Front, left to right: Mark Handy, Tuncay Taymaz, Jean-Pierre Brun, Walter Mooney. Back, left to right: Patience Cowie, Christian Teyssier, Alain Vauchez, Kevin Furlong, Gregory Beroza, Brian Wernicke

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Group Report: Nucleation and Growth of Fault Systems

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INTRODUCTION

Faults are key components of the Earth system. They are sites of great earthquakes, which have an obvious, direct impact on the biosphere, including human societies. Faults have repeatedly affected the evolution of the Earth's lithosphere and surface. Mountain belts, rift valleys, mid-ocean ridges, and other features at plate boundaries are dominated by faults. In addition, faulting triggers surface motion on a broad range of scales (Figure 4.1). Thus, it is critical that we understand how faults form.

The focus of our discussion group was on finding ways to improve our understanding of how the structure of major fault systems evolves in space and time. Key questions relevant to understanding the characteristics of fault systems include:

1. What is the behavior of fault systems on different timescales?
2. What is the pattern(s) of fault system localization?
3. What drives localization?
4. Do different types of faults (strike-slip, thrust, normal) interact with the lithosphere in fundamentally different ways?
5. What is the style of faulting at different lithospheric levels, and how do transitions occur?

Identifying these basic questions is straightforward, but answering them is problematic. There have been substantial advances in our understanding of fault

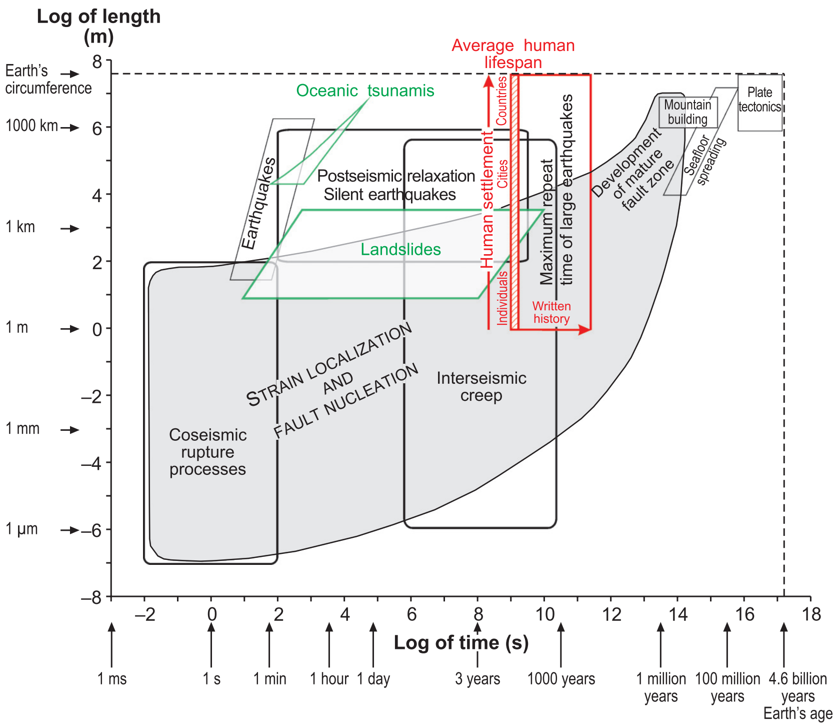

Figure 4.1 Length (m) versus time (s) on a log-log plot showing scales of fault processes (black, gray) and surface processes triggered by faulting (green) compared to the human scale (red). Diagram includes information from the following sources: Coseismic, postseismic and interseismic processes (modified from Figure 7.2), mountain building (Pfiffner and Ramsay 1982), landslides from rock falls to earth flow (www.landslides.usgs, www.planat.ch) and tsunamis (wwww.walrus.wr.usgs.gov/tsunami). Gray area in background shows the large range of conditions for faulting, from nucleation and localization to maturation on the scale of plate boundaries. Diagram contributed by M. R. Handy.

systems within the lithosphere, but we are faced with some fundamental limitations. For one, we cannot directly observe deformation along faults except near the Earth's surface. Despite the fact that faulting has direct impacts on human civilization (see Chapter 10), most processes related to faulting occur on time- and/or length scales other than the human scale, as shown in Figure 4.1. For example, faults nucleate on very short timescales, but reach structural maturity only after millions of years. The repeat time of large earthquakes on some seismogenic faults exceeds the human life span. We therefore rely on a combination of tools to study faults—from geophysical imaging of the fingerprints of active processes, to integrated field studies of exhumed, inactive (fossil) fault zones—and to see the deformation itself, even after faulting has ended. Unfortunately, many key observations are still missing; techniques to improve the resolving power of our deep structural images need improvement, as does our understanding of the physical processes that drive faulting.

