comparison between local slip rates and predictions from the regional kinematics is also possible farther north, near Jisr esh-Shugur (Fig. 1). The DSFZ locally consists of subparallel active north–south-striking left-lateral faults across a c. 20-km-wide zone; this complexity is represented (from west to east in Fig. 1) by three subparallel faults, which Westaway (2003, 2004) called the Qanaya–Babatorun, Salqin, and Armanaz faults (Fig. 2). Our kinematic model (Fig. 13) predicts 6.9 mm/a of left-lateral slip in this area (c. 35.75°N), again indicating c. 35 km of total slip if this rate has persisted since c. 3.7 Ma. As Westaway (2003) noted, sedimentation in the adjacent Ghab Basin (Fig. 1), interpreted as a pull-apart basin at a leftward step between the Misyaf Fault and the Qanaya–Babatorun Fault (Fig. 2), began in the Pliocene, consistent with Pliocene initiation of the present geometry of faulting in this region.

The southern limit of lower Pliocene marine sediment is offset left-laterally across the Qanaya–Babatorun Fault by c. 10 km (Westaway 2003, 2004; ‘L’ in Fig. 2). Taking this as the total slip on this fault, with an age of c. 3.7 Ma, gives a time-averaged slip rate of c. 2.7 mm/a. East of Jisr esh-Shugur (‘M’ in Fig. 2), the southern end of the Salqin Fault offsets Early Pleistocene basalt (dated to c. 1.1–1.3 Ma; Kopp *et al.* 1999) by c. 1.1 km, giving a slip rate of c. 1.2–1.4 mm/a (see Mart *et al.* 2005, for a local map). Subtracting these rates from our model prediction (Fig. 13) suggests that c. 3 mm/a of left-lateral slip is taken up on the Armanaz Fault and any other subparallel faults in this area. The low left-lateral slip rate on the Salqin Fault was used by Mart *et al.* (2005) to argue against significant left-lateral slip on any part of the DSFZ (these authors have indeed claimed that no more than c. 20 km of left-lateral slip has occurred anywhere on the DSFZ). The fact now evident, i.e. that the modern geometry of the northernmost DSFZ has taken up only c. 20 km of left-lateral slip, reflects its young (Pliocene) age, and does not invalidate the generally accepted view (e.g. Freund *et al.* 1970; Garfunkel 1981) that >100 km of left-lateral slip has occurred on the southern DSFZ, which became active in the Middle Miocene.

Conclusions

We have reported four new Ar/Ar dates and 18 new geochemical analyses of Pleistocene basalts, from the Karasu Valley of southern Turkey, which have become offset left-laterally by slip on the N20°E-striking Amanos Fault. The geochemical analyses help to correlate some of the less-obvious offset fragments of basalt flows and thus to measure amounts of slip; the dates enable slip rates to be calculated. On the basis of four individual slip-rate determinations, obtained in this manner, we estimate a weighted mean slip rate for this fault of 2.89 ± 0.05 mm/a ($\pm 2\sigma$). We have also obtained a slip rate of 2.68 ± 0.54 mm/a ($\pm 2\sigma$) for the subparallel East Hatay Fault farther east. Summing these values gives 5.57 ± 0.54 mm/a ($\pm 2\sigma$) as the overall left-lateral slip rate across the DSFZ in the Karasu Valley. These slip-rate estimates and other evidence from farther south on the DSFZ are consistent with a preferred Euler vector for the relative rotation of the Arabian and African plates of $0.434 \pm 0.012^\circ \text{ Ma}^{-1}$ about 31.1°N, 26.7°E. The Amanos Fault is misaligned to the tangential direction to this pole by 52° in the transpressive sense. Its geometry thus requires significant fault-normal distributed crustal shortening, taken up by crustal thickening and folding, in the adjacent

Amanos Mountains. The vertical component of slip on the Amanos Fault is estimated as *c.* 0.15 mm/a. This minor component contributes to the uplift of the Amanos Mountains, which reaches rates of *c.* 0.2–0.4 mm/a. These slip-rate estimates are considered representative of times since 3.73 ± 0.05 Ma, when the modern geometry of strike-slip faulting developed in this region; an estimated 11 km of slip on the Amanos Fault and *c.* 10 km of slip on the East Hatay Fault have occurred since then. It is inferred that both these faults came into being, and the associated deformation in the Amanos Mountains began, at that time; before this the northern part of the Africa–Arabia plate boundary was located farther east.

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