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Fault death: a perspective from actively deforming regions

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Abstract

Patterns of activity on large faults in regions of distributed continental deformation can change quite rapidly (in less than 1 Ma). Such changes include faults becoming permanently or temporarily inactive, spatial migration of activity between different faults or faults sets within a deforming region, and faults apparently going through episodes of intense activity separated by long quiescent periods. Greater knowledge of these changes, combined with our increasingly detailed understanding of the geometry of large faults, have implications for a variety of structural and tectonic issues, such as: (1) the nature of the interaction between neighbouring faults; (2) the link between the discontinuous deformation on faults in the seismogenic upper crust and the more distributed deformation in the aseismic lithosphere beneath; and (3) the relation between the instantaneous kinematics and the cumulative, finite deformation. Examples from the active extensional province of central Greece are used to illustrate the nature and possible significance of changing fault patterns. © 1999 Elsevier Science Ltd. All rights reserved.

1. Fault geometry, growth and interaction

When the *Journal of Structural Geology* celebrated its first 10 years, one of the topical issues was the geometry of large faults, particularly large normal faults in regions of extension (Jackson and White, 1989). Earthquake seismology had just come of age, in that new techniques (mostly involving synthetic seismograms) were available that could determine the faulting in earthquakes with sufficient accuracy to participate in debates in structural geology. At that point, some important structural questions became accessible through looking at active faults that moved in earthquakes, particularly when seismological data were combined with observations of surface rupture, geomorphology or geodesy (e.g. Stein and Barrientos, 1985; Yielding, 1985; Braunmiller and Nabalek, 1996). Techniques for resolving details of the geometry, segmentation and slip distribution on large active faults have become ever more powerful, and concern is now about the significance of differences (which 10 years ago would have been regarded as subtle) between results obtained from seismology, GPS surveying and SAR interferometry for the same earthquake: see, for example, studies of the 1994 Northridge, California, earthquake by Massonet et al. (1996) and Hudnut et al. (1996), or studies of the 1995 Grevena, Greece, earthquake by Clarke et al. (1997a) and Meyer et al. (1996).

Over the period of these developments, there has also been considerable interest in how faults grow or evolve with time, pioneered by the theoretical work of Watterson (1986), Walsh and Watterson (1988) and Cowie and Scholz (1992a), based mainly on empirical scaling relations observed on faults or on considerations of fault mechanics. There is now widespread recognition that an important scale is imposed on earthquakes and faulting in any particular region by the local thickness of the seismogenic upper crust or the length of the largest faults (e.g. Shimazaki, 1986; Cowie and Scholz, 1992b; Scholz, 1994). Faults that are small compared with this scale length often show features that are scale invariant over some part of their size range, but those simple rules are thought to change (and so, presumably, does the physics of the controlling processes) for larger faults. Many excellent field studies of scaling relations and the processes of

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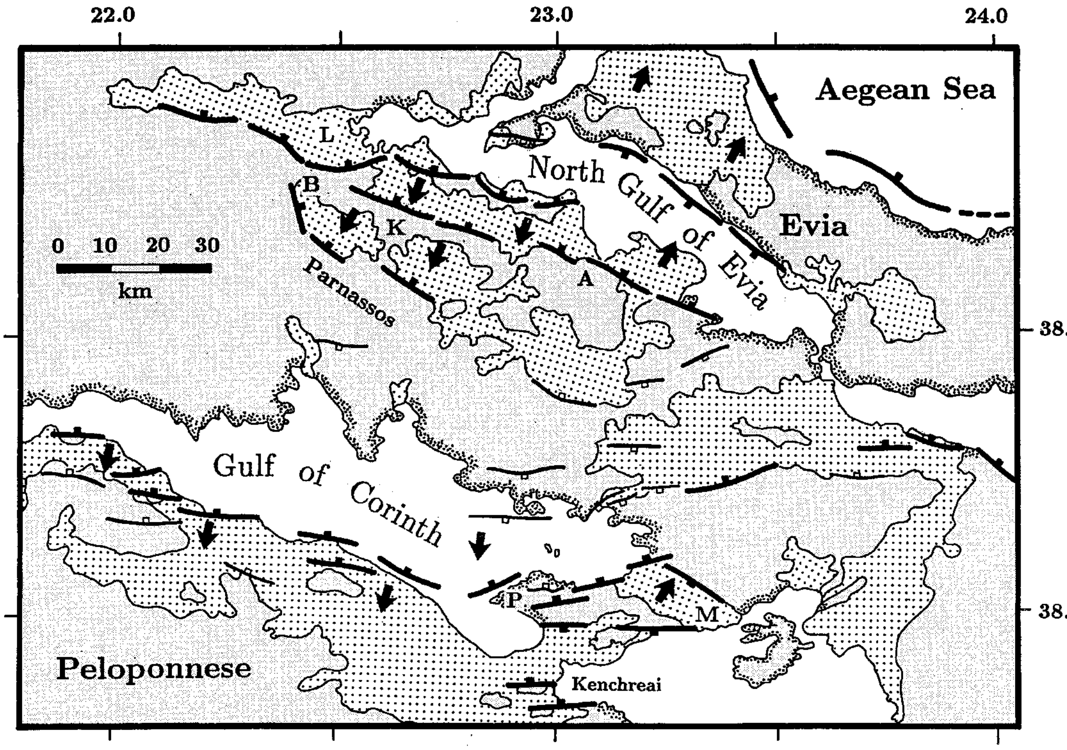

Fig. 1. Summary map of the major Plio-Quaternary normal faults in central Greece, together with their associated Neogene basins (stippled) and regional dip directions of sediments (arrows). Most sediment dips are gentle; usually less than 10–15°. Faults which bound range fronts with footwall ridges higher than about 500 m are shown as thick lines; faults with smaller topographic expression are shown as thin lines. Not all the faults on this map are equally active today. This map does not distinguish between the very active and less active (or completely dead) faults because this separation is conjectural in some cases, and so this map should not be used for seismic risk assessment. Nor is the map complete: knowledge of offshore faults is mostly sketchy, and there are likely to be more than are shown here. L marks the Lamia basin; M the Megara basin; B the Bralos basin; A the Atalanti fault; K is Mount Kallidromon; P is the Perakora peninsula.

fault propagation and linkage have concentrated on faults at the smaller end of the faulting spectrum (e.g. Trudgill and Cartwright, 1994; Dawers and Anders, 1995); but this is not the scale of faulting which has been illuminated by the advances in seismology, GPS geodesy and SAR interferometry, which can only image large faults. This note is concerned only with such large faults.

Perhaps surprisingly, it has been possible to recognize some of the processes of fault growth and interaction in operation on large, seismogenic faults in active regions by examining the evolution of their associated drainage patterns; in some cases even obtaining quantitative estimates of growth per earthquake to compare with theoretical models (e.g. Jackson and Leeder, 1994; Jackson et al., 1996; Mueller and Talling, 1997). From observations of range-front and fault-scarp morphology, Wallace (1978, 1984, 1987) suggested that fault activity occurs in episodes that switch from one range to another in the Basin and Range province of Nevada on a time scale of less than 1 Ma; a view supported by the drainage patterns around active normal faults (Jackson and Leeder, 1994) and by paleoseismology (Coppersmith, 1989).

Where is this increasingly sophisticated knowledge of the geometry and evolution of large faults taking us? The purpose of this note is to suggest that one challenge lies in understanding the connection between the instantaneous fault kinematics and the final geometry of the faulting at the end of the episode of deformation.

Within the wide areas of dispersed deformation that characterize the continents, it is now clear that the geometry and distribution of activity of large faults can change rapidly compared with the dating resolution in older terranes. Thus changing fault patterns are most accessible for study in active or recently active regions. Some faults continue to grow; others become permanently or temporarily inactive. The death or dormancy of large faults in an otherwise actively deforming province is sometimes revealed by the geomorphology of features in their footwalls and hanging walls, and in the interaction of those features with the geomorphology created by neighbouring faults. This note is to point out some examples of temporal and spatial changes in fault activity that are thought to have occurred over the last ~1 Ma within the active extensional province of central Greece. The most important structures are in and around the two major graben systems of the Gulf of Corinth and north Gulf of Evia. These structures are first described, then the activity between the two gulfs is compared. Finally, some possible causes and implications of these changing patterns are discussed.

2. Fault death, dormancy and episodicity in central Greece

The extensional province of central Greece (Fig. 1) is thought to have been active at present rates for about the past 5 Ma (e.g. Taymaz et al., 1991; Armijo et al., 1996). Major range-bounding normal faults that have been active in the Plio-Pleistocene, their associated sedimentary basins and the regional dips of bedding in those basins are shown in Fig. 1. The main morphological structures today are the big graben systems of the Gulf of Corinth and the north Gulf of Evia, both associated with numerous normal faults. However, not all the faults in Fig. 1 are equally active now.

2.1. The Gulf of Corinth

The seismicity and morphology of the Gulf of Corinth clearly indicate that the most active faults are those along the southern shore (in the centre and west) and the Perakora peninsula (in the east; P in Fig. 1). Some are also active offshore. South of the active coastal faults are other normal faults, which were responsible for Plio-Pleistocene basins in their hanging walls. In addition to the general lack of morphological evidence for late-Quaternary activity on the southern faults, their hanging walls contain Quaternary marine sediments that are now uplifted (Doutsos and Piper, 1990; Doutsos and Poulimenos, 1992; Armijo et al., 1996). Some of this uplift may be regional, related to the subduction zone to the south, but some is interpreted to be related to footwall uplift on the coastal faults, as its amplitude decays with distance from those faults over the expected length scale (e.g. Leeder and Jackson, 1993; Armijo et al., 1996). Whatever the origin of the uplift, the southern faults are not moving sufficiently fast to keep their hanging walls below sea level, whereas the coastal ones are. This contrast in the rate of activity between the faults has continued for at least the last 300 ka, based on the correlation of uplifted marine terraces with global sea level curves (Armijo et al., 1996) and on corals of that age which have been dated in the hanging wall of the Kenchreai Fault (Fig. 1) in the eastern Gulf (Collier, 1990; Dia et al., 1997).

At the eastern end of the Gulf, the E–W coastal faults cut the NW-trending normal fault system bounding the Megara basin (M in Fig. 1). The Plio-Pleistocene sediments of this basin dip northeast towards the bounding fault, yet the geomorphology is clearly dominated by the coastal fault, which has uplifted and tilted the basin towards the southeast. The change in activity from faults with NW–SE to E–W strikes happened within the last ∼1 Ma (Bentham et al., 1991; Leeder and Jackson, 1993).

Thus in the Gulf of Corinth the main activity seems to have switched from one set of large faults to another in the last ∼1 Ma. But is this change permanent? The southern faults in the west and the NW-trending Megara fault in the east show no signs of late Quaternary or Holocene activity. The Kenchreai fault is more enigmatic: though it has marine sediments 200–300 ka old in its hanging wall that are uplifted to a height of ∼100 m above sea level, it also has Holocene scarps, submerged coastlines and a submerged archeological site of ∼AD400 in its immediate hanging wall at its eastern end (Noller et al., 1997). Although this fault has evidently been less active than the faults on the Perakora peninsula to the north over the last 200–300 ka, it is not completely dead.

2.2. The north Gulf of Evia

Like the Gulf of Corinth, the north Gulf of Evia region also contains a set of large coastal faults that have Holocene scarps, and which are known to have moved in earthquakes (e.g. Roberts and Jackson, 1991). In the footwalls of these coastal faults, particularly at the western end of the Gulf, are other normal faults which are much less clearly expressed in the morphology and show only degraded escarpments. In the Bralos region (B in Fig. 1) the interaction between the faults bounding Mount Parnassos and the coastal faults of the Lamia basin (L in Fig. 1) is revealed in the geomorphology. Sediments in the Bralos basin dip towards the Parnassos fault system, but the whole region is uplifted and tilted by activity on the frontal Lamia basin faults (L in Fig. 1, Leeder and Jackson, 1993), in a geometry similar to that in the Megara basin. If the Parnassos faults are active, they are not active enough to dominate the geomorphology.