The questions raised above are interlinked. We cannot isolate temporal from spatial behavior; deformation in fault zones at shallow levels can have profound impacts on the style and amount of deformation at depth, and vice versa. Fault systems develop under widely varied kinematic and stress regimes, producing fundamentally different styles of deformation in convergent, divergent, and transform fault systems. However, there are apparent similarities among these fault systems. Possibly, we can exploit these similarities to discover the fundamental processes underlying the development and evolution of fault systems. In so doing, we must recall that these processes may only be apparently similar; we may be misled to assume that all faults have common factors controlling their development.

The suite of questions above falls under three major themes, which directed our deliberations. The timescales of fault development (question 1) was our first theme. Localization of fault systems, both from an observational and process-oriented perspective (questions 2 and 3), provided our second. Our third theme focused on how fault systems interact within the lithosphere, both in terms of how fault types (strike-slip, thrust, normal faults) interact with compositional, thermal, and rheologic layering of the lithosphere (question 4) and how fault behavior varies with depth, not just within the lithosphere but also at the lithosphere–asthenosphere boundary (question 5).

In addressing these themes, we reviewed the current state of knowledge, identified critical open questions, and attempted to formulate strategies to address these questions. In this report, we focus primarily on the latter two components of our discussion. Much of the current state of knowledge is provided in the collection of background papers in this volume. However, to put our discussions of open questions and strategies into context, we interject brief discussions of some background material and provide short descriptions of key examples and processes.

TIMESCALES OF FAULT DEVELOPMENT

Faulting occurs over a wide range of timescales (Figure 4.1): Earthquakes accomplish significant displacements in seconds to at most a few minutes. At the other end of the time spectrum, shear zones deforming by aseismic (i.e., ductile or viscous) creep mechanisms develop over thousands to millions of years. Faulting on the scale of orogens, continental rifts and seafloor-spreading systems lasts even longer. Although it is clear that the earthquakes are episodic by nature, determining the degree to which other fault-related processes are episodic is an important research frontier in the study of active and ancient natural systems, as well as in laboratory and numerical experiments on fault system behavior.

Two new developments provide insight into this issue and summon a re-evaluation of our concepts of the timescales of fault zone behavior: the deployment of continuous GPS (Global Position System) networks and the application of cosmogenic nuclide dating to fault-related surfaces.

Monitoring of active faults with continuous and semi-continuous GPS instruments shows a variety of transient deformational events. The recognition of creep events on large areas of several major subduction zone interfaces shows that, for at least some parts of some subduction zone systems, much interplate deformation occurs neither continuously for years nor suddenly during earthquakes, but rather over days to months. Sometimes these events appear to be quasi-periodic, strain release that makes up a large proportion of the total strain release along the plate boundary. Although the location and geometry of these so-called silent or slow earthquakes are becoming understood (Dragert et al. 2001), the details of what goes on at depth are still unresolved.

Other evidence for transient deformation from continuous GPS networks includes intraplate extension in the northern Basin and Range Province (western U.S.A.) on a characteristic timescale of years to decades (Figure 4.2). Although the tectonic setting of this transience is well understood, there is still no satisfactory explanation for the spatial extent or timing of such events.

We are beginning to recognize transient behavior on longer timescales associated with the interaction of individual faults within fault systems. For example, the San Andreas and San Jacinto faults in southern California together accommodate a significant portion of the oblique convergence between the North American and Pacific Plates. However, the partitioning of this plate motion on the two faults has not been constant through time, but has alternated between them on a hundred to thousand to million year timescale (Bennett et al. 2004).

The use of cosmogenic radionucleides to date erosional surfaces near faults together with paleoseismic studies indicates that individual fault segments can have complex histories. Clustering of earthquakes on timescales of a thousand years and longer is an increasingly common observation in many fault zones. How this clustering affects the loci of deformation through time on large fault systems, such as in the Basin and Range Province, is an important, unanswered question.

Although it is evident that some fault zones show transient behavior on timescales ranging from seconds to perhaps millions of years, we do not fully understand the mechanics that govern such transients. Are there previously unrecognized mechanisms of lithospheric deformation?

If we can distinguish periods of fault movement with different strain rates, we may find that there are transients on longer timescales. For example, is it possible that periods of contrasting strain rates reflect changes in the relative activity of competing deformation mechanisms at depth? An example of this is

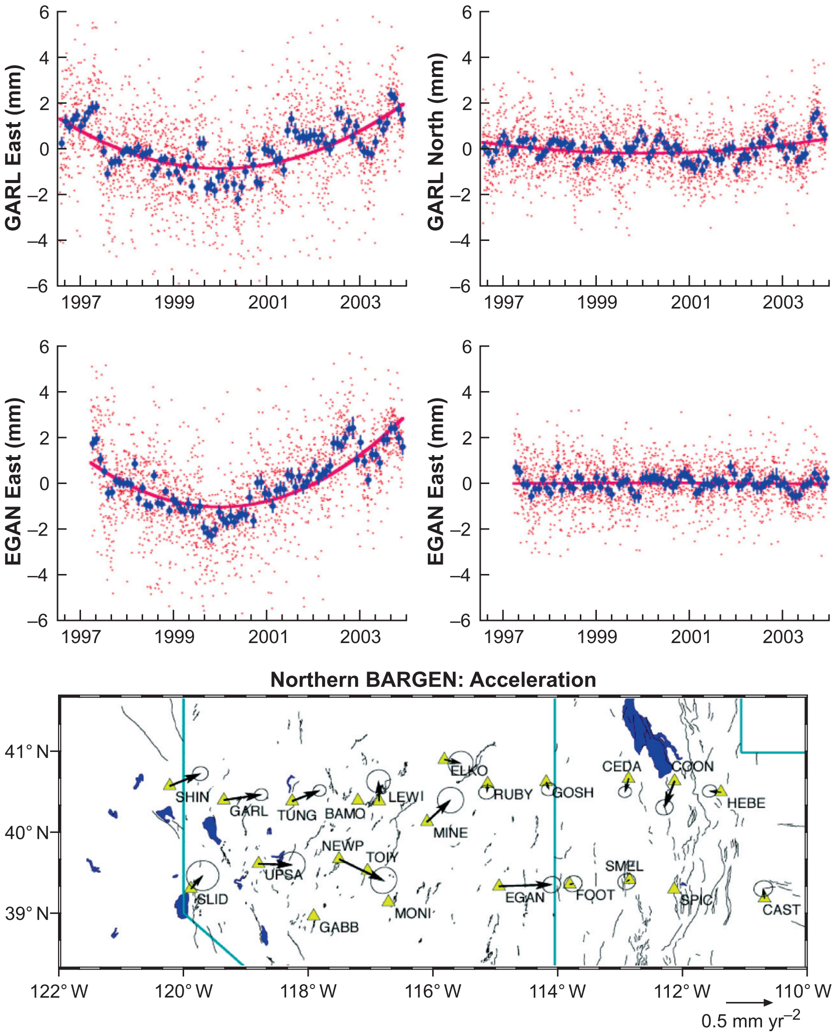

Figure 4.2 Top: Time series for continuous GPS sites GARL and EGAN in the intraplate rift setting of the northern Basin and Range Province from 1997–2004, with linear trend subtracted to show deviations from steady state. Accelerations of the sites from regression of second-order polynomial are 0.2 and 0.4 mm yr^{-1}. Bottom: Vectoral accelerations of continuous GPS sites showing regionally coherent acceleration of western sites relative to eastern sites of ~0.2 mm yr^{-2}. Accelerations are not correlated with earthquakes or patterns of regional seismicity. Source: J. L. Davis and B. Wernicke, unpublished data.

the possibility that the development of S–C fabrics in mylonite triggers accelerated fault motion at depth, loading the fault system in the upper crust and potentially leading to large earthquakes (see Chapter 6).

There are some basic strategies that could prove useful in addressing the problem of deformational transients. Clearly, we need to improve the spatial resolution of our observational methods. Increasing the density of observational networks (e.g., GPS networks) can provide both higher resolution locations and much lower detection levels in seismic magnitude. Combining seismic and geodetic observations has proven to be a powerful tool in providing observational continuity across broadly differing timescales of deformation. Extending monitoring beyond the plate-boundary zones to adjacent areas may provide different insight into the temporal scales of deformation. We must move beyond seismology and geodesy to incorporate detailed geologic studies, providing us with information on even longer timescales. Such studies may help us understand whether transient behavior results from time-dependent rheology or from changes in the kinematic boundary conditions.

As mentioned above, deformational fabrics may provide constraints on strain rate, not only strain. For example, microstructures in dynamically recrystallized quartz aggregates have been calibrated with laboratory experiments to yield estimates of temperature and strain rates during dislocation creep in natural shear zones (Stipp et al. 2002). Similarly, the calculation of heat production and dissipation rates in pseudotachylites has placed limits on paleostress and slip rate during earthquakes (e.g., Sibson 1975; Otsuki et al. 2003). Calibrations of yet other deformation mechