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Scales of structural heterogeneity within neotectonic normal fault zones in the Aegean region

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Abstract—Structural heterogeneities within neotectonic normal fault zones in the Aegean region are important controls on the geomorphic expression of their uplifted footwall blocks. Contrasting fault rocks and slip-plane phenomena (corrugations, comb fractures, slip-parallel fractures and pluck holes) are responsible for mesoscopic inhomogeneities within fault scarps. Along-strike disruptions in the continuity of fault scarps (step-over zones, step-up zones, step-down zones, footwall cross-faults and hangingwall cross-faults) are products of perturbations in slip-plane geometry or intersections with transverse faults.

Large-scale variations in range-front morphology reflect contrasting patterns of deformation within fault zones. Some major fault zones, and many minor fault zones, are characterized by a migration of the active slip plane towards the hangingwall, resulting in a distributed network of high-angle faults and a range front which is stepped in profile. The presence of a localized mass of uneroded bedrock (hangingwall salients) protruding from the hangingwall of some major fault zones, results in intense deformation along a principal slip plane, and a range front which is ramp-like in profile. As a consequence of structural heterogeneities within neotectonic normal fault zones, fault scarps and range fronts in the Aegean region are subject to a temporally and spatially variable pattern of degradation.

INTRODUCTION

THIS paper synthesizes and interprets field data collected during a 5-year programme of surveying neotectonic normal fault zones in the Aegean extensional province (Fig. 1). Emphasis is given to three sectors of the Aegean region within which normal faults juxtapose Mesozoic carbonates with Quaternary deposits: (1) the Corinth region of mainland Greece (Fig. 1a); (2) the Izmir region of western Turkey (Fig. 1b); and (3) central and eastern Crete (Fig. 1c). In particular, we examine varying scales of structural heterogeneity that disrupt the continuity and geometry of the uppermost slip plane within a fault zone. As will be discussed later, the uppermost slip plane occurs at the bedrock–Quaternary contact and is generally the youngest or active slip plane within a zone.

The increasing recognition that earthquakes do not produce ruptures along the entire length of many faults but, instead, surface breaks are restricted to segments bounded by rheological and structural heterogeneities within fault zones, possesses considerable potential for seismic hazard assessment (Schwartz & Coppersmith 1984, Crone & Haller 1991). Although the fault segmentation concept has most commonly been applied to strike-slip faults (Schwartz & Coppersmith 1984, 1986), investigations of normal faults (Menges 1987, Wheeler 1987) and, in particular, the well-studied Wasatch fault zone in the eastern part of the Basin and Range province (Schwartz & Coppersmith 1984, Bruhn & Parry 1987, Bruhn *et al.* 1987, Machette *et al.* 1987, Wheeler & Krystinik 1987), reveal that they display comparable attributes to those of strike-slip systems. The most commonly recognized features are along-strike variations in the surface orientation and dip of faults (*geometric discontinuities*—dePolo *et al.* 1989) and intersections between faults and other major geological structures (*structural discontinuities*—dePolo *et al.* 1989).

An essential feature of fault segmentation studies is the ability to identify long-lived or persistent structural features which define *earthquake segments*—that is, parts of a fault that rupture as a unit during an earthquake (Barka & Kadinsky-Cade 1988, dePolo *et al.* 1989). In this context, normal fault zones possess an advantage in that a record of tectonic uplift may be reflected in the morphology of the uplifted footwall block. Morphological studies, based on correlation of erosional benches, along different parts of the Wasatch fault zone (e.g. Hamblin 1976, Anderson 1977, Osbourne 1978), for example, demonstrate spatial variations in the uplift history along the range front. In addition, morphometric analyses of certain geomorphologically-produced range front components, e.g. mountain-front sinuousity (Bull & McFadden 1977), valley spacing (Wallace 1978) and river gradients (Keller 1977), permit areas of contrasting degrees of tectonic activity to be identified (Bull & McFadden 1977, Mayer 1985, 1986, Rockwell *et al.* 1985). Although these studies express the long-held consensus that fault-scarp morphology is a product of geomorphological modification (Davis 1903, Gilbert 1928, Hamblin 1976, Wallace 1978, Nash 1986), more recent studies of range fronts have highlighted the role of non-uniform fault activity in influencing relief (e.g. Wheeler 1987). A study of the western Sangre de Cristo mountain front in

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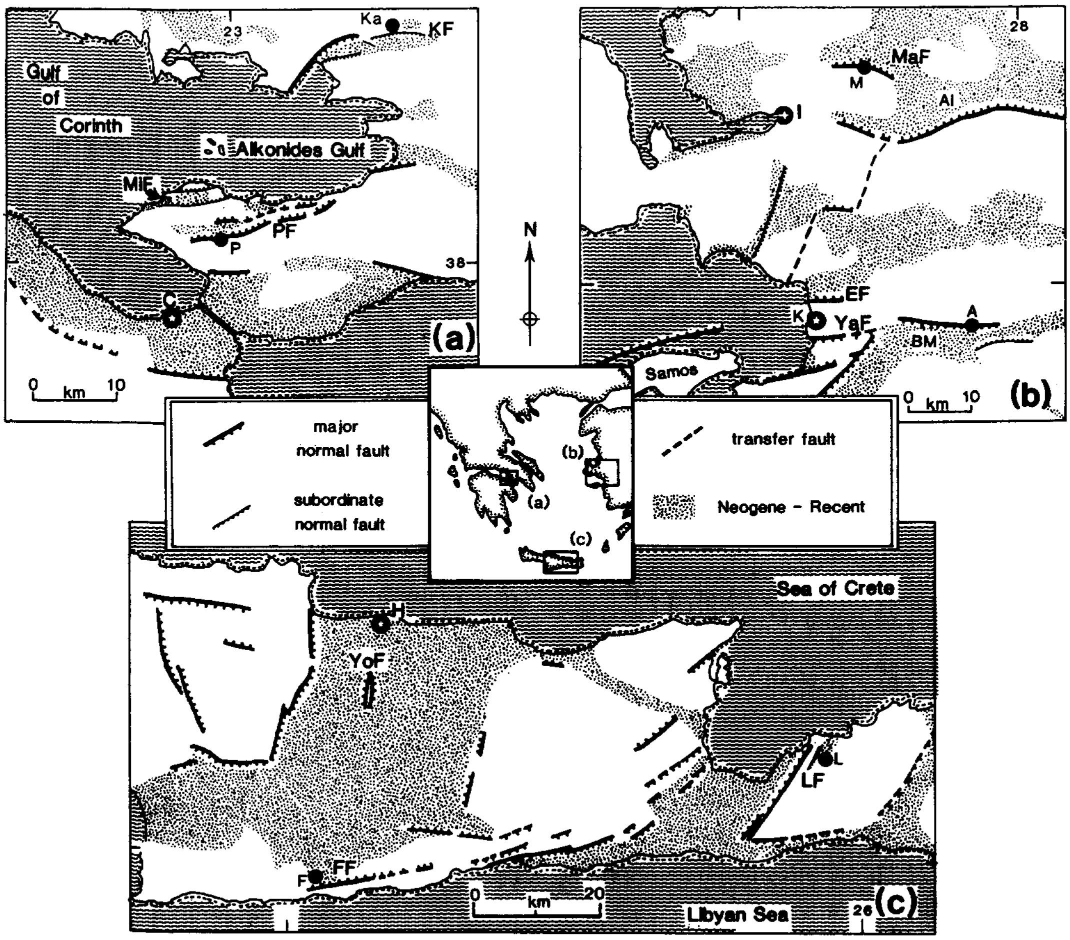

Fig. 1. Location of studied neotectonic normal fault zones. (a) Corinth region: PF, Pisia fault; KF, Kaparelli fault; MiF, Milokopi fault. (b) Western Turkey: MaF, Manisa fault; EF, Ephesus fault; YaF, Yavansu fault. (c) Crete: FF, Furnofarango fault, YoF, Youchtas faults; LF, Lastros fault. Grabens: BM, Büyük-Menderes; AL, Alesehir. Towns: (a) P, Pisia; C, Corinth; Ka, Kaparelli; (b) I. Izmir; K, Kusadasi; A, Aydin; M, Manisa; (c) H, Heraklion; F, Furnofarango; L, Lastros.

northern New Mexico by Menges (1987, p. 219), for example, showed that "... many of the geomorphic patterns of the range front may reflect large-scale variations in net displacement and structural geometry of the range front fault zone". It is becoming increasingly apparent, therefore, that the morphological characteristics of uplifted footwalls may reflect the distribution of major structural heterogeneities along normal fault zones.

According to Schwartz (1988, p. 13), "... faults are geometrically and mechanically segmented at a variety of scales. Segments may represent the repeated coseismic rupture during a single event on a long fault and may be tens to hundreds of kilometres in length, they may represent a part of the rupture associated with an individual faulting event and may only be a few kilometres long or they may represent local inhomogeneities along a fault plane and be only tens to hundreds of metres in length". Although only the largest scale irregularities influence rupture propagation (Schwartz 1988, dePolo *et al.* 1989) and range-front morphology (Menges 1987, Wheeler 1987), much of what is known about the internal structure of well-documented heterogeneities within strike-slip fault zones, such as fault jogs (step-overs) and fault bends (Sibson 1985, 1986, 1987), has been derived from analogous smaller-scale features, on the premise that fault attributes are self-similar (e.g. Sibson 1985, fig. 3). Despite this, there has been little attempt to examine the detailed forms of small-scale (10^1–10^2 m) heterogeneities within active normal fault zones. In this study we document the internal architecture of a variety of minor bedrock features, some of which have not previously been recognized in other normal fault zones (including well-studied examples in the western U.S.A.). Although many of the features occur at scales unlikely to affect the rupture characteristics of earthquakes, they do appear to be important controls on the geomorphic expression of fault zones.

FAULT-ZONE ARCHITECTURE

Neotectonic normal fault zones in the Aegean region are characterized by contrasting arrangements of fault rocks that reflect the activity of different deformation mechanisms during fault-zone evolution (Hancock & Barka 1987, Stewart & Hancock in press a). Minor fault zones, which are characterized by metres to tens of

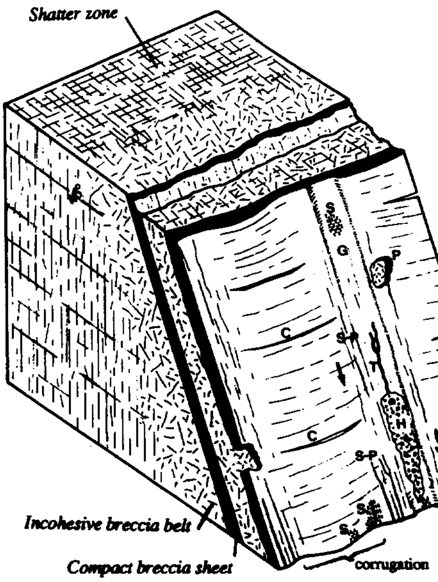

Fig. 2. Scaleless block diagram illustrating architecture within neotectonic normal fault zones. The uppermost layers within fault zones are characterized by alternating zone and slope-parallel *compact breccia sheets* and *incohesive breccia belts*, both within broader *shatter zones*. The uppermost slip plane is cut by a variety of slip-parallel and slip-normal structures. Slip-plane phenomena: C, comb fracture; G, gutter; P, pluck hole; S, frictional-wear striae; S-P, slip-parallel fracture; T, tool track; H, trail of brecciated material derived from hangingwall Quaternary sediments. Arrow indicates slip direction.

metres of offset, record a relatively simple structural history and display a well-ordered architecture, comprising alternating erosionally resistant *compact breccias* and erosionally weak *incohesive breccias* within a broader *shatter zone* of regularly fractured carbonate bedrock (Fig. 2). Compact breccias, which occur as indurated sheets of recomminuted and recemented breccia immediately beneath slip planes, form armoured carapaces on many fault scarps. Such carapaces protect underlying closely fractured incohesive breccias from degradation. The thickness of a compact breccia sheet is related to scarp height (i.e. amount of slip) and slip-plane inclination (Fig. 3). Along the Lastros fault zone in Crete, for example, the thickest compact breccia sheets (*ca* 1 m) generally accompany 10-m-high, 50°-dipping slip planes, while near-vertical, metre-high scarps are underlain by centimetre-thick compact breccia carapaces.

In contrast, major fault zones, whose longer tectonic history is expressed by range fronts several hundred metres high, commonly possess a more varied assemblage of fault rocks. In particular, uppermost slip planes within some major fault zones are underlain by stacked sequences of metre-thick sheets of intensely deformed and highly resistant fault rock containing stylolitic sutures and widespread recrystallized calcite. Commonly, these *stylobreccia zones* are confined to specific parts of major fault zones which, otherwise, are characterized by a similar layered architecture to that described from minor fault zones. As will be demonstrated later, this variation in fault-rock development along many major fault zones gives rise to marked lateral changes in range-front morphology.

SLIP-PLANE PHENOMENA

Slip planes are rarely uniformly smooth surfaces but are, instead, disrupted by a variety of tectonic phenomena (Hancock & Barka 1987). Some small-scale tectonic features, such as frictional-wear striae and tool tracks (mm-wide grooves along the slip plane), are well known slip-plane phenomena (e.g. Tjia 1972, Engelder 1974, Means 1987) but of greater significance from the perspective of scarp morphology are corrugations, comb fractures, slip-parallel fractures and pluck holes.

Corrugations are decametre-long undulations of a slip plane, the long axes of the corrugations being oriented parallel to the slip direction (Dumont *et al.* 1981, Hancock & Barka 1987). Commonly, corrugations occur in a continuum of size orders from ripple-like features to wave-forms that possess wavelengths and amplitudes of several metres. While the uppermost slip plane within minor fault zones is often characterized by a single corrugated slip plane that can be traced laterally for many metres within intermediate and large-scale fault zones, the uppermost slip surface comprises a network of laterally and longitudinally coalescing slip planes.

Commonly, the uppermost slip plane and its underlying compact breccia carapace within a fault zone are cut by *comb fractures* and *slip-parallel fractures*. Comb fractures, which subtend angles of 60–90° with a slip plane and, therefore, in profile, resemble the teeth of a hand-comb, give rise to intersection lineations on slip planes that are perpendicular to the slip direction (Hancock & Barka 1987). In contrast, slip-parallel fractures are high-angle (75–90°) structures which form an intersection lineation on a slip plane, that is aligned roughly parallel

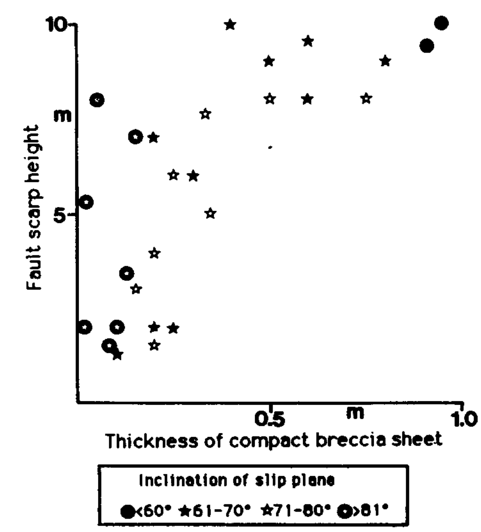

Fig. 3. Relationship between fault scarp height, slip-plane inclination and thickness of compact breccia sheets in Lastros fault zone. In general, the thickest compact breccia sheets are associated with high, moderately-dipping slip planes while the thinnest compact breccia sheets are associated with small, high-angle slip planes.

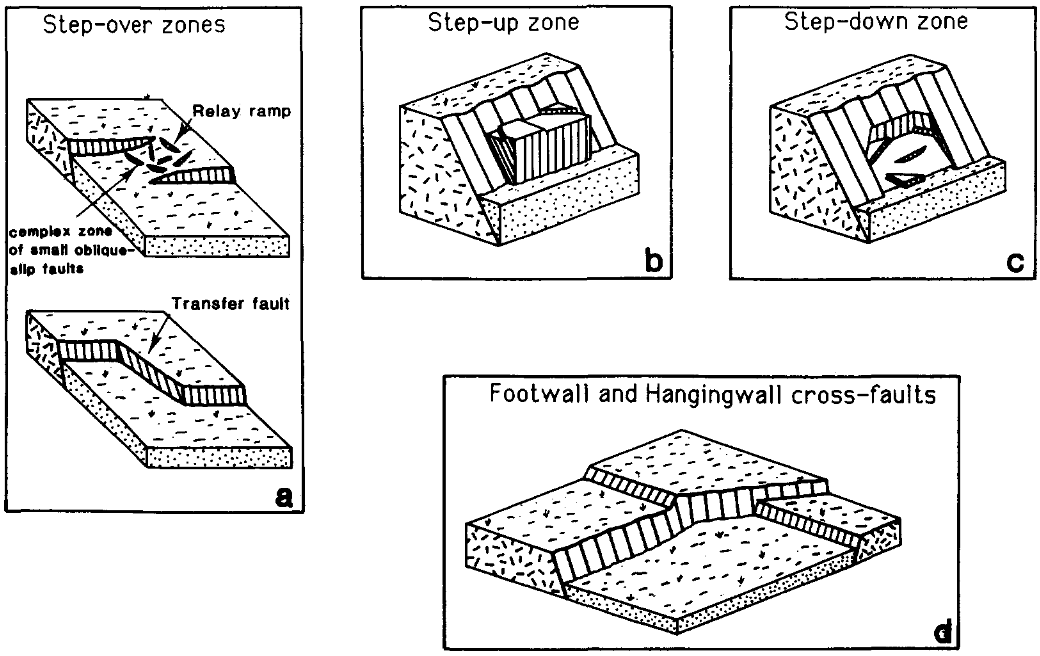

Fig. 4. Schematic diagram illustrating structural complexities within neotectonic normal fault zones in the Aegean region. (a) Step-over zones. (b) Step-up zone. (c) Step-down zone. (d) Footwall and hangingwall cross-faults.

to the slip direction (Stewart & Hancock in press a). The distribution of these fractures along a slip plane can vary because comb fractures are preferentially developed on corrugation crests, while slip-parallel fractures occur predominantly in corrugation troughs. Where slip planes are roughly planar there is generally an orthogonal trellis-like pattern of equally developed fracture sets. The lack of consistent abutting relationships between members of the two sets indicates that they contain approximately contemporaneous fractures. Because both comb fractures and slip-parallel fractures offset the youngest or active slip plane within a fault zone, their reactivation must post-date the last increment of fault motion. Possibly they result from post-seismic stress release.

Additional structures that disrupt the uppermost slip plane within many fault zones are *pluck holes*, centimetre- to decimetre-wide and deep cavities produced by sidewall plucking during fault movement (Hancock & Barka 1987).

GEOMETRIC DISCONTINUITIES

Neotectonic normal fault zones in the Aegean region display decametre- to hundred-metre-scale along-strike disruptions of the active or youngest fault trace. Although *step-over zones* are relatively well-documented components of both 'palaeotectonic' (Larsen 1988) and neotectonic (Jackson *et al.* 1982, Wallace 1984, Rosendahl *et al.* 1986, Frostick & Reid 1987, Wheeler 1987, dePolo *et al.* 1989) normal fault systems, *step-up zones* and *step-down zones* have not previously been reported.

Step-over zones

Step-over zones, also called fault jogs (Sibson 1985, 1986), occur where fault motion is transferred from one slip plane to another (Larsen 1988). Commonly, they are expressed as a right- or left-stepping en échelon arrangement of offset fault segments (e.g. Wallace 1984, Larsen 1988). Switching of fault motion between offset fault tips is generally accommodated by a zone of discontinuous and chaotically oriented oblique-slip minor faults within a monoclinal downwarp or *relay ramp* (Larsen 1988) (Fig. 4a). In the Corinth region of mainland Greece, for example, Jackson *et al.* (1982) observed that a zone of warping separating the Pisia fault and the Gulf of Corinth fault (Higgs 1988), both of which were active during the 1981 Corinth earthquake sequence, exhibited complex and discontinuous minor faulting and groundcracking. Occasionally, however, motion may be transferred via a discrete slip plane, commonly bearing oblique-slip striations (cf. a *transfer fault*, Gibbs 1984) (Fig. 4a). A step-over zone, for example, within the central part of the West Youchtas fault zone is accommodated via both complex networks of small, high-angle faults oriented obliquely to the main fault zone and by discrete oblique-slip planes (Fig. 5a). The step-over zone has transferred the uppermost slip plane from a 60- to 65°-dipping, 15-m-high slip plane to a series of high-angle (70–80°) slip planes that are only a few metres high. Immediately to the west, these high-angle slip planes become discontinuous, and within 10 m they have been removed by denudation.

Although they are most commonly identified at the scale of many hundreds of metres or a few kilometres (Jackson *et al.* 1982, Wallace 1984, dePolo *et al.* 1989), neotectonic normal fault zones in the Aegean region contain step-over zones at a variety of scales (Figs. 5a–c). Hancock & Barka (1987, fig. 3), for example, showed that the uppermost slip plane along a 500-m-long section of the Yavansu fault zone in western Turkey comprised a series of ENE-striking, decametre-long segments connected by shorter WNW-striking slip planes. Immediately west of the section investigated by Hancock & Barka (1987), the uppermost slip plane steps right-handed and northwards by over 100 m. Within this step-over zone there is an array of NW-striking slip planes (Fig. 5b). The northern slip plane can be traced eastward as a degraded scarp in the footwall of the fault zone. To the west of the step-over zone it is a fresh and generally striated slip plane separating bedrock from Quaternary deposits. While the inclination of the uppermost slip plane remains relatively constant (55–60°) on either side of the step-over zone, it progressively increases in dip towards the lateral tips of the fault zone, becoming, for example, 80–85° at its western limit. This increase in slip-plane inclination towards the western end of the fault zone is accompanied by its progressive degradation, indeed, in places it has been removed by erosion.

An excellent example of a step-over zone also occurs within the N–S-trending, E-dipping Lastros fault zone in Crete (Fig. 1c). For over 2 km along the northern end of the fault zone the main trace, which is marked by a fresh 60-70°-dipping, 6–10-m-high fault scarp, follows the lower slopes of an 800-m-high escarpment (Fig. 5c). About 1 km southwest of the village of Lastros, however, the lower trace dies out, via, initially, a near-vertical scarp several metres in height, and then, a low topographic ridge. The fault trace reappears, across a right-handed step-over, more than 200 m upslope. Here it is a bedrock scarp of similar height and inclination to that below. This scarp trends N–S for a further 2 km to the south. Streams cutting through the step-over zone expose several high-angle faults which strike obliquely to the main fault trace and which form conspicuous topographic ridges. In addition, bedding, which throughout the main segment of the fault zone dips gently (0–15°) towards the west, becomes downwarped in the vicinity of the step-over zone, generally dipping moderately (20–30°) southeast although immediately adjacent to slip planes the dip increases to 50° (Fig. 5c). Bedding is downwarped across a similar relay ramp towards the north at the northern end of the Lastros fault zone.

Step-up zones

Step-up zones are localized zones of high-angle faults interrupting a moderately-inclined fault segment (Fig. 4b). On the mesoscopic scale, step-up zones are discontinuous, near-vertical slip planes bounding slices of compact fault breccia which have been uplifted in the immediate hangingwall of pre-existing bedrock slip planes (Fig. 6a). Striations reflecting the last increment of fault motion, which, along adjacent fault segments, occur at the base of the bedrock scarp, are exposed on the near-vertical slip planes in the step-up zone. The

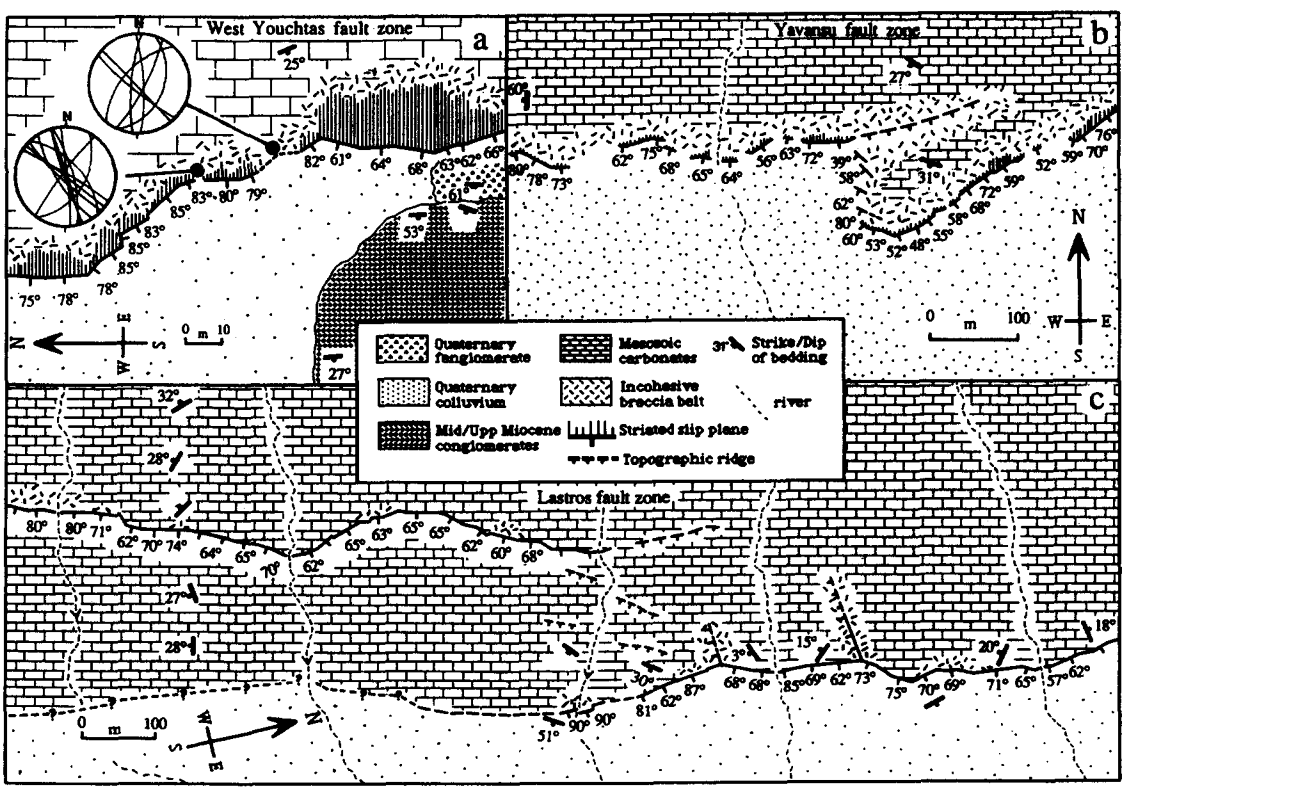

Fig. 5. Step-over zones within (a) the West Youchtas fault zone, (b) the Yavansu fault zone, and (c) the Lastros fault zone. Diagrams show how slip-plane inclination and height (lengths of vertical lines in ornament indicate heights directly proportional to horizontal scale) vary in relation to the location of step-over zones. 'Stereoplots' are lower-hemisphere equal-area diagrams.

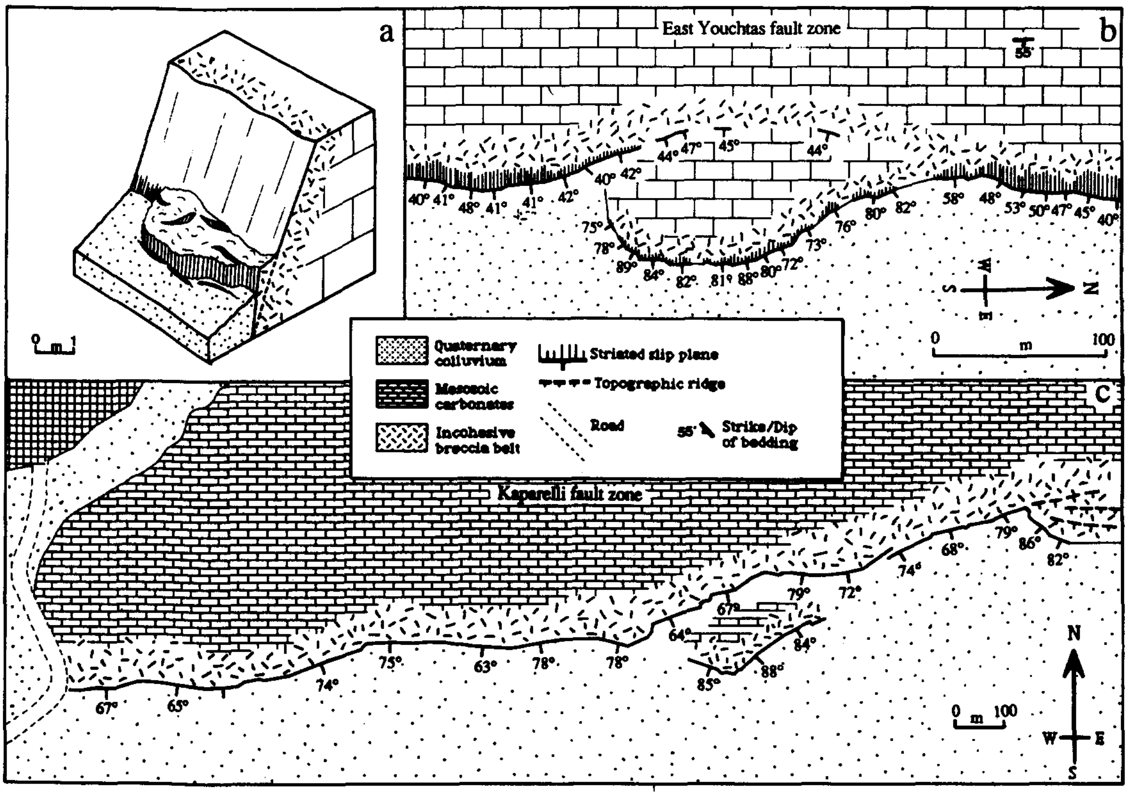

Fig. 6. Step-up zones within neotectonic fault zones in the Aegean region. (a) Schematic diagram illustrating general relationships. (b) Step-up zone in the East Youchtas fault zone. (c) Step-up zone in the Kaparelli fault zone. Note that in (b) and (c), moderately-dipping slip planes, several metres in height (vertical line ornament as in Fig. 5), are disrupted by shorter segments of high-angle to near-vertical slip planes and groundcracking. Cross-hatched ornament shows site of Kaparelli town.

older bedrock–bedrock fault contact is not reactivated. At some localities, blocks of compact hangingwall breccia have been sheared through and are 'stranded' several metres above the base of the fault scarp.

Although the above mesoscopic phenomena are likely to be short-lived at the free surface, some step-up zones are larger and more likely to be preserved for longer. Figure 6(b), for example, shows that along much of the central part of the East Youchtas fault zone in Crete the main fault trace, between bedrock and Quaternary materials, is defined by 5–15-m-high, moderately-inclined (40–50°) slip planes, generally bearing relatively fresh striations. Along a 200-m-long stretch of the fault zone, however, the active trace is marked by a belt of fissuring, extensive groundcracking and discontinuous, near-vertical (80–90°) slip planes, commonly 0.5–2.0-m-high and ornamented by striae. By contrast, although moderately-dipping slip planes can be traced several tens of metres into the footwall, they are laterally discontinuous, variable in height and do not display striations, indicating that they were not reactivated during the latest increment of displacement.

A comparable geometry is also displayed at the eastern end of the Kaparelli fault zone, an E–W-trending, S-dipping minor normal fault zone reactivated during the 1981 Corinth earthquake sequence (Jackson *et al.* 1982, Mariolakos *et al.* 1982). Along much of its length, a 0.5-m-high 1981 surface break reactivated a pre-existing 3-m-high limestone scarp that dips steeply (*ca* 65°) south (Fig. 6c). It also broke through previously unfaulted Holocene alluvium and colluvium in the immediate hangingwall of the fault scarp. In the central part of the fault zone, however, a several hundred metre-long bedrock scarp, located up to 200 m into the hangingwall of the main fault scarp, was also reactivated. In contrast to the main scarp, this 2-m-high, near-vertical wall shows only centimetres of offset, again accompanied by some groundcracking along the bedrock–Quaternary contact.

Step-down zones

Step-down zones are concave depressions within the footwall block of a normal fault zone (Fig. 4c). The clearest example of a step-down zone occurs as a 5-m-wide depression along the north-central part of the East Youchtas fault zone (Fig. 7a). The depression exhibits some characteristics in common with superficial slumps. For example, backwalls are defined by steep, planar or curved faults. The internal part of the step-down zone comprises a chaotic arrangement of oblique-slip faults cutting compact fault breccia. An adjacent segment of the East Youchtas fault zone contains a larger (decametre-wide) but less well-defined step-down zone (Fig. 7b). Along much of this part of the fault zone, the bedrock–Quaternary contact is defined by NNW-trending slip planes which dip moderately (50–55°) with well-developed striations that indicate that there was almost pure normal dip-slip. A few metres to the north of the mesoscale step-down zone, the uppermost fault

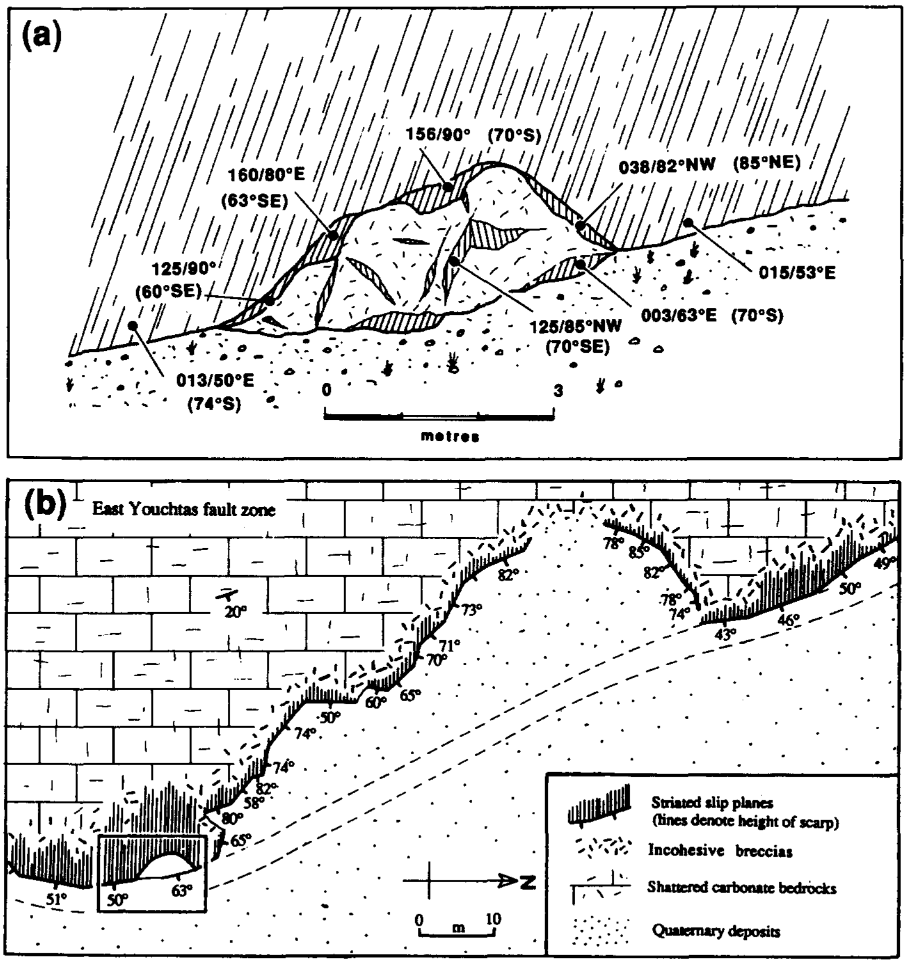

Fig. 7. (a) A metre-wide step-down zone disrupting a moderately-dipping, striated slip plane that defines the bedrock–Quaternary contact along the central part of the East Youchtas fault zone (Crete). (b) Decametre-scale step-down zone in the East Youchtas fault zone, in which a segment of moderately-dipping slip planes is disrupted by a zone of high-angle slip planes striking obliquely to the main fault zone. Locality of part (a) marked by box.

trace swings sharply towards the northwest, as a WNW-trending steep normal fault bearing obliquely pitching striations. Coincident with this change in fault strike, the height of the bedrock scarp is reduced from 15 to 3 m. Further to the northwest the fault scarp continues as a NW-striking, metre-high, steep to near-vertical slip plane until it dies out below a talus cone, only to reappear further north as a NE-trending scarp. Twenty metres towards the northeast, however, the scarp undergoes a sharp change in strike and dip, reverting to a NNW-striking, moderately-dipping (*ca* 50°) slip plane.

DISCONTINUITIES RELATED TO TRANSVERSE STRUCTURES

Structures trending obliquely or transversely to an active fault zone also influence its architecture. While some normal fault systems are reported to be disrupted by older thrusts (Smith & Bruhn 1984) or major lithological contacts (dePolo *et al.* 1989), important influences within active normal fault systems in the Aegean region are normal faults that are transverse to the main fault. These structures have been called *footwall cross-faults* (Hancock *et al.* 1987) where they segment the footwall block of a fault zone, and *hangingwall-cross-faults* (Hancock *et al.* 1987) where they segment the hangingwall block. Although, footwall cross-faults are widely recognized as prominent segment-bounding structures in the Basin and Range province of the western U.S.A. (Wheeler 1987, Crone *et al.* 1987, Schwartz 1988, Crone & Haller 1991), hangingwall cross-faults are less well-documented.

Footwall cross-faults

Investigations in the Aegean region have revealed that many of the studied fault zones are characterized by slip planes striking obliquely to and abutting the range-front fault zone (Fig. 4d). Commonly, these faults are associated with a sharp change in the strike of the main fault zone. A footwall cross-fault segmenting the Furnofarango fault zone (Fig. 1c), for example, is coincident with a 25° change in strike of the main range-bounding fault scarp, from WNW–ESE to NE–SW. In addition, in the Lastros fault zone (Fig. 5c) two prominent transverse faults, expressed as E–W-striking, 3–4-m-high ridges of compact breccia, occur at the edge of a step-over zone, and coincide with a 30–35° change in the strike of the main fault scarp.

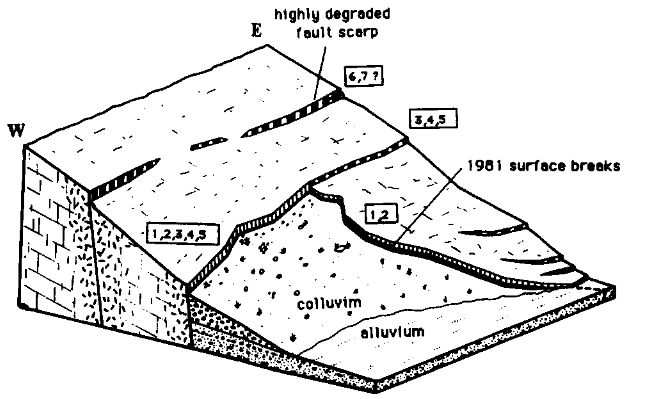

Fig. 8. Hangingwall cross-fault at the eastern end of the Kaparelli fault zone (Corinth). During the 1981 Corinth earthquake sequence, the active fault break followed the pre-existing ENE-striking fault scarp along much of its length but at its eastern end, motion was transferred to the NW-striking hangingwall cross-fault, leaving part of the ENE-striking scarp unreactivated. Numbers 1–7 refer to increments of motion, 1 being that during the 1981 events and 2–7 being successively older increments.

Hangingwall cross-faults

Downthrow of a hangingwall block during normal faulting is commonly accompanied by the formation of minor accommodation structures such as hangingwall cross-faults (Hancock *et al.* 1987, fig. 9a) (Fig. 4d). Hangingwall cross-faults are generally located near the lateral tips of active fault zones, distributing deformation within the proximal part of the hangingwall block. At the eastern end of the Kaparelli fault zone, for example, the main ENE-trending fault scarp continues eastward as a prominent though degraded scarp surface, despite the fault having reactivated a NW–SE-trending fault scarp in its immediate hangingwall (Figs. 6c and 8). It is interesting to note that King *et al.* (1985) suggested that the eastern end of the Pisia fault zone, a major E–W-trending, N-dipping fault zone which was reactivated during the 1981 Corinth earthquake sequence, is similarly characterized by N-trending, W-dipping hangingwall cross-faults.

FAULT-ZONE EVOLUTION

Range-front morphology is influenced not only by the distribution of slip planes within a fault zone but also by how they have propagated and migrated with time. We use the term propagation to describe the increase in area of an existing slip plane, and the term migration to describe the formation of a new linked slip plane in either the hangingwall or footwall of an older slip plane. Studies of fault zones in the Aegean region have identified two principal factors during fault zone evolution that strongly influence both range-front morphology and the persistence of fault zone heterogeneities.

Intrafault-zone hangingwall-collapse

In many active normal fault zones, both in the Basin and Range province of the western U.S.A. and the Aegean region, the youngest or active slip plane either coincides with the bedrock–Quaternary contact at the base of a range front, or it cuts Quaternary deposits in its immediate hangingwall (Slemmons 1957, Witkind *et al.* 1962, Wallace 1978, 1984, Jackson *et al.* 1982, Mariolakos *et al.* 1982, Mercier *et al.* 1983, Crone *et al.* 1987, Menges 1987). An additional important observation from the perspective of fault zone evolution is that slip-plane inclinations at the surface are commonly steeper than those at depth (the latter are known from fault plane solutions of earthquakes or geophysical surveys). For example, in the Wasatch fault zone (Smith & Bruhn 1984), the Lost River fault zone (Crone & Haller 1991) and the Pisia fault zone (Jackson *et al.* 1982, King *et al.* 1985), faults at depths of several kilometres are predominantly moderately-dipping (35–45°) and planar, although at the surface they may dip as steeply as 60–80°. A common explanation for such surface steepening of faults is that within Quaternary superficial materials, lateral confining pressures are low and, hence, the fault is expressed by a near-vertical tension fissure (Jackson *et al.* 1982, Jackson & McKenzie 1983, Mercier *et al.* 1983, Sébrier *et al.* 1985).

The architecture of many active fault zones in bedrock carbonates suggests that a comparable process also operates in bedrocks. The Kaparelli fault zone (Fig. 8), for example, displays a stepped morphology comprising a series of fault scarps that are less degraded towards the base of the profile. The main ENE-striking fault scarp experienced 0.5–0.7 m of displacement during the 1981 Corinth earthquakes (Jackson *et al.* 1982, Mariolakos *et al.* 1982), suggesting that the 3-m-high scarp may be a product of five or six increments of motion. In contrast, the similar-sized NW-striking fault scarp, which corresponds to a hangingwall cross-fault, experienced up to 1.6 m of offset during the 1981 events (Mariolakos *et al.* 1982) and, therefore, could have been formed by only two increments of motion. While the upper part of this scarp is slightly karstified, the upper part of the main ENE-striking scarp and, in particular, the 2-m-high section of the scarp that was not reactivated during the 1981 events, is markedly more degraded. Several decametres up slope of the main fault scarp is a laterally discontinuous and highly degraded 1–2.5-m-high scarp. This relationship suggests there has been a hangingwall-directed migration of the active slip plane, a characteristic which has been described by Stewart & Hancock (1988) from other fault zones in the Aegean region. Stewart & Hancock (1988) proposed that this *intrafault-zone hangingwall-collapse* possibly reflects a tendency for an upwards-propagating slip plane to cut through the more brecciated hangingwall block, rather than the more intact footwall block.

Hangingwall salients

Although the stepped appearance of many range-front fault zones reflects predominantly high-angle faulting within footwall blocks, some range fronts or segments of range fronts are characterized by decametre-high, moderately-inclined fault scarps at their base. These scarps are mainly underlain by metre-thick zones of highly resistant stylobreccias. More importantly, there is an association between the presence of stylobreccias and the occurrence of relic salient blocks of pre-faulting cover rocks in the immediate hangingwall of the fault zone.

The Pisia fault zone, which dips moderately (*ca* 40°) northwards beneath the eastern Gulf of Corinth (Fig. 1a) (Jackson *et al.* 1982, Vita-Finzi & King 1985) comprises a series of left-stepping segments (Jackson & White 1989). A detailed study of the architecture of the most western segment, a 2.5-km-long section at Pisia (Fig. 9), demonstrates that stylobreccias are confined to its eastern end where the fault zone is in contact with a prominent block of Mesozoic flysch. During the 1981 Corinth earthquake sequence, surface rupturing probably propagated eastwards (King *et al.* 1985), generally following the Quaternary–bedrock contact, but it died out against the block of flysch, reappearing a kilometre to the east where the fault zone is again emergent. As a result, the block of flysch in the hangingwall of this section of the Pisia fault zone acted as a barrier to surface rupturing. This eastern part of the Pisia fault zone displays a ramp-like morphology reflecting the occurrence of 30–50-m-high scarps, dipping at 39–45°, at the base of the range front. In contrast, the western end of the Pisia segment is characterized by a network of high-angle (60–70°) surface faults and possesses a step-type morphology comparable to that of minor fault zones. Such a variation in scarp morphology over a distance of about a kilometre is unlikely to be a product of deep-level structural heterogeneities. Instead, it is more likely that the load imposed by the greater volume of cover rocks in the eastern part of the Pisia segment inhibited splay faulting and intrafault-zone hangingwall-collapse.

A similar relationship between the presence of a ramp-type scarp and the occurrence of an erosional salient characterizes the N-dipping Furnofarango fault zone. At its western end, the 300-m-high range front displays a step-type geometry that reflects underlying high-angle (60–80°) faults. A section through the western part of the fault zone shows that these faults splay from a moderately-dipping (*ca* 40°) planar normal fault, several tens of metres below the scarp surface. Eastwards, the range front becomes lower and surface faulting is expressed by moderately-dipping (35–45°) slip planes, until in the extreme east the fault zone dies out laterally within an outcrop of Mesozoic flysch. This eastward decline in fault inclination is accompanied by an increase in the thickness of the uppermost compact breccia, or in places, stylobreccia carapace with metre-thick complexes of highly resistant fault breccia occurring in the most eastern part of the fault zone. Comparisons of slope profiles across the fault zone in its eastern and western parts, show that there is, towards the east, an increase in the amount of material overlying the hangingwall block.

Here, we propose that range-front morphology is partly controlled by the shape of the free surface towards which and through which, faults propagate (Fig. 10). Where the free surface is dominated by a large and prominent relic block of pre-faulting cover in the hangingwall, here called a *hangingwall salient*, intrafault-zone hangingwall collapse is inhibited and deformation is restricted to the main fault zone. The resulting intense deformation produces moderately-

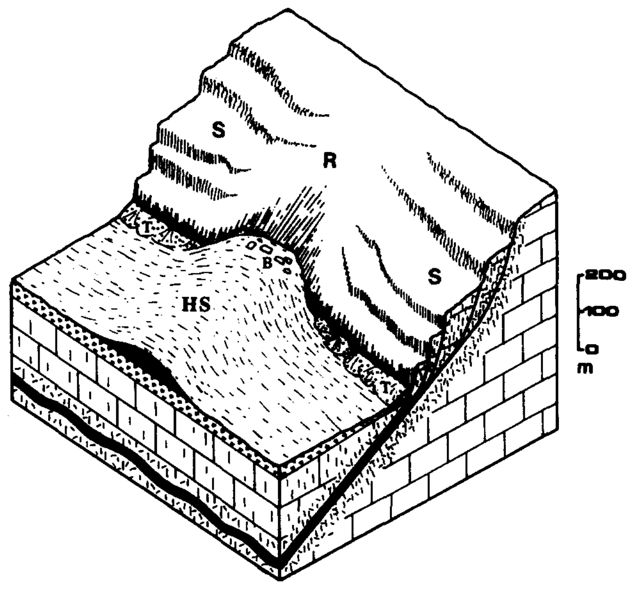

Fig. 10. Schematic diagram illustrating how range-front morphology varies with near-surface fault geometry. Where the free surface is dominated by a large and prominent hangingwall salient (HS), deformation is restricted to the main fault zone, resulting in moderately dipping stylobreccia zones and a ramp-type (R) scarp morphology. Where the hangingwall salient is absent, the relative lack of overburden promotes intrafault-zone hangingwall-collapse and the formation of high-angle splay faults which branch into the hangingwall of the main fault and result in a step-type (S) scarp morphology. Degradation of ramp-type scarps results in the creation of localized boulder fields (B) while degradation of step-type scarps produces extensive talus deposits (T).

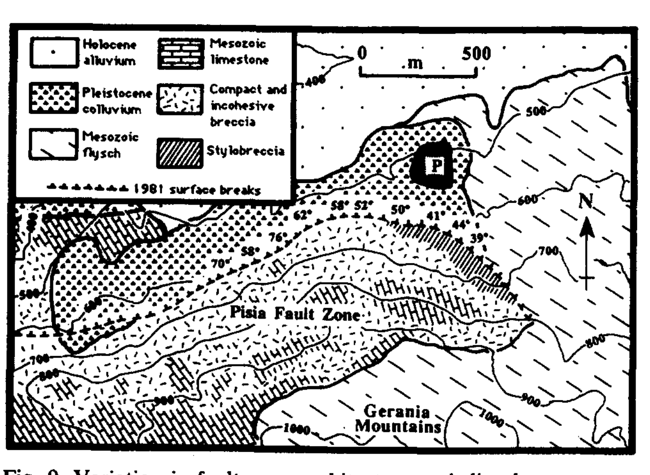

Fig. 9. Variation in fault zone architecture and slip-plane geometry within the westernmost segment of the Pisia fault zone (Corinth), near the village of Pisia (P). Coincident with the eastward decrease in slip-plane inclination and the change from a zone of compact and incohesive breccias to a zone of stylobreccias, is the eastwards thickening of a Mesozoic flysch wedge in the hangingwall of the fault zone. During the 1981 Corinth earthquakes, the active surface break followed the bedrock–Quaternary contact along the Pisia fault zone and died out within the flysch. Contours at 100 m intervals.

dipping, metre-wide stylobreccia zones which give rise to a *ramp-type* scarp profile (Fig. 10). Where hangingwall salients are not preserved as a result of erosion of the pre-faulting cover, the relative lack of overburden promotes intrafault-zone hangingwall-collapse and the formation of high angle (60–90°) splays in the hangingwall of the main fault. This process results in distributed deformation and a *step-type* scarp morphology (Fig. 10).

In addition to influencing the style of surface faulting, and, therefore, the degradational characteristics of fault scarps, the presence of hangingwall salients may also affect the larger-scale relief of uplifted footwall blocks. For example, the geometry of the West Youchtas fault zone, a W-dipping normal fault bounding the western flank of Youchtas mountain in central Crete, was modified by two salient blocks of Upper Miocene lacustrine sediments which stand proud of the surrounding topography (Fig. 11a). In contrast to the mainly high-angle (65–90°) slip planes along much of the length of the fault zone, slip planes in the vicinity of these upstanding blocks are inclined at 55–65°. Figure 5(c) shows that the transition from decametre-high moderately-dipping slip planes, adjacent to a salient block, to metre-high steeply-dipping slip planes, where a salient block has been eroded, is achieved by a step-over zone comprising a complex network of high-angle oblique-slip faults. A comparison of variations in net relief (measured as the height difference between the crest of the range front and the bedrock–Miocene contact) along the fault zone (Fig. 11b) shows that the central part of the West Youchtas fault zone consistently displays about 300 m of relief. This observation indicates that faulting and uplift probably generated the relief, because erosion and exhumation are more likely to create an irregular topography. Although footwall relief decreases markedly at both ends of the fault zone, particularly the southern end, an additional zone of reduced relief occurs in the central part of the fault zone, coincident with the most prominent salient block of Upper Miocene sediments. Although areas of reduced structural relief may be interpreted as zones of slip deficit and, therefore, as segment boundaries (Wheeler 1987), it is perhaps more likely that along the West Youchtas fault zone, relief reflects differential uplift on faults of contrasting inclination. Thus the same net extension perpendicular to a fault zone is accommodated by a much greater vertical offset along steep normal faults than along moderately-dipping faults.

EXHUMED FAULT ZONES

Because fault zone architectures comprising broad shatter zones and alternating incohesive and compact breccias have not been reported from studies of ancient, exhumed fault zones, we interpret these assemblages as being confined to the epidermis of the crust and, hence, having a low preservation potential (Hancock & Barka 1987, Stewart & Hancock in press b). In the central Arabian graben system (near Riyadh), for example, late Cenozoic erosion of a few hundred metres of limestone bedrocks has exposed early Cenozoic normal faults (Hancock *et al.* 1987). These *exhumed faults*, although accompanied by compact breccia sheets, do not display broad shatter zones or incohesive breccia belts, indicating that fault-precursor fracturing is restricted to the

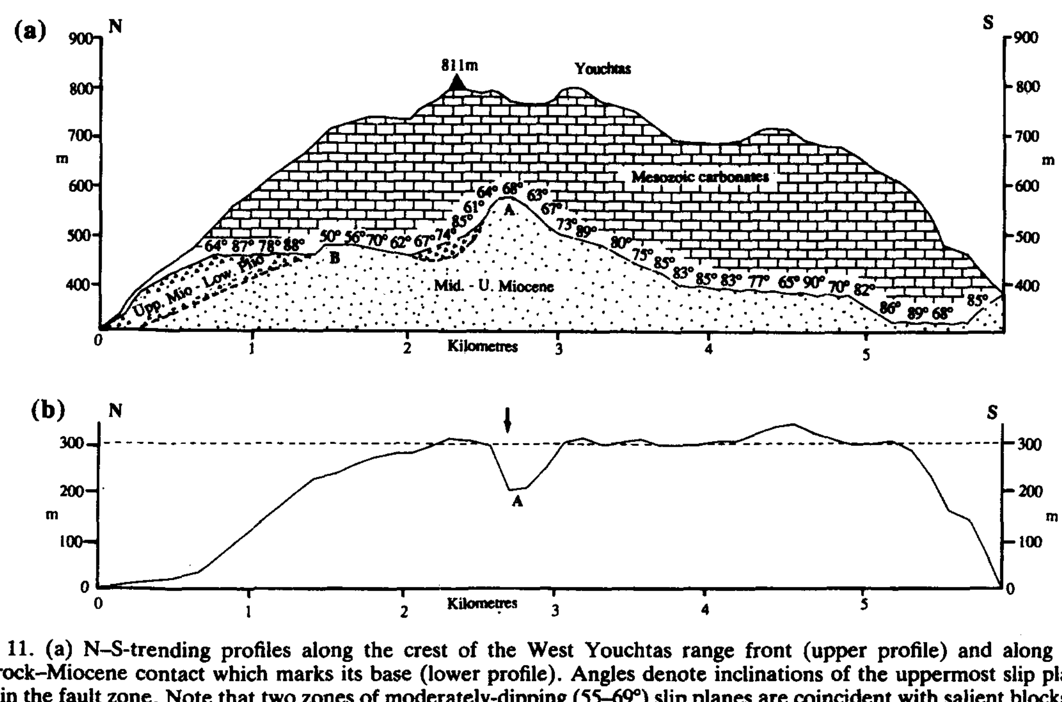

Fig. 11. (a) N–S-trending profiles along the crest of the West Youchtas range front (upper profile) and along the bedrock–Miocene contact which marks its base (lower profile). Angles denote inclinations of the uppermost slip plane within the fault zone. Note that two zones of moderately-dipping (55–69°) slip planes are coincident with salient blocks of Upper Miocene rocks (A and B) in the immediate hangingwall of the fault zone. (b) N–S profile showing variations in net relief along the West Youchtas fault zone. The main part of the footwall block displays about 300 m of relief except where a hangingwall salient (A) occurs. The reduced relief coincident with this salient is interpreted as reflecting differential uplift on faults of contrasting inclination.

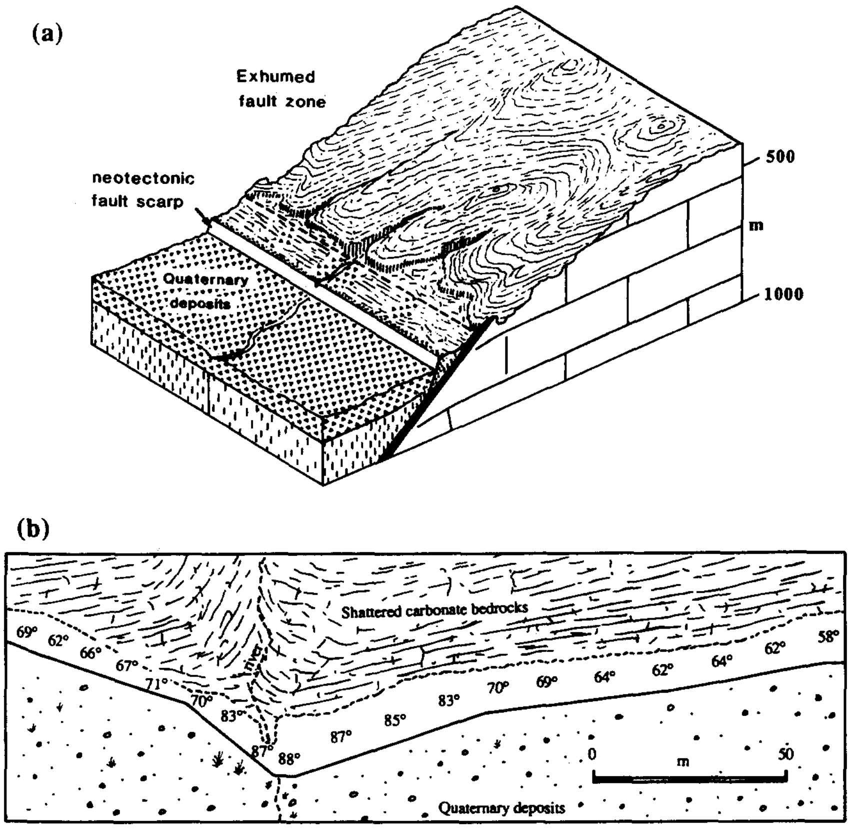

Fig. 12. (a) Schematic diagram illustrating the main characteristics of inherited fault zones. See text for details. (b) Changes in the geometry of the uppermost slip plane within the Lastros fault zone across an obsequent drainage channel. Slip-plane inclination progressively increases towards the river valley as a result of localized intrafault-zone hangingwall-collapse, initiated as a consequence of stream downcutting having eroded part of the immediate hangingwall of the fault zone.

highest levels of the crust while attrition and mineralization continue to accompany faulting at greater depths.

INHERITED FAULT ZONES

Although the Lastros fault is characterized by a metre-high fresh-looking scarp along its base, the fault zone displays many features indicative of it being a long-lived structure that was emergent before neotectonic reactivation. For example, the footwall block exhibits several hundred metres of relief, displays well-developed triangular facets and is dissected by valleys showing relatively gentle longitudinal profiles, all features typical of mature range fronts (Wallace 1978). Furthermore, a poorly preserved stylobreccia zone occurs in the footwall of the main fault scarp, while thick sequences of Quaternary colluvium juxtaposed with the fault zone are indurated and uplifted in places (Fig. 12a). In addition, and in common with the exhumed fault zones of central Arabia, the Lastros fault zone does not contain well-developed incohesive breccia belts. Well-bedded carbonate bedrocks are in direct contact with compact breccia sheets. We interpret these features as reflecting the neotectonic reactivation of an exhumed fault zone. Deformation along an already existing slip plane results in the creation of compact breccia sheets but not the development of fault-precursor fractures or breccia belts. We call such fault zones *inherited fault zones* in contrast to neoformed fault zones (nomenclature of Angelier 1989) that have propagated to the free surface through intact rock.

IMPLICATIONS FOR FAULT SCARP DEGRADATION

In this section we discuss the implications for fault scarp degradation of the heterogeneities described above. The topic is of relevance from a structural perspective because it is an axiom in many palaeoseismological investigations that scarp morphology can be used to estimate the amplitudes and recurrence of slip events.

The highly irregular or 'ragged' (Vita-Finzi & King 1985, p. 389) appearance of many neotectonic normal fault scarps in carbonate bedrocks in the Aegean region can be interpreted as reflecting the inherent structural heterogeneity of uplifted footwall blocks. Both structural heterogeneity and the uneven pattern of degradation it produces, can be seen at a variety of scales within the investigated fault zones.

Local or metre-scale inhomogeneities are produced by anastomosing slip planes. They result in the vertical and lateral overlapping and pinching out of compact breccia or stylobreccia sheets. In addition, unevenly thick armoured carapaces are randomly breached by tectonic structures such as pluck holes, and by erosional features such as sub-surface solution pipes. Comb fractures (on corrugation crests) and slip-parallel fractures (in corrugation troughs) are more regular structures breaching fault zone carapaces. Superimposed upon this structural template are the effects of unevenly distributed geomorphological agents. Surface water flow, for example, is preferentially channelled down corrugation troughs while vegetation, in the form of lichen and moss, which serve to retain moisture after rainfall, preferentially colonise the drier corrugation crests. An additional factor influencing scarp degradation is introduced by the irregular distribution, over space and time, of Quaternary talus which affords a degree of protection to scarp surfaces.

Degradational inhomogeneities may become enhanced, sustained or dampened by interaction with other structures. Corrugation troughs, for example, are often sites of particularly intense degradation because, in addition to compact breccia sheets being thinner beneath them, slip-parallel fractures are favourably oriented to be exploited by surface wash. As a result, localized talus cones are common at the bases of corrugation troughs. In contrast, corrugation crests are areas of relatively slow degradation, experiencing only the gradual widening of comb fractures.

Decametre-scale irregularities in the continuity of fault scarps are important factors in the lateral dissection of a scarp, particularly where they coincide with changes in slip-plane inclination. For example, because compact breccia sheets are generally thinner where scarps are low or steeply inclined, step-up and step-down zones are sites of increased degradation potential. This inference is supported by the observation that talus cones or sheets are common beneath these features. In general, step-over zones do not appear to mark a sharp change in slip-plane inclination or fault scarp height, although there may be minor changes in geometry in the immediate vicinity of fault tip-lines (such as in the Lastros fault zone, Fig. 5c). An exception is the step-over zone in the West Youchtas fault zone (Fig. 5a) where there is a change from a 20-m-high, 55–65°-dipping slip plane to a series of metre-high, 70–80°-dipping slip planes. Because several hundred metres to the north, slip-plane inclinations revert to a more moderate angle (Fig. 10a), it appears likely that this step-over zone represents one part of a more extensive step-up zone between two hangingwall salients. Within relay ramps of step-over zones, the complex network of small, high-angle faults also enhances degradation and talus accumulation. This effect can also be detected on a regional scale, such as in the southern Gulf of Corinth (Leeder & Gawthorpe 1987, Jackson & White 1989) and the East African rift system (Frostick & Reid 1987), where sedimentation is locally concentrated where active faults overlap. Intersections of footwall or hangingwall cross-faults with a main fault zone generally produce complex patterns of fracturing and brecciation. As a result, these junctions correspond to areas of intense, localized degradation, and are often accompanied by abnormally thick talus accumulations.

Large-scale (kilometre) variations in range-front morphology reflect changes in structural style and fault-rock fabric, rather than contrasting geomorphologic environments. Whereas intrafault-zone hangingwall-collapse induces a pattern of distributed deformation, the presence of hangingwall salients concentrates deformation along the main fault plane, and results in intensely deformed breccias which are eventually exposed as highly resistant stylobreccia carapaces. Thus, ramp-style fault scarps are subject to lower rates of geomorphologic modification than step-type fault scarps (although step-type scarps may be mantled by talus which inhibits degradation). In addition, ramp- and step-type fault scarps are subject to contrasting geomorphological processes. Extensive talus slopes banked against the bases of the western ends of both the Pisia and Furnofarango fault zones, for example, attest to progressive erosion by granular disintegration and rockfall. Localized boulder fields at the bases of the eastern ends of both these fault zones reflect the repeated 'peeling off' of stylobreccia sheets by slab failure.

Although the pattern of degradation that fault scarps and range fronts undergo in the Aegean region is strongly controlled by the style of faulting and the variety of fault rocks, the role of irregular relief, in the form of hangingwall salients, in determining the architecture and evolution of a fault zone is also critical. In addition evidence from the Lastros fault zone indicates that the surface expression of an inherited fault scarp is influenced by the topography that exists after prolonged exhumation and during neotectonic reactivation. Figure 12(b), for example, shows the details of the geometry of the Lastros fault zone along a 200-m-long section where the main 5–16-m-high fault scarp crosses a well-developed and deeply dissected stream channel, with the stream forming a hanging valley 8 m above the base of the scarp. Immediately below the incision of the steam channel, the slip plane is nearly vertical, although slip-plane inclinations decrease to 60–70° away from this point. The increase in slip-plane inclination towards that part of the slip plane breached by the stream is interpreted as indicating there was minor intrafault-zone hangingwall-collapse in the vicinity of the dissected valley, where erosion had led to a relative lack of overburden. Thus, over time, denudation of a fault zone may influence the style of subsequent fault behaviour.

CONCLUSIONS

(1) The morphologies of neotectonic normal fault scarps and range fronts in the Aegean region reflect inherent structural complexities at a range of scales.

(2) Minor fault zones, and some major fault zones, possess a relatively simple architecture of multiple high-angle slip planes underlain by alternating compact breccia sheets and incohesive breccia belts, both contained

within broad shatter zones. While incohesive breccia are relatively erodible, compact breccias are resistant to erosion, commonly forming slip-parallel and slope-parallel armoured carapaces on fault scarps. In many major fault zones, however, the fault zone architecture is more complex, with multiple zones of intensely deformed and erosionally resistant stylobreccias.

(3) Slip planes are ornamented by a diverse suite of meso-scale tectonic phenomena, the most important of which are metre-scale undulations of slip planes (corrugations) oriented parallel to the slip vector. In addition, slip planes are cut by an orthogonal network of stress release fractures oriented normal to (comb fractures) and parallel with (slip-parallel fractures) the direction of slip.

(4) The along-strike continuity of fault zones is disrupted by several types of geometric and structural discontinuities. While step-over zones, step-up zones and step-down zones reflect localized changes in the geometry of slip planes, footwall and hangingwall cross-faults serve to segment fault zones internally.

(5) Kilometre-scale variations in the style of deformation within fault zones record contrasting patterns of evolution controlled largely by the shape of the free surface towards and through which a fault propagates. Where there has been removal of a pre-faulting cover in the hangingwall block, the relative lack of overburden encourages upward-propagating slip planes to splay into the hangingwall block, a tendency here called intrafault-zone hangingwall-collapse. This process results in distributed deformation and a step-type scarp morphology. Where hangingwall salients (relic blocks of pre-faulting cover protruding from hangingwalls) occur, however, intrafault-zone hangingwall-collapse is inhibited and slip planes are confined to the main fault zone, within which there are metre-wide zones of intense deformation. These metre-wide zones of intense deformation contain moderately-dipping stylobreccia zones and result in a ramp-type scarp profile. In addition, hangingwall salients may act as barriers to fault rupture.

(6) As a consequence of the various scales of structural heterogeneities within fault zones, fault scarps are subjected to an irregular pattern of degradation, with the rates and types of geomorphic processes being largely controlled by fault zone architecture.

(7) Modifications to the shape of the free surface of a fault zone as a result of geomorphic exhumation can influence the behaviour of subsequent fault activity by encouraging or discouraging localized intrafault-zone hangingwall-collapse.

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