The contrast between the large faults bounding the northeast side of the Kallidromon range (K in Fig. 1) and those along the coast is also dramatic. Both have footwall ridges made of the same resistant Triassic–Jurassic platform carbonates and their topographic expression might be expected to degrade at the same rate, but the scarps along the coast are much clearer and better expressed in the morphology. The indications are that the Kallidromon faults are either inactive, or much less active than the coastal ones (Philip, 1974; Mercier, 1976). Other evidence suggests that the Kallidromon faults are also older than the coastal

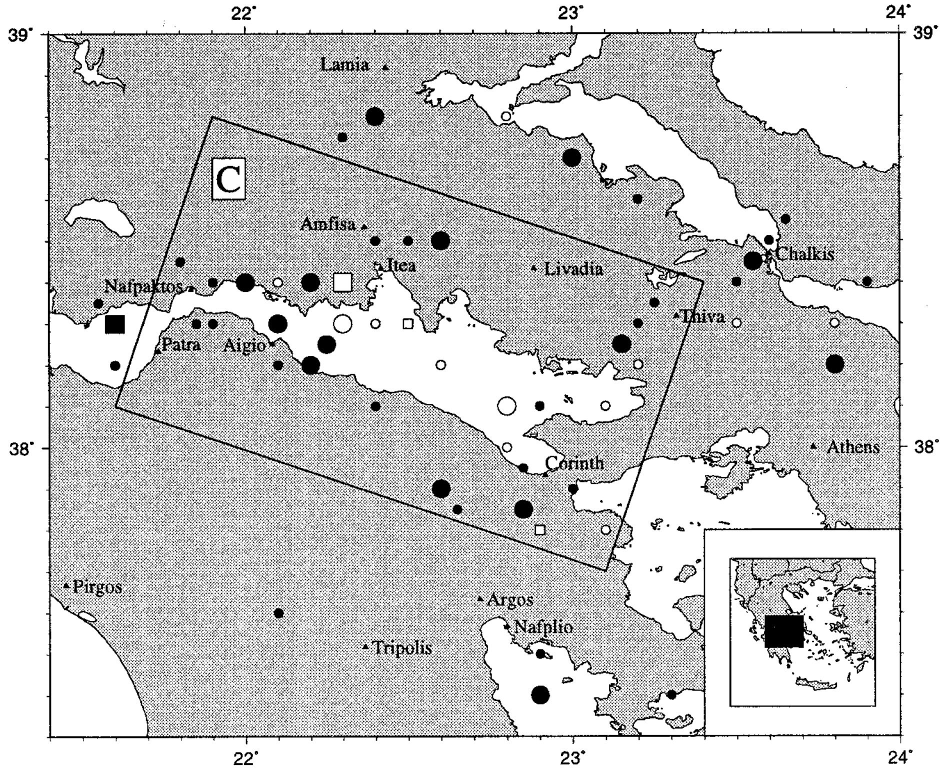

Fig. 2. Approximate epicentres of earthquakes in central Greece since 1694. Large symbols are events with Ms ≥ 6.5, smaller symbols are those with 6.0 ≤ Ms < 6.5. Filled symbols are for events prior to 1900, open ones are post 1900. Squares are for events that may have sub-crustal depths. The seismic moment release in the box marked C can account for an average of ~10 mm/y extension, which is about the same as that determined geodetically in the same region (Clarke et al., 1997b; Davies et al., 1997). Adapted from Ambraseys and Jackson (1997).

faults. Near the stratigraphic base of the tilted half-graben between the Kallidromon and coastal faults are conglomerates containing granite pebbles that have no conceivable source to the south or west in Greece. Their only likely source is to the north, and it is hard to see how they could have reached their present position if the coastal faults were active during their deposition. Thus although the Parnassos, Kallidromon and coastal faults and their associated sediments resemble a set of classic 'dominoes' or tilted blocks, the faults are not now equally active and may never have been active simultaneously.

2.3. Relations between the two major Gulfs

The north Gulf of Evia and the Gulf of Corinth are the most dramatic extensional structures of central Greece. Yet the Gulf of Corinth is overwhelmingly the most active of the two today. The total N–S extension across the two Gulfs is known to be ~12 mm/y from GPS and triangulation surveys (Billiris et al., 1991; Davies et al., 1997; Clarke et al., 1998), but the Gulf of Corinth alone accounts for ~10 mm/y of this total (Clarke et al., 1997b). The same contrast is seen in the historical seismicity, where the earthquakes of the last 300 years can account for ~10 mm/y extension in the Gulf of Corinth, but only ~1 mm/y in the north Gulf of Evia (Ambraseys and Jackson, 1997 and Fig. 2). There is no doubt that the north Gulf of Evia is of major importance in the late Quaternary extension of central Greece: the large coastal faults have footwalls 1000 m high and there is abundant evidence of Holocene activity on scarps at their base. In addition, the coastline shows evidence of Holocene uplift and subsidence that is controlled by the location of the faults (Roberts and Jackson, 1991; Stiros et al., 1992). Yet only one substantial earthquake has occurred in the north Gulf of Evia in the last 300 years (rupturing the Atalanti Fault, marked A in Fig. 1, in 1894). Though the north Gulf of Evia is not completely dead, the much smaller late Quaternary uplift in the footwalls of its coastal normal faults is evidence that it has been moving much more slowly than the Gulf of Corinth over the last 200–300 ka at least.

3. Discussion

3.1. Causes of fault inactivity

Central Greece shows abundant evidence that the position and rates of activity on large faults has changed over the last million years or so. The change involves some faults becoming less active. However, whether the inactivity is permanent, temporary or episodic is not always clear, and so the causes of it are largely conjectural.

An obvious possible cause of permanent fault death is a rotation of the fault with time, so that it becomes unfavourably oriented for movement. Rotations can occur about vertical and horizontal axes, and the lock-up mechanism in this case may be frictional (Nur et al., 1986) or even physical truncation by a new set of faults with a more favourable orientation (e.g. Proffett, 1977). The northwest strike of the apparently inactive faults bounding the Megara basin and Mount Parnassos relative to the more active E–W coastal faults suggests this as a possibility, since the faults in central Greece are thought to rotate clockwise with time about a vertical axis at rates as much as 5°/Ma (e.g. Kissel and Laj, 1988; Taymaz et al., 1991). Thus a discordance in strike of ~30° is quite possible over the Plio-Pleistocene period of deformation. Widespread tilting of footwall and hanging wall blocks generates stresses that may also contribute to frictional lock-up (e.g. Buck, 1988; Forsyth, 1992).

Another cause of fault inactivity could be the stress interactions between adjacent or nearby faults, such that slip on one fault can trigger or inhibit slip on another (King et al., 1994). This process almost certainly influences the seismic history of a group of relatively close sub-parallel normal faults (e.g. Hubert et al., 1996; Cowie, 1998), and is likely to be very important for small faults. However, whether this interaction can permanently deactivate large faults and whether it is important over distances several times the thickness of the seismogenic layer or the largest fault segment lengths (such as between the north Gulf of Evia and the Gulf of Corinth) is less obvious, and other alternatives are possible. For example, patterns of deformation over large distances may be more influenced by distributed flow in the ductile part of the lithosphere than by stresses transmitted through the seismogenic crust (e.g. England and Jackson, 1989; England and Molnar, 1997; Bourne et al., 1998).

Consideration of lithosphere rheology raises the possibility that the activity in the north Gulf of Evia has diminished, perhaps permanently, as the result of some 'strain hardening' effect during the extension. One process that may be capable of achieving this is the cooling and strengthening of the mantle lithosphere as it is brought closer to the surface, which depends on the interaction between strain and strain rate (Sonder and England, 1989; Newman and White, 1997). At first sight, the comparable heights of the topography surrounding the Gulf of Corinth and the north Gulf of Evia, their similar depths (900 m and 450 m, respectively), and the similar tilting of their syn-rift sediments (typically 10–15°), do not indicate dramatically different total extensions in each place, suggesting that this strain hardening mechanism is unlikely. However, the north Gulf of Evia is much closer to the Aegean Sea itself, where greater subsidence and crustal thinning do argue for larger finite extensions (Makris and Stobbe, 1984). If extension in the mantle is distributed over a wider area than in the upper crust (e.g. White and McKenzie, 1988), then mantle cooling beneath the north Gulf of Evia might be significant.

The Gulf of Corinth, the north Gulf of Evia, and the basin northeast of the island of Evia itself, all lie at the southwest end of a system of NE-trending strike-slip faults that cross the Aegean Sea to link with the North Anatolian Fault in Turkey (e.g. Taymaz et al., 1991). Armijo et al. (1996) suggest that the strike-slip faults have been propagating southwest through the Quaternary, and that the Gulf of Corinth is now the most active structure in central Greece because it is at the tip of this strike-slip system. This then represents another possible explanation for the relative quiescence of the north Gulf of Evia, though it should be pointed out that the tip of the strike-slip system is poorly defined, and may lie offshore to the northeast of Figs. 1 and 2, as there is no evidence for strike-slip faulting on land in either the observed faults or the earthquake focal mechanisms.

3.2. Implications

The changing patterns of large fault activity in central Greece are similar to those known to have occurred in the Basin and Range province of the western USA (Wallace, 1978, 1984, 1987), and those that are implied by the evolution of the drainage patterns in other regions, such as parts of New Zealand (Jackson et al., 1996). Changing temporal and spatial patterns of fault activity of the kind described here may be a general feature of regions that are extending or shortening, though the changes may occur too rapidly to be resolvable in older terrains (Nicol et al., 1997). Understanding such behaviour has obvious implications for earthquake hazard assessment as well as for the distribution and integrity of oil and gas reservoirs and perhaps even for migration pathways in hydrocarbon provinces. Perhaps more interesting are the less obvious and predictable tectonic and structural implications of understanding the changing patterns of large fault activity. At one level, it may help us understand why earthquakes have a worrying habit of occurring on faults that were unknown or unappreciated before: usually because their geomorphological expression was more subtle than other faults nearby, such as at San Fernando (California, USA) in 1971, Landers (California, USA) in 1992, Northridge (California, USA) in 1994, or Kobe (Japan) in 1995.

At another, it may make us re-examine some simplistic structural models more closely. For example, the simple domino model of extension, in which fault blocks rotate simultaneously about horizontal axes, can often provide a reasonable estimate of the average extension (White, 1990) and even the associated vertical motions (Yielding, 1990) in a basin. Yet the classic domino arrangement of the faults bounding Parnassos, Kallidromon and the coast of the north Gulf of Evia involves faults that are probably neither active simultaneously nor formed at the same time.

At the largest scale, changing fault patterns may tell us about the relationship between discontinuous deformation in the seismogenic crust and flow in the mantle lithosphere beneath. In Greece, we have a good idea of both the overall velocity field today (Jackson et al., 1992; Davies et al., 1997; Clarke et al., 1998) and of how slip on faults achieves it (Taymaz et al., 1991). The relationship between the two is quite close: only certain orientations of faulting can accommodate the velocity field if the faulting is to be in simple patterns (Holt and Haines, 1993). Yet we also know the faults are rotating clockwise about a vertical axis. If the faults change orientation they cannot accommodate the present-day velocity field. Either the faults, or the velocity field, must change. Which change actually happens may depend on whether the strength of the seismogenic upper crust, or the ductile lower lithosphere, controls the deformation of the continents (Jackson et al., 1992; Bourne et al., 1998).

Observing fault patterns that change on a time scale of ~1 Ma is not easy, and attaching firm dates to those changes is even more difficult. Nonetheless, it is easier to see and describe these temporal and spatial changes than to be sure of what causes them; a problem in the relationship between kinematics and dynamics that occurs throughout the Earth sciences. We certainly do not yet understand what causes the changing fault activity outlined here in central Greece, and there may well be more than one mechanism operating. Insights into the dynamics of these changes are likely to come when we have more, and better-described, examples that allow us to compare different regions. Perhaps the best way to sum up the significance of understanding changing fault patterns is as follows. As earthquake seismology, GPS geodesy and SAR interferometry give us an ever more detailed instantaneous picture of tectonic motions, and as three-dimensional seismic reflection surveys provide us with ever more accurate images of the finite cumulative deformation, we might finally be able to connect the two, and answer the question 'how did it get like that?'

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