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Migration of activity within normal fault systems: examples from the Quaternary of mainland Greece

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Abstract

We examine five areas of mainland Greece where active extension occurs on sub-parallel systems of normal faults, and where geomorphological and stratigraphic evidence indicates that the faulting has migrated basinwards into the original hanging walls, in several cases within the late Quaternary. By comparing fault slip rates estimated from geomorphological data with current extension rates known from geodetic measurements, it appears that the newest faults can account for effectively all the present-day motions. Fault migration of this sort is easy to recognize in young systems close to sea level, because vertical movements of footwalls and hanging walls are obvious and reveal which faults are currently most active, but is less easy to confirm away from reliable reference levels or in older terrains with poorer time resolution. It is probably more common than is appreciated, and has a profound effect on syn-rift sedimentation and erosion patterns. Fault migration is probably an inevitable consequence of the interplay between stresses generated by the fault-related topography and the ultimate strength of major faults. It is likely to be further encouraged in places where lower crustal flow or rotations about a vertical axis are important. However, it is not clear why migration should preferentially occur into the hanging walls, as observed in central Greece.

1. Introduction

A striking feature of actively extending regions on the continents is the organization of the normal faulting into sub-parallel systems distributed over regions tens or hundreds of kilometres wide. This can be observed in many regions that are active today, such as the Basin and Range province of the western U.S.A. (e.g. Stewart, 1980), the Aegean Sea (McKenzie, 1978; Roberts and Jackson, 1991) and Tibet (e.g. Armijo et al., 1986), as well as in many older extended continental margins and basins. The major active normal faults in such regions have attracted much study, particularly using earthquake seismology, surface rupture in earthquakes and geomorphology, and as far as generalizations are possible, most are: (1) roughly planar in cross-section (e.g. Stein and Barrientos, 1985; Braunmiller and Nabelek, 1996) with (2) a quite restricted dip range of ~30–65° (Jackson and White, 1989) and (3) are generally segmented along strike, with a maximum segment length probably related to the local thickness of the seismogenic layer (Jackson and Blenkinsop, 1997; Scholz and Contreras, 1998). Within these systems of sub-parallel faults, activity is usually not distributed uniformly over the duration of the extension, but migrates from one fault system to another, either permanently or episodically, with such changes occurring on time scales that may be rapid compared with dating resolution in older terranes (e.g. Wallace, 1987; Coppersmith, 1989; Jackson and Leeder, 1994; Goldsworthy and Jackson, 2000). These changes are of great interest, particularly if patterns can be seen that give some clue as to why they occur. Not only may they influence factors of human or economic importance, such as earthquake hazard, the source and preservation of syn-rift sediment and the timing of trap formation relative to hydrocarbon migration, but they may also allow insights into fundamental questions of continental dynamics, such as the relationship between discontinuous deformation on faults in the upper crust and more continuous flow in the lower part of the lithosphere (see Jackson, 1999). Migration of fault activity on the spatial and temporal scales of interest here (10–20 km and less than 1 million years) is much easier to study on structures that have been active in the late Quaternary, where we can use earthquake seismology, surface faulting, active geomorphology and relative vertical motions, than in older terranes. In this paper, we concentrate particularly on the geomorphological features that result from fault migration,

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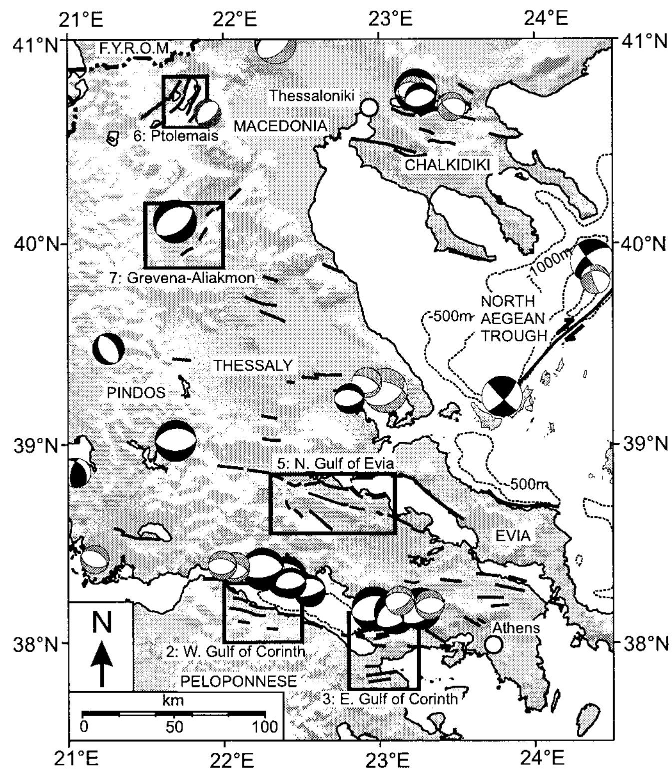

Fig. 1. Topographic map (illuminated from NE) showing the locations of the areas discussed in the text (black boxes). Also marked are major faults (black lines) and fault plane solutions of the main events since 1963 (pale grey from Harvard CMT catalogue, medium grey from first motion analysis, and black from P and SH waveform modelling).

which can often be recognized in surface slopes and drainage. It is these features that ultimately control how the evidence for fault migration is preserved in the structure and sedimentology of the geological record (e.g. Paton, 1992; Dart et al., 1995).

We describe five examples from the actively extending region of mainland Greece in which normal fault activity has migrated into the hanging wall basins on a time scale of about 1 million years. In each case, this involves uplifting and eroding hanging wall basins of the old faults in the footwalls of the new faults. Not all the examples are new and some, such as the western Gulf of Corinth, have been known for some time. The purpose of this paper is to gather these examples together, in each case summarizing the evidence for a shift in the active faulting and concentrating on the geomorphological signature left by the migration. This should then help the recognition of such fault evolution in other places where local conditions make it less easy to identify than in Greece. Another interest is in having several examples of the same phenomenon, all active today and all in the same region, so that a comparison can be made between them.

2. Tectonic setting

Mainland Greece (Fig. 1) is part of a broader region of active extension covering the Aegean Sea and western Turkey (McKenzie, 1978; Taymaz et al., 1991; Clarke et al., 1998; McClusky et al., 2000). It is one of the most rapidly extending regions on the continents today, with frequent normal-faulting earthquakes as large as M_w 6.6, corresponding to slip on faults approximately 10–20 km long. In the central area, between 38°N and 40°N, the normal faults occur with a WNW–ESE or E–W strike and approximately N–S slip vectors (e.g. Roberts and Jackson, 1991; Taymaz et al., 1991; Hatzfeld, 1999), and GPS measurements indicate a stretching rate across this region of 15–20 mm/year (Clarke et al., 1998). The discrepancy between the N–S slip vectors on the faults and the overall NW–SE motion across central Greece suggests that the faults rotate clockwise about a vertical axis as they move (e.g. McKenzie and Jackson, 1983, 1986; Jackson, 1994). Three of our examples are taken from this rapidly extending region of central Greece: two from the Gulf of Corinth and one from the Gulf of Evia (Fig. 1). In NW Greece, the strike of the active normal faulting changes to NE–SW with NW–SE slip vectors and the overall rates of extension are much less (e.g. McClusky et al., 2000). We discuss two examples from this region of slower extension: near Grevena and Ptolemais (Fig. 1).

3. The Gulf of Corinth

The Gulf of Corinth is the most rapidly extending graben

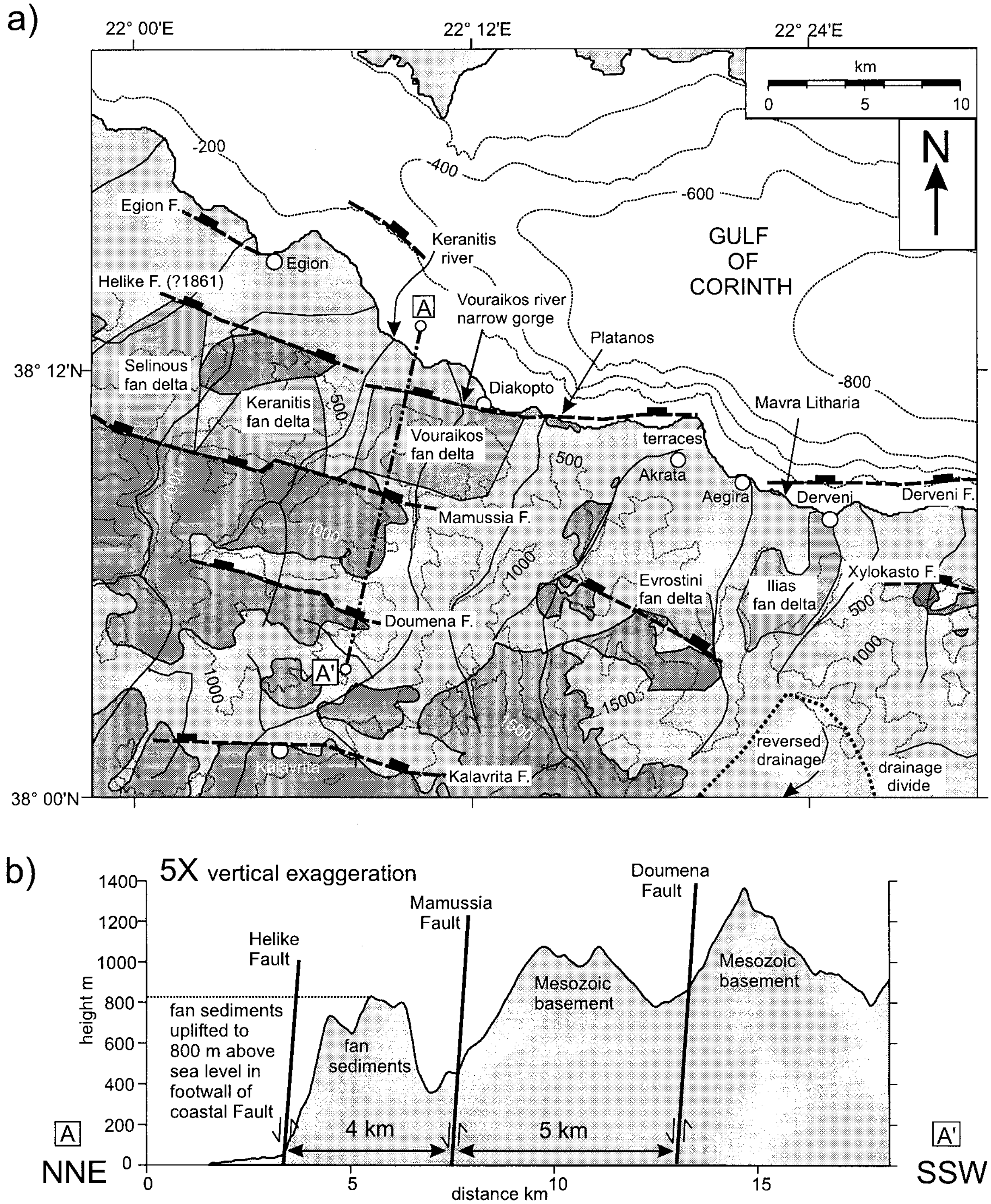

Fig. 2. (a) Simplified geology map of the western Gulf of Corinth region showing the location of the major uplifted fan delta deposits (medium grey), major rivers (black lines) and faults (thick black dashed lines). Outcrops of Mesozoic rocks are shown in dark grey and Neogene sediments in pale grey. (b) Topographic profile across the western Gulf of Corinth fault system.

system in Greece. It is 120 km long and up to 30 km wide, with a WNW–ESE trend that cuts the NW–SE structural grain created by thrust nappes in the early Tertiary. There is up to 3000 m relief between the highest footwalls on land and the deepest bathymetry offshore, with an additional 1000 m or more of syn-rift sediment seen in offshore seismic reflection profiles (Brooks and Ferentinos, 1984; Higgs, 1988). The graben is asymmetric, bounded by north-dipping faults along the southern coast that control the topography, bathymetry, sediment dips and vertical motions of the coastline. The faults are generally arranged in a right-stepping en échelon pattern, with maximum segment lengths of about 15–25 km (Roberts and Jackson, 1991; Roberts and Koukouvelas, 1996).

Extension probably started in places during the late Miocene, with some uncertainty because of the difficulty in dating largely unfossiliferous continental or lacustrine sediments (e.g. Bentham et al., 1991; Collier and Dart, 1991). However, much of the extension on the prominent E–W faults has occurred during the Quaternary (Doutsos and Piper, 1990) and is continuing today.

The eastern and western parts of the Gulf have different characters, though they share some evolutionary features, and we discuss them separately.

3.1. The western Gulf of Corinth

The western Gulf of Corinth is extending at a rate of ~13 mm/year (Clarke et al., 1997a). The north coast of the Peloponnese, on the southern side of the Gulf, is at the foot of a staircase of parallel north-dipping normal faults (Fig. 2), mostly within relatively soft Neogene sediments which nonetheless generate relief of up to 800 m across the faults (Fig. 2b). The faults are spaced approximately 5 km apart and typical segment lengths are approximately 10–25 km. Segmentation of the faulting is the principal control on the location of the largest fan delta systems entering the Gulf (Roberts and Jackson, 1991; Leeder and Jackson, 1993). This region has attracted much attention and only a brief summary of the relevant observations is necessary here.

The most active faults today are along the coast or offshore. Seismological and geodetic studies of recent earthquakes in 1992 and 1995 (Hatzfeld et al., 1996; Bernard et al., 1997) indicate major slip at depths of 4–10 km on offshore faults east of Egion. During the 1995 earthquake, a small amount of motion may also have occurred at the surface on the Egion fault (Lekkas et al., 1998; Koukouvelas, 1998). A damaging earthquake in 1861 caused fissuring and coastal subsidence along the line of the Helike fault between Egion and Diakopto, and may have involved rupture on that fault, though this is not certain (Ambraseys and Jackson, 1997). Other evidence of coastal fault activity is seen in the uplift of footwalls relative to sea level, including Holocene marine fauna uplifted up to 10 m above msl at Diakopto, Platanos and Mavra Litharia that indicate Holocene uplift rates of 1–1.5 mm/year (Papageorgiou et al., 1993; Stewart and Vita-Finzi, 1996; Stewart, 1996). Longer-term and larger-amplitude uplift is seen in uplifted marine terraces evident along the southern coast of the Gulf. These represent platforms cut during late Quaternary highstands, and which have been preserved through uplift in the footwalls of the coastal faults (Keraudren and Sorel, 1987; Armijo et al., 1996). Uplift rates are 1–2 mm/year over the last 300,000 years, requiring fault slip rates of ~10–15 mm/year (Armijo et al., 1996), in agreement with the present-day extension rates measured with GPS (Clarke et al., 1997a).

The most obvious evidence for migration of the fault activity is that the uplifted coastal terraces are in the hanging walls of the faults further south. In addition, the footwalls of the coastal faults contain delta deposits (thought to be Plio-Quaternary, but poorly dated), similar in nature to those forming in the Gulf today (Ori, 1989; Dart et al., 1994). The nature of these Gilbert-delta deposits, with continuous foresets reaching elevations of 1000 m above present sea level and back-tilted (i.e. south-dipping) topsets, implies both a basinward shift of the faulting and deposition and also considerable uplift relative to sea level. In addition, the morphology of the southern faults, which once controlled the location of the now-uplifted delta deposits, is much more subdued than that of the active coastal faults to the north (Armijo et al., 1996).

Drainage patterns are also helpful indicators of fault evolution. The locations of many modern rivers and fan systems entering the Gulf are controlled by segmentation breaks in the coastal fault system (Roberts and Jackson, 1991). In some cases, these same rivers may have been responsible for the older Gilbert-deltas now uplifted in the footwalls of the coastal faults, but have changed their course in response to the new outlets to the Gulf created by the segmentation of the younger coastal faults. The uplifted Keranitis fan delta (Fig. 2a), situated between the old Mamussia Fault and the young Helike Fault, provides an example of this process (Dart et al., 1994). Further east, two other rivers which used to drain into the Gulf have reversed their courses after failing to incise fast enough through the rising footwalls of the younger coastal faults (Dufaure, 1977; Dart et al., 1994; Armijo et al., 1996). The resulting wind gaps are now raised to elevations of 800–1200 m in delta sediments probably younger than 450,000 years (Armijo et al., 1996). Where the rivers have maintained their course and cut through the young uplifted sediments of the active footwalls, they form narrow steep gorges such as the Vouraikos River valley east of Egion, which is several tens of metres deep and has a minimum valley width of just 1 m (Stiros and Pirazzoli, 1998).

Thus, many authors have concluded that normal faulting activity in the western part of the Gulf of Corinth has migrated north during the Quaternary. Evidence supporting this view comes from the seismicity, geomorphology, sedimentology, drainage and vertical motions. Perhaps the most conclusive and useful is the uplift of the coastal footwalls, illustrating the importance of sea level as a marker. It is difficult to prove that the southern faults are completely inactive and there may be other causes of uplift besides footwall uplift. For example, microearthquake locations suggest that the subducted slab beneath the Peloponnese is relatively flat (Hatzfeld et al., 1989), raising the possibility that a regional component of uplift may come from the underplating of subducted sediments beneath the Peloponnese (e.g. Collier et al., 1992), as has been suggested for Crete (Angelier et al., 1982). However, even if the southern faults are still active and even if regional uplift is important, it is clear that the coastal faults are moving fast enough to keep their hanging walls below sea level, whereas the southern faults are not. Thus, the conclusion that the

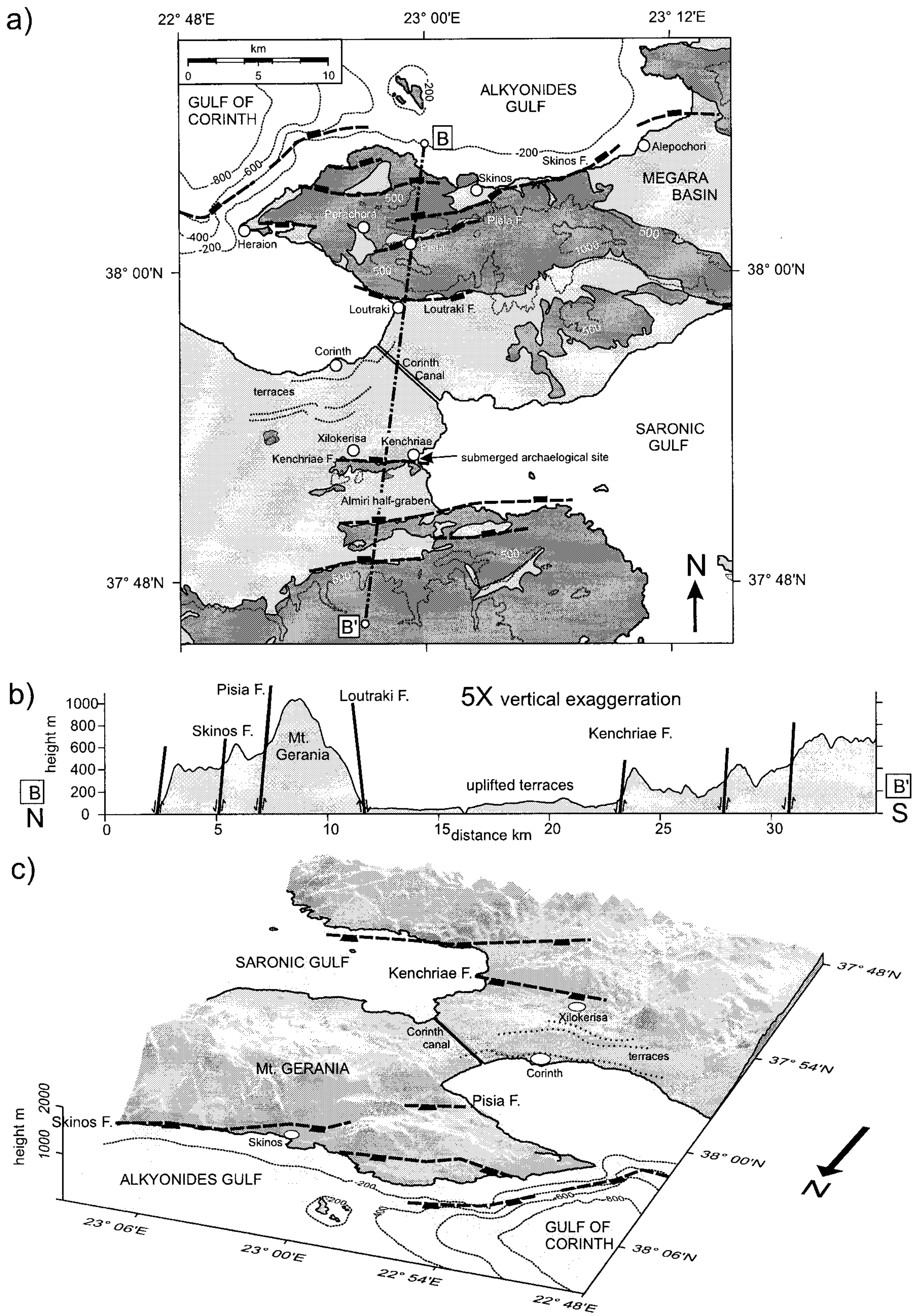

Fig. 3. (a) Simplified geology map of the eastern Gulf of Corinth region showing the location of the major faults (thick black dashed lines). Outcrops of Mesozoic rocks are shown in dark grey and Neogene sediments in pale grey. (b) Topographic profile across the eastern Gulf of Corinth fault system. (c) Illuminated perspective view of the topography in the eastern Gulf of Corinth region.

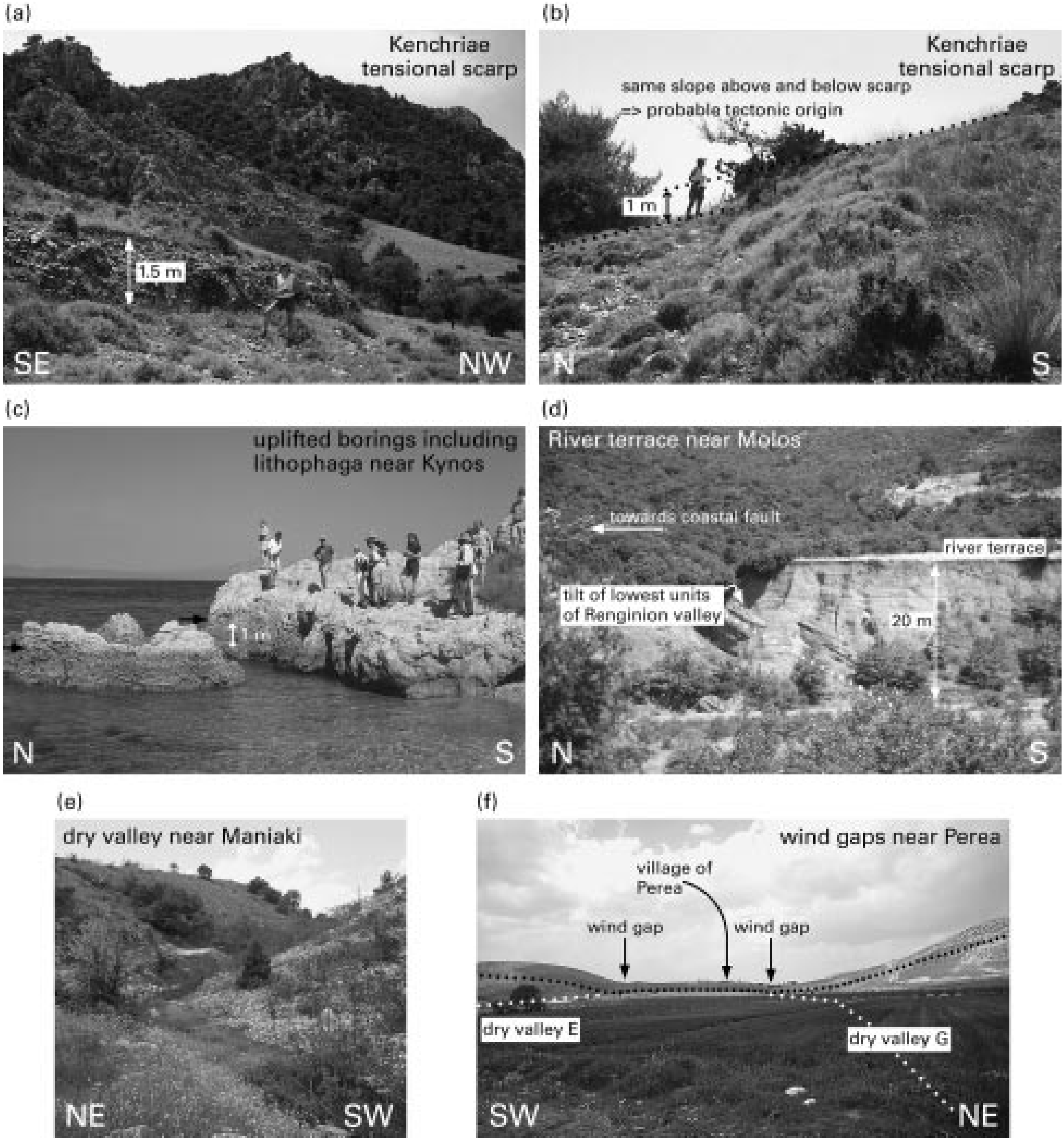

Fig. 4. Field photos. (a, b) Probable coseismic fault scarps adjacent to the Kenchriae fault at two sites south of Xilokerisa, discovered by Noller et al. (1997). In both places, the scarps are tensional features 1–2 m high in cemented scree, with free faces that would have originally been subvertical. They are both about 50 m away from the main limestone escarpment, in the hanging wall. (a) Eastern site, view south-west. (b) Western site view east. (c) View looking east at the level of the lithophaga borings within the limestone outcrop exposed along the coastline near Kynos. Black arrows indicate the height at which borings stop. (d) View looking east across a river valley showing the preserved river terrace 20 m above the river bed covering the dipping units of the lowest part of the Renginion valley sedimentary fill. (e) View looking south-east along a dry gorge preserved in the Perea region. (f) View looking north-west towards the wind gaps along the Perea fault either side of the village of Perea. These features lie along strike of dry valleys E and G.

northern coastal faults are the most active today seems inescapable.

From the known extension rates across the Gulf (~13 mm/year), the probable slip rates on the faults (~10 mm/year), and the uplift rates (~1–2 mm/year) and heights of the footwalls (~1000 m), we conclude that most of the morphology associated with the active coastal faults can be generated in less than 1 million years. It is therefore probable that the concentration of activity on to the coastal fault system also occurred in the last 1 million years (e.g. Dart et al., 1994; Armijo et al., 1996), consistent with limited palaeontological evidence in the exhumed fan deltas and their underlying formations. There is, however, little constraint on the geometry of the fault system at depth and in particular whether the sub-parallel surface faults merge on to a low angle fault or shear zone (e.g. Doutsos and Poulimenos, 1992; Sorel, 2000). An ~15° dipping zone between 8 and 12 km has been observed in both the microseismicity (Rigo, et al., 1996; Rietbrock et al., 1996) and an aftershock study of the 1995 Egion earthquake (Bernard et al., 1997). However, fault plane solutions for both the major events of 1992 and 1995 (Hatzfeld et al., 1996; Bernard et al., 1997) and some of the smaller events (Hatzfeld et al., 2000) indicate dips of 30–35°, which lie within the range typically seen worldwide (Jackson and White, 1989). It seems probable that the location of the gently dipping zone of microseismicity marks the base of the seismogenic layer rather than slip on a low angle surface (Hatzfeld et al., 2000). If the steeper normal faults do merge on to a low angle shear zone, this probably happens below the seismogenic layer.

3.2. The eastern Gulf of Corinth

GPS measurements (Clarke et al., 1997a) indicate that present-day extension rates at the eastern end of the Gulf of Corinth (Fig. 3) are slower (~6 mm/year) than in the west (~13 mm/year). This may be reflected in the water depth, which is only 360 m in the Alkyonides Gulf compared to 860 m in the deepest part of the Gulf of Corinth to the west (Heezen et al., 1966).

There are two main north-dipping fault systems in the eastern Gulf of Corinth (Fig. 3). In the north, a system of en échelon and sub-parallel fault segments occupy the Perachora peninsula and its offshore region to the north and west (Jackson et al., 1982; Roberts and Koukouvelas, 1996). A second system bounds the southern side of the Corinth isthmus and forms an array of three E–W-trending faults, spaced at 4-km intervals. All these faults form steep footwall ridges 300–500 m high in Mesozoic limestone. There are several indications that the Perachora system has been the more active of the two in the late Quaternary.

Once again, the most compelling evidence for relative activity of the two fault systems is in the geomorphology and vertical motions relative to sea level. On the Perachora peninsula, there is a strong correlation between the uplift or subsidence of the coastline and its position relative to the main fault segments (Jackson et al., 1982). Uplift in the footwalls is revealed by the preservation of marine terraces and fauna, including *Cladocora* corals and *Lithophaga* bivalves, many of which have now been dated to give uplift rates of 0.3–0.6 mm/year over the last 300,000 years (e.g. Collier et al., 1992; Pirazzoli et al., 1994; Dia et al., 1997). Uplift is also seen in the Corinth isthmus itself (Fig. 3), with a flight of marine terraces rising southwards from Corinth town to reach a height of almost 100 m near the fault at Kenchriae (Freyberg, 1973; Dufaure et al., 1975; Keraudren and Sorel, 1987). These terraces can be correlated with those further west, where their elevations increase (Armijo et al., 1996). Limited dating of corals in the isthmus area suggest uplift rates of ~0.3 mm/year (Collier et al., 1992; Dia et al., 1997). The decrease in terrace elevation with distance from the northern faults to the west of the Perachora peninsula (Armijo et al., 1996) and the tilted geomorphology of the Megara basin (Leeder and Jackson, 1993) suggest that the dominant cause of uplift is footwall uplift rather than any regional effect. Once again, the conclusion is that the northern set of faults are moving fast enough for their hanging walls to be subsiding, whereas the Kenchriae fault to the south is not. The (presumed) Plio-Quaternary marls and conglomerates of the Corinth basin in the isthmus region were deposited in a basin that may have been bounded to the south by the Kenchriae fault system (Collier and Dart, 1991). However, over the last ~300,000 years the uplifted marine terraces cut into those sediments, showing that the hanging wall of the Kenchriae fault has been more affected by uplift in the footwall of the Perachora system than by any subsidence related to activity on the Kenchriae fault itself.

This picture from the vertical motions is broadly confirmed by what is known of the faulting and extension rates. The fault segments at Pisia and Skinos ruptured in earthquakes in 1981 (Jackson et al., 1982; Hubert et al., 1996), and trenching on the Skinos segment indicated slip rates of ~1–3 mm/year over the last 1000 years (Pantosti et al., 1996; Collier et al., 1998). This estimate, together with footwall uplift rates of ~0.3–0.6 mm/year that imply fault slip rates of ~2–5 mm/year (e.g. Dia et al., 1997), suggests that slip on the Perachora faults can account for most of the present-day extension rate in this part of the Gulf, which is only about 3 mm/year across the Alkyonides Gulf itself (Clarke et al., 1997a).

Many historical earthquakes are known from this region because of the economic and strategic importance of Corinth, but contemporary accounts are nearly always too vague to associate these with particular faults (e.g. Ambraseys and Jackson, 1997). The discovery by Noller et al. (1997) of relatively fresh scarps of probable coseismic origin in the immediate hanging wall alluvium of the Kenchriae fault near Xilokerisa (Fig. 4a, b) is of great significance. The archeological site of Kenchriae, in the immediate hanging wall of the fault on the coast, was abandoned after earthquakes in the fourth century AD, and is now submerged 2 m below sea level (Stiros and Pirazzoli, 1998). It now seems likely that the Kenchriae fault was responsible, and is thus not completely inactive, even though the longer-term uplift of its hanging wall indicates it has been less active than the Perachora faults in the late Quaternary.

The heights (~500–800 m) and uplift rates (~0.5 mm/

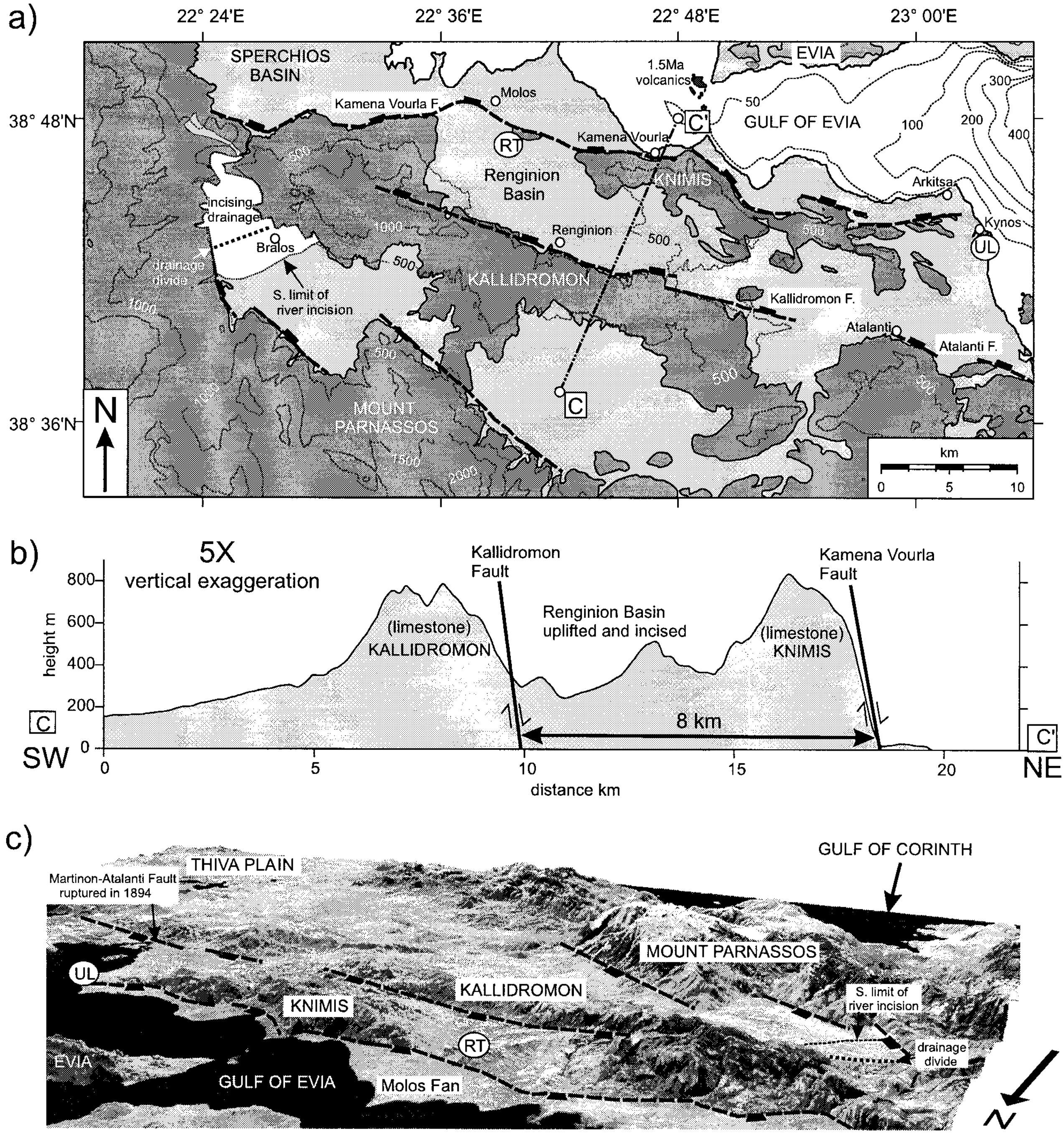

Fig. 5. (a) Simplified geology map of the North Gulf of Evia region showing the major faults (thick black dashed lines), and the location of the river terraces (RT) and uplifted lithophaga (UL). Outcrops of Mesozoic rocks are shown in dark grey and Neogene sediments in pale grey. (b) Topographic profile across the Kamena Vourla and Kallidromon faults. (c) Perspective view of the Landsat TM image overlain on the digital topography in the region of the North Gulf of Evia.

year) of the footwalls on the Perachora peninsula indicate that most of their morphology can be created in about 1 million years, suggesting a concentration of activity on to the northern fault system approximately 1 million years ago, as in the western Gulf of Corinth. In this case, the distance between the two fault systems of Perachora and Kenchriae is about 15 km, which is more than in the western Gulf (typically 5 km). On the other hand, each system contains several overlapping or en échelon segments. On the Perachora Peninsula, the geomorphology suggests a true en échelon arrangement, with one fault segment dying out along strike as another starts. They are probably all active now. Little is known about the faults south of Kenchriae.

4. The North Gulf of Evia

The North Gulf of Evia (Fig. 5) is a major graben of similar size to the Gulf of Corinth. Geodetic measurements and the seismicity of the last 300 years indicate that it is extending much less rapidly than the Gulf of Corinth, probably at about 1–2 mm/year (Clarke et al., 1998; Ambraseys and Jackson, 1997). The North Gulf of Evia is an asymmetric half-graben, but the polarity of the faulting and tilting changes midway along its length at about 23°15′E (Roberts and Jackson 1991). In this example we discuss only its western part, where the major faults dip north and the syn-rift sediments dip south (Fig. 5).

There are three main systems of faults in this region. A coastal fault system borders the North Gulf of Evia and Sperchios basin, extending for a distance of 100 km from Arkitsa in the east to Sperchios in the west, and uplifting Mesozoic limestone and Neogene sediments in its footwall (Roberts and Jackson, 1991; Eliet and Gawthorpe, 1995). The footwall ridge has a relief of up to 1 km, similar to the Gulf of Corinth, although the adjacent water depth is only 100 m, probably because of high sedimentation rates. The sub-parallel Kallidromon fault system (Fig. 5a) is approximately 8 km south of the coastal fault system, which may truncate it at its western end. The footwall of the Kallidromon fault is also largely Mesozoic limestone and rises 600–700 m above the Neogene sediments of the Renginion basin. The most southerly fault system has an NW–SE strike and borders the Mt. Parnassos range (Fig. 5a), which rises to a height of 2300 m above sea level.

As before, the geomorphology and vertical movements are the most convincing guides to the relative rates of activity on these fault systems. Of the three systems, the coastal faults have by far the clearest morphology, with clear, probably Holocene, escarpments (e.g. Jackson and McKenzie, 1999). By comparison, the Kallidromon and Parnassos fault escarpments are much more degraded, even though all three footwalls are mostly formed in the same Mesozoic limestone. At the western end of the Parnassos system, near Bralos (Fig. 5a), drainage systems in the hanging wall of the Parnassos faults are incised because of uplift in the footwall of the coastal faults bounding the Sperchios basin (Leeder and Jackson, 1993), showing that the coastal system dominates the geomorphology and is the more active of the two. Other indications of uplift in the footwalls of the coastal faults include uplifted river terraces which stop abruptly at the fault near Molos (Fig. 4d) and uplifted marine borings including *Lithophaga* near Kynos (Fig. 4c). ^{14}C dates on *Lithophaga* shells by Pirazzoli et al. (1999) are difficult to interpret, as the higher shells apparently have younger ages than the lower. If, instead, we assume that the top of the bored zone, at ~1 m elevation, formed when eustatic sea level stabilized ~6000 years ago, as in the Gulf of Corinth (Pirazzoli et al., 1994), then we estimate an uplift rate of ~0.2 mm/year. A similar figure is suggested by the prominent river terrace near Molos, which is 20 m above the river bed at the fault. If the river originally graded to a base level determined by the last high-stand at 125 ka, then an uplift rate of ~0.2 mm/year is inferred here too. These rates imply fault slip rates of the order of 1–2 mm/year, much slower than in the Gulf of Corinth, but compatible with the present-day estimates of extension rates from GPS and seismicity.

Thus, the Renginion basin, located between the Kamena Vourla and Kallidromon fault, is now being incised and appears more influenced by uplift in the footwall of the coastal fault system than by any subsidence related to slip on the Kallidromon faults. Furthermore, exotic granite pebbles are found at the base of the south-tilted syn-rift fluvial sediments between the Kallidromon and Kamena Vourla faults. Possible sources for this granite can be found to the north and east, but there is no known provenance to the south. For the pebbles to have reached their present location, they must have been transported into this basin before footwall uplift on the coastal fault isolated them from their source (Jackson, 1999).

Once again, the coastal fault system seems the most active and recent. Some historical earthquakes are known in the region, but most cannot be associated with a particular fault. Only one, in 1894, is known to have ruptured the surface and this occurred on the coastal fault system east of Atalanti (Fig. 5), causing coseismic subsidence of its immediate hanging wall (Ambraseys and Jackson, 1990). The age of the coastal fault system is unknown. On the basis of mammalian faunas in lignites near the base of the Renginion basin fill with an estimated age of 1.2 Ma, Ioakim and Rondoyanni (1988) suggest that the Kallidromon fault was active in the lower Quaternary. It is probable that the concentration of activity on the Kamena Vourla–Arkitsa coastal fault system occurred within the last 1–2 million years (Philip, 1974; Mercier, 1976).

5. Northern Greece

Recognition of the migration of fault activity in the three examples discussed so far is helped by the obvious vertical movements relative to sea level as a marker. In each case, the uplift of the footwall blocks of the coastal faults, which are also the hanging wall blocks of the faults in the hinterland, provides a clear indication of which faults are now dominant. Presumably, the migration itself is not caused by proximity to the coast, so this process should occur elsewhere, but may be more difficult to recognize without the help provided by sea level. We offer two possible examples below from Ptolemais and Grevena in inland northern Greece (Fig. 1). In both places, the drainage systems are strongly influenced by the tilted slopes created by the faulting itself, in contrast to the coastal-dominated systems that typify the regions to the south, and thus the drainage and its evolution are more critical to our understanding of fault migration than direct evidence of uplift. Indeed, it is

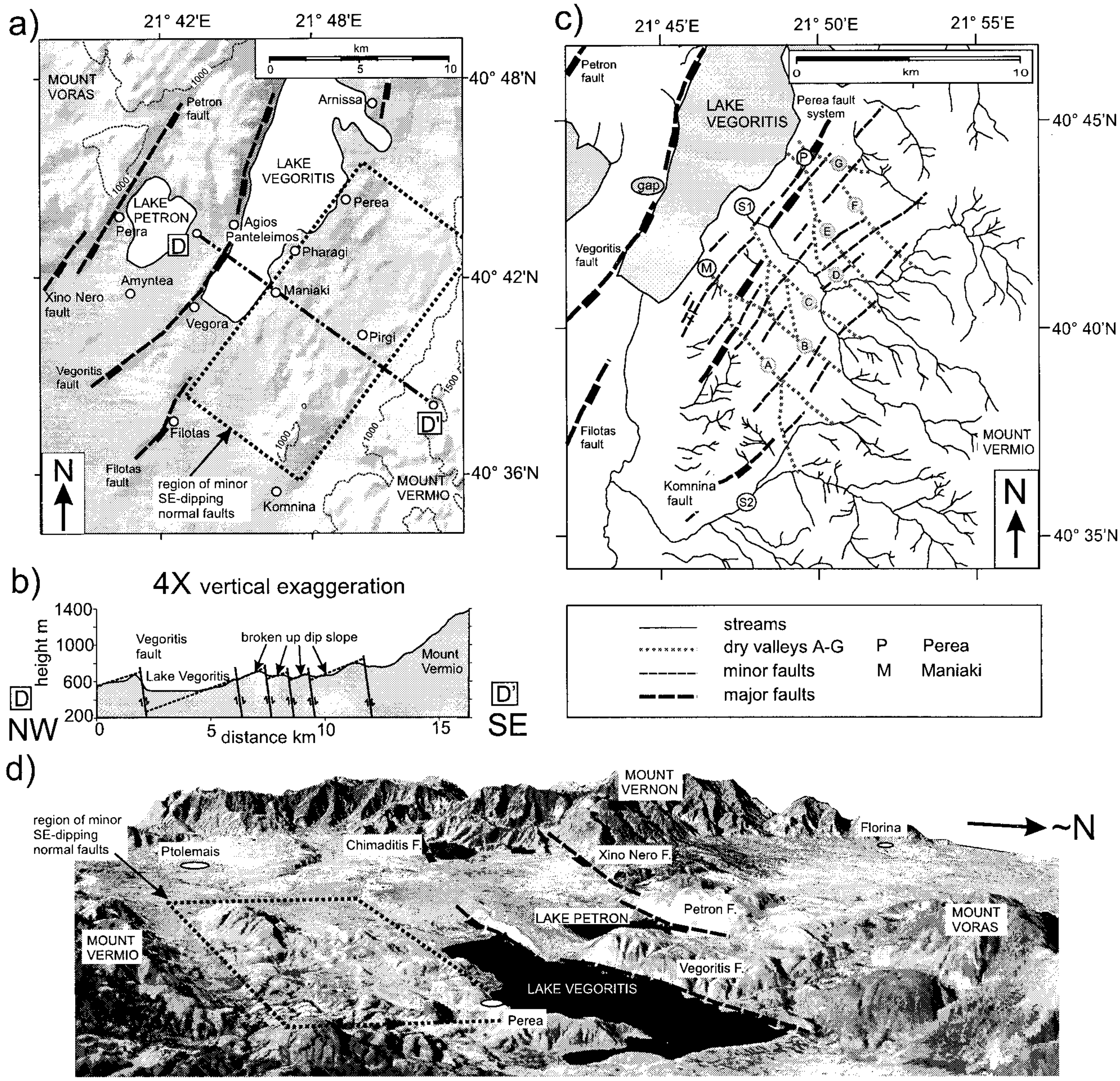

Fig. 6. (a) Illuminated topographic image of part of the Ptolemais–Florina basin of NW Macedonia, showing the location of the Petron and Vegoritis faults and the complexly faulted region around Perea. (b) Topographic profile across the densely faulted region showing the closely spaced small faults of the Perea region dissecting a possible older dip slope. (c) Drainage map of the Perea region, showing the two main stream networks S1 and S2, the faults, and the dry valleys A–G. (d) Perspective view of the Landsat TM image overlain on the digital topography in the Ptolemais–Florina basin.

hard to gauge genuine vertical motions in these interior regions of Greece because of fluctuating local base levels. For example, lake levels in the Ptolemais region have fallen rapidly in recent years through agricultural and industrial usage. Such changes can outweigh those produced by faulting and hence river terraces are often found within both the footwall and hanging wall blocks.

5.1. The Ptolemais basin

The Ptolemais basin in NW Macedonia contains a network of NE-trending normal faults that are younger than the lower Pliocene lignite–marl units uplifted and tilted in their footwalls (Pavlides and Mountrakis, 1987; van Vugt et al., 1998).

Of interest here is a parallel set of faults about 7 km apart, separated by the hanging wall lakes of Petron and Vegoritis (Fig. 6). Most of the faults in this region cut hard Mesozoic limestone that forms prominent footwall ridges and striated fault planes are exposed along several strands of the Perea system and along the Vegoritis fault (Pavlides and Mountrakis, 1987; Goldsworthy and Jackson, 2000). In

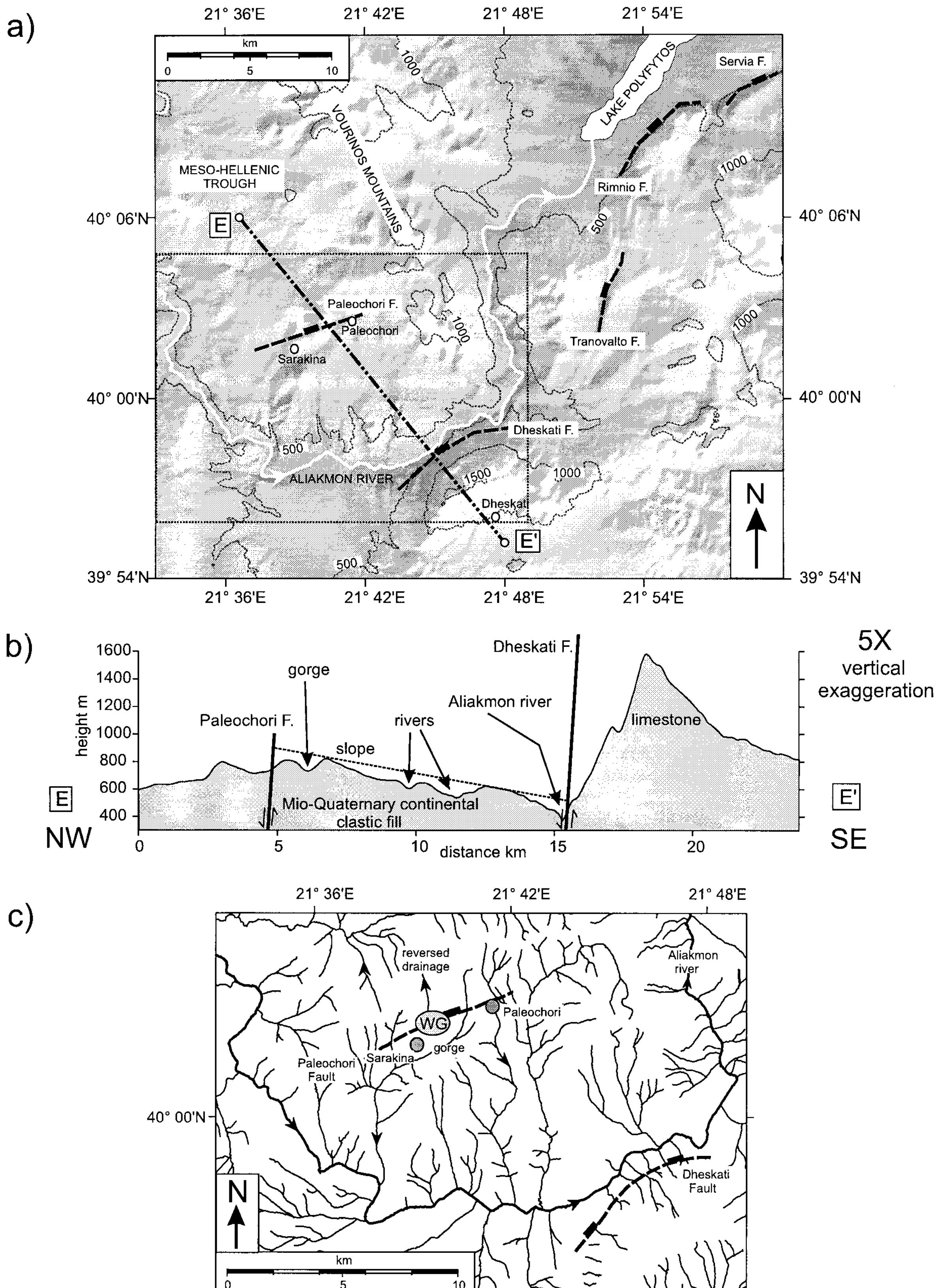

Fig. 7. (a) Illuminated topographic image of the Grevena–Aliakmon region showing the location of the major faults (thick black dashed lines). (b) Topographic profile across the Paleochori and Dheskati faults. (c) Drainage map of the Paleochori region showing the location of the wind gap (WG), reversed stream and gorge.

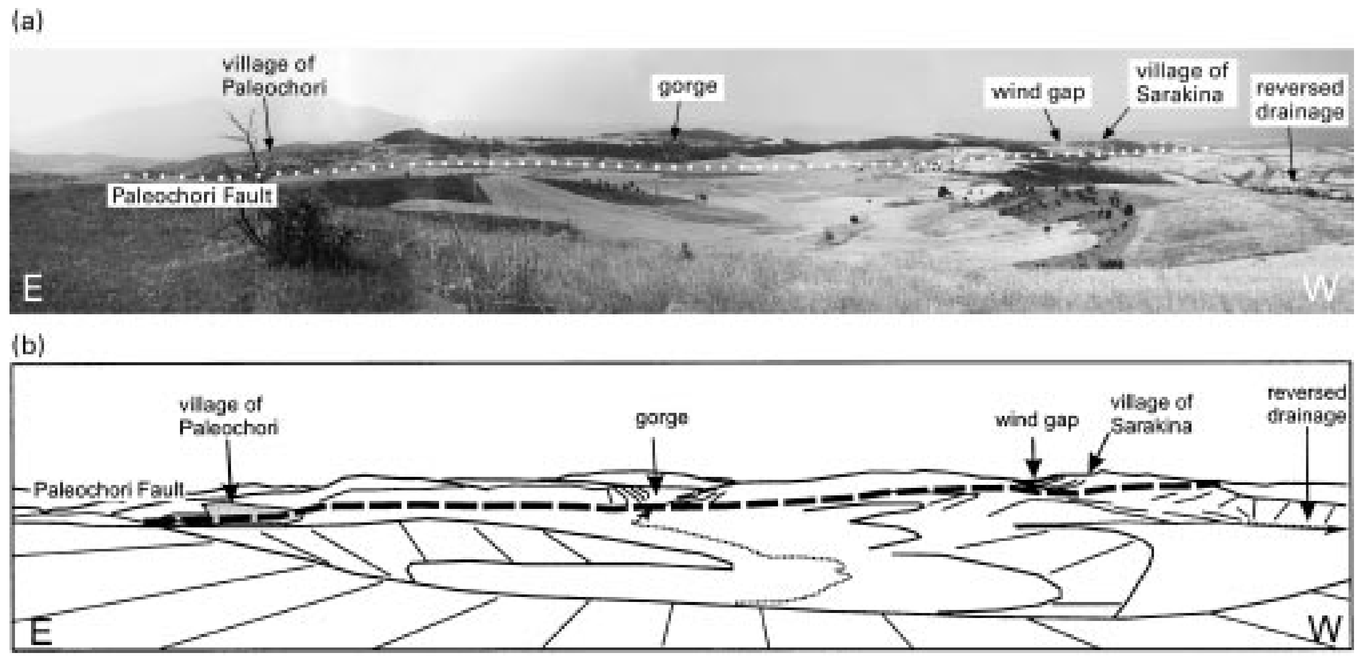

Fig. 8. (a) Photographic panorama looking south towards the Paleochori fault showing the location of the gorge, wind gap and reversed stream. (b) Sketch map of the region covered in (a).

1984, an earthquake of m_b 5.1, with a mechanism indicating normal faulting with a NE–SW strike, occurred close to the Perea fault system (Anderson and Jackson, 1987).

The Petron and Vegoritis faults are simple-looking structures, with single footwall escarpments rising 200–500 m above the basin. The Perea fault system, however, is different, composed of a series of parallel fault strands up to 5 km long, each with a much lower relief of typically 20–100 m, spread over a region 5–8 km wide (Pavlides and Mountrakis, 1987). The Perea faults are all in limestone and break up a recognizable 8° slope dipping towards the Vegoritis fault, cutting a system of NW-flowing streams that once drained into the lake. Many of the original stream courses are now preserved as dry valleys. The headwaters of these streams have now either been diverted round the ends of the fault segments or else captured and concentrated into larger catchments that are able to maintain gorges through the rising Perea footwalls, leaving a prominent flowing gorge (S1 in Fig. 6c) and several dry wind gaps (Fig. 4e, f). This geomorphology is discussed in greater detail by Goldsworthy and Jackson (2000). The evolution of the drainage implies that the system of parallel streams was established in the hanging wall of the Vegoritis fault and then interrupted by the later formation of the Perea faults.

There is little comparable evidence for the relative age of the Petron and Vegoritis faults. Lake Vegoritis is bounded on its NW side by two fault segments separated by a prominent low gap at Agios Panteleimos, which might once have been a drainage course flowing NW. If this route through the Vegoritis footwall system formed by a concentration of drainage into large catchments in the same way as gorge S1 (Fig. 6c) has formed through the Perea faults, it may imply that the Vegoritis fault is younger than the Petron fault, but this is obviously speculative.

The character of the Perea fault system may provide a key to understanding how new faults develop in this hanging wall position of a larger, more established fault (the Vegoritis fault in this case). The Perea faults are distributed and relatively minor, resembling the pattern described in theoretical simulations by Cowie (1998), in which the gradual growth of a distributed fault network causes strain to localize on a few major strands that coalesce (perhaps the main fault through Perea itself in this case), while others become inactive. This may be a more common process than is generally realized. In the Perea region, it is visible because the smaller faults are clearly exposed in the regular and almost barren limestone slope at the foot of Mt. Vermio, and are not obscured by soil cover.

5.2. The Grevena–Aliakmon region

Our final example is more ambiguous than the rest, but we include it to illustrate the difficulty in identifying fault migration in a place where the base level is controlled by a major river whose course is in turn influenced by the young faulting.

In the region of the Aliakmon river near Grevena (Fig. 7), NW-dipping normal faults occur in a NE–SW zone extending from NE of the man-made Polyfytos Lake to west of Dheskati. West of the Vourinos mountains, these faults truncate the NNW–SSE-trending Mesohellenic trough, an early Tertiary molasse-like basin (Pavlides et al., 1995). Of interest to this paper are the parallel faults of Dheskati and Paleochori, separated by 12 km. The footwall of the Dheskati fault forms a ridge in Mesozoic limestone rising 1000 m above the Aliakmon. The Paleochori fault to the northwest cuts a sequence of Plio-Pleistocene sediments at least 200 m thick and has created a much more subdued footwall ridge only ~100 m high.

The 1995 Grevena earthquake (M_w 6.5) was associated with slip on the Paleochori fault, though surface offsets were minor (~10–20 cm) and most of the coseismic slip failed to reach the surface (Meyer et al., 1996; Pavlides et al., 1995; Clarke et al., 1997b; Hatzfeld et al., 1997). Trenching across the Paleochori fault has revealed evidence for at least three prior events during the Quaternary (Chatzipetros et al., 1998). The Dheskati fault is not known to have moved in historical earthquakes, but the historical seismicity of the region has been very low, and locations poorly constrained (Stiros, 1998; Ambraseys, 1999).

The Paleochori fault has created a linear footwall ridge (Fig. 8), comparable to other active faults in Neogene sediments elsewhere in Greece (Goldsworthy and Jackson, 2000). In the footwall of the Paleochori fault is an incised surface slope in the Neogene sediments dipping south to the Aliakmon river. The surface is well preserved on the interfluves between incised streams and can be extrapolated to provide an estimate of the fault throw of ~120 m (Fig. 7b) and which also indicates only a small amount of erosion, suggesting a relatively young age. The footwall of the Dheskati fault is higher and steeper, as is typical of faults in limestone, but compared to the active faults of Pisia (Fig. 3), Kamena Vourla (Fig. 5) or even the Rimnio–Servia system along strike to the NE (Fig. 7a), the morphology of the Dheskati fault is subdued without the abrupt cliffs or exposed fault faces that are common for those other faults in limestone.

The main river draining the northern Mesohellenic trough and surrounding hills is the Aliakmon. It flows south down the basin before diverting through the limestone and ophiolite sequence of the Vourinos mountains in the immediate hanging wall of the Dheskati fault, and then north into the hanging wall basins of the Rimnio and Servia fault segments (Fig. 7a). It seems probable that this diversion of the drainage was facilitated by activity along these faults providing a lower course for the river. The Aliakmon is now incising in the hanging wall of the Dheskati fault and has eroded the base of the Paleochori footwall slope. This incision by itself is not conclusive: the whole Mesohellenic basin is incised with many abandoned terraces of the Aliakmon visible in gorges. Of greater interest are details of the drainage around the Paleochori fault (Fig. 7c), whose footwall contains a system of parallel streams flowing down the dip slope into the Aliakmon. Some of these streams originate NW of the Paleochori fault and are thus likely to have been established before it became active. There is no evidence that the streams have crossed the Paleochori fault by headward incision to the NW. One of these rivers has since ponded in the hanging wall and reversed its flow as it failed to incise fast enough to keep pace with uplift on the Paleochori fault. The old course is preserved as a wind gap in the ridge near Sarakina (Figs. 7c and 8).

The interpretation of the fault evolution depends on the origin of the sub-linear drainage pattern which is cut by the Paleochori fault and now flows down its footwall (Fig. 7c). If the N–S stream system formed on the hanging wall dip slope of a westward continuation of the Dheskati fault, then the formation of the Paleochori fault represents a migration of activity into the hanging wall of that older fault, as at Perea (Fig. 6). On the other hand, the N–S stream may exist simply because the Aliakmon flows E–W south of Paleochori. However, there are two other significant characteristics of the N–S stream pattern: (1) it seems better organized into a regular, linear system than other tributaries of the Aliakmon nearby (Fig. 7c), and (2) the streams flow down the dip direction of the recognizable surface preserved on their interfluves. These two characteristics seem more compatible with a tectonic origin of that slope and stream system than with an origin by headward erosion of streams from the Aliakmon. Since the Paleochori fault post dates that stream system, this interpretation implies a northward migration of the faulting.

Thus, although there is some evidence for the migration of fault activity into the hanging wall of the Dheskati fault, it is certainly equivocal and is less convincing than in the other examples. The ambiguity is mostly because the behaviour of the base level itself (the Aliakmon) is not well understood. We include this example to illustrate this point and to emphasize that the temporal evolution of fault systems is not likely to occur only where convenient base levels make it easy to recognize and confirm.

6. Discussion

The migration of normal faults into their hanging walls is not a new discovery and has been identified from stratigraphic evidence in a number of other places, including western Turkey (Paton, 1992; Dart et al., 1995) and various parts of the Basin and Range province of the western U.S.A. (e.g. Wernicke and Axen, 1988; Horton and Schmitt, 1998). In this study, we have concentrated on examples where some of the faulting is still active today, and we can therefore relate what we see to evidence from seismology and geodesy. In addition, we have used geomorphological data, not just to confirm that fault migration has taken place, but to see what happens when it occurs. With several examples in the same extensional province, it is worth summarizing what lessons there are concerning the likely consequences and causes of such fault evolution.

6.1. Consequences of hanging wall fault migration

The fact that we can identify several cases where normal faults have migrated into their hanging walls in mainland Greece and western Turkey (Paton, 1992; Dart et al., 1995), all within the same extensional province, suggests that this migration is an important process. Although much of the present-day extension of central Greece is localized in the Gulf of Corinth, hanging wall migration of the faulting is also seen in the Gulf of Evia and in northern Greece, where extension rates are probably 10 times slower. The total amount of extension on the faults discussed in Greece and also in western Turkey is not large, with β factors of about 1.2–1.3 estimated from changes in crustal thickness and from the tilting of Neogene sediments (e.g. Roberts and Jackson, 1991; Paton, 1992). These observations suggest that fault migration of this type may be common in other areas of modest extension, such as the North Sea, but how easy will it be to recognize?

The examples illustrated here suggest it may be difficult. In at least three places in mainland Greece (Figs. 2, 3 and 5) there is evidence that the switch of activity from one fault system to another can occur quickly, within the Quaternary. Furthermore, the en échelon arrangement of faulting along the margins of the major graben systems, such as the Gulf of Corinth and the North Gulf of Evia, often leads to an inactive fault being along strike from an active one. For example, the right-stepping relay between the coastal Arkitsa and Atalanti faults (Fig. 5a) causes the Atalanti fault, which ruptured in 1894, to be along strike from the probably inactive Kallidromon fault. Similar examples can be seen in the western Gulf of Corinth, particularly west of Egion, where the active Helike fault is along strike from the inactive Lakka fault to the west (e.g. Roberts and Jackson, 1991; Armijo et al., 1996). Thus, recognizing the processes described here in the older geological record may put severe demands both on the dating of stratigraphy and on the spatial resolution of lateral facies changes. However, the processes themselves have profound implications for synrift facies distributions and hydrocarbon potential as the hanging wall deposits of one fault are uplifted and re-worked in the next: a subject explored in greater detail by Dart et al. (1995). In the Gulf of Evia and the Gulf of Corinth, it is clear that basins can change from substantial subsidence to substantial uplift, erosion and re-deposition all within 1 million years.

An interesting structural consequence of the evolution described here concerns the origin of the arrays of tilted half-grabens that are so common in extended terrains. All the examples in this study show sets of sub-parallel faults dipping one way and the blocks between them tilted the other. It is commonly assumed that the blocks rotate simultaneously like dominoes or books on a shelf. This simple model is widely used and not only restores the pre-rift geometry satisfactorily (e.g. Proffett, 1977), but appears to give reasonable estimates of the vertical motions during extension (e.g. Yielding, 1990) and also of the total extension when compared with independent evidence, such as crustal thinning and subsidence (e.g. White, 1990). However, the parallel fault systems in the regions described in this study did not move simultaneously: they moved sequentially. Here is an interesting example of a model which, if not taken too literally, can yield useful estimates of some parameters related to the overall extension, but may be incorrect in its evolutionary implications, particularly of sedimentary processes during the extension.

6.2. Causes of fault migration

There are several possible reasons why a fault system might become inactive and another one start nearby but, as we shall see, explaining the direction of fault migration is more problematic.

A clue may lie in the stresses generated by the faulting and its associated topography. Any basin-and-range style topography requires stresses to maintain the lateral density contrasts, especially those at the surface. These stresses are likely to be contained within the elastic seismogenic upper crust (Maggi et al., 2000). For domino-style topography, the stresses required in the elastic layer depend on the wavelength (block width), amplitude (fault displacement), density contrast (between graben fill and footwalls) and thickness of the elastic layer. For typical block widths of less than 25 km, amplitudes of 2–3 km, an elastic thickness of 10–15 km and graben filled with compacted sediment, the stresses generated in the elastic layer are less than about 10 MPa (e.g. Jackson and White, 1989; Foster and Nimmo, 1996). This is a significant value, as it is approximately the largest stress drop seen in earthquakes (Scholz, 1982) and is probably close to the maximum strength of faults on Earth (see Lachenbruch and Sass, 1992; Bird, 1995, for reviews of this large subject). The dimensions of the fault blocks and topography in Greece lie within these bounding values, and so do not require shear stresses on the faults that are greater than the typical stress drops in earthquakes. However, the elastic restoring stresses increase rapidly with both fault offset (slip) and block spacing (Foster and Nimmo, 1996). Therefore, at some limiting offset, motion on an existing fault will cease, and a new fault will start up at such a distance away that the stresses generated by the topography do not exceed $\sim$10 MPa. It is the maximum offset on the fault that probably controls the maximum segment length, through a displacement–length relationship (e.g. Scholz and Contreras, 1998). In this scheme, the fault evolution is ultimately controlled by the strength of major faults themselves.

A new fault that forms in this way need not dip in the same direction as the original fault. However, if it dips towards the original fault, the two faults may intersect within the seismogenic layer if the spacing between them is less than about 15 km, as it is for all the examples shown in Greece. If the two faults are both active together over a limited time period, as they appear to be in the Corinth area where the Kenchriae Fault still seems to be active (Fig. 4a, b), then this requirement will favour new faults that dip the same way as the original. However, other explanations are also possible, the most likely being the reactivation of some older structural fabric.

None of these considerations explain why faulting should migrate into the hanging walls, rather than the footwalls, which is the pattern we see in Greece. To take the discussion beyond probable stress magnitudes and to specify actual locations and orientations of shear stresses is much more difficult and very dependent on the assumptions that must be made regarding the initial stress-free state and its relation to the topography. For example, the topography cannot be generated by elastic stresses acting on these faults alone, as the amplitude of the deformation far exceeds the elastic limit on faults, so the correct stress-free state is almost certainly not also topography-free (though this is always the easiest assumption to make). Some intuitive generalizations about relative effects are nonetheless possible: loading the hanging wall basin with sediment reduces the density contrast between the basin and its suroundings, thereby reducing the shear stresses in the hanging wall relative to the footwall. By contrast, footwall erosion reduces the topographic contrast between the footwall and its surroundings, thereby reducing stresses in the footwall relative to the hanging wall. However, to go further than this, and claim that either effect is dominant, or substantially increases the stresses, in one particular place (hanging wall or footwall), requires knowledge of the pre-existing stress state, which we do not have.

Fault migration may also be expected in regions where lower crustal flow occurs to smooth out lateral variations in crustal thickness. The flow is likely to occur as a front beneath the seismogenic upper crust, causing a topographic step that will move in the flow direction (McKenzie et al., 2000). Faulting at the topographic step will migrate with the flow, giving the ‘rolling-hinge’ evolution of faulting described by Buck (1988) and Wernicke and Axen (1988). However, such flow is only likely to be important where the lower crust is weakened through heating, caused either by a particularly thick crust or by magmatism. This cause of migration was probably important during the Miocene extension of the Basin and Range province of the western U.S.A. (Kruse et al., 1991) and the central Aegean Sea (Jolivet and Patriat, 1999), but probably not in the active regions of central Greece discussed here, where there is no evidence for substantial syn-rift magmatism and where variations in topography broadly correlate with variations in crustal thickness (Makris and Stobbe, 1984; Saunders et al., 1998), suggesting that lower crustal flow is unimportant there.

Finally, fault migration may be related to adjustments in the fault pattern that can be required in places where rotations about a vertical axis are important. This subject has been reviewed by Jackson (1994, 1999). In central Greece, where clockwise rotation rates about a vertical axis may average ~5°/Ma over the last 5 Ma, such rotations may be the origin of the discordance in strike direction between the old NW–SE faults that bound Parnassos (Fig. 5) and also the Megara basin (Fig. 4) and the roughly E–W faults that are active today (see Jackson, 1999). However, this process cannot account for systems in which the faults are sub-parallel, as in most of the Gulf of Corinth and northern Greece. In any case, even where the faulting migrates because of lower crustal flow or vertical-axis rotations, the basic control on geometry is likely to be exercised through the stress and strength considerations discussed above.

In conclusion, we can think of various general reasons why faulting may migrate, but none of these, except lower crustal flow (McKenzie et al., 2000), which we do not believe is important in central Greece, require it to do so towards the hanging walls rather than the footwalls. Yet that is clearly the pattern described in this paper. A proper explanation for this direction of migration requires a realistic dynamic analysis of the stress fields generated by the faulting, topography and regional loading.

7. Conclusions

Within the active extensional province surrounding the Aegean Sea, there are sufficient examples of normal faulting migrating basinwards into hanging walls to conclude that this is a significant structural process. Such migration is easy to recognize in active regions where sea level provides a base level, because the once-subsiding hanging wall block of an older fault can be seen to be uplifting in the footwall of a newer fault. It is more difficult to demonstrate in inland regions, but geomorphological indicators such as organized stream systems and fault-related dip slopes allow it to be recognized in some circumstances. In central Greece, the migration is evidently rapid, with the change from one fault system to another occurring within 1 million years. Without a reliable base level, adequate timing resolution and the ability to detect lateral facies changes, such migration may be difficult to recognize in old, inactive terrains, and is probably more common than we realize.

Fault migration of this type has a profound effect on the evolution of syn-rift sediment patterns, as the uplift, erosion and re-deposition of sediments can occur soon after they were initially deposited, and the distribution of the sediment supply itself is strongly influenced by fault segmentation. Fault migration is probably linked to the stresses generated within the elastic upper crust by the fault-related topography and the maximum stresses that can be supported by the faults themselves, which are likely to be of the order of 10 MPa. However, we cannot offer a simple explanation for preferred direction of migration into the hanging walls that we see in central Greece.

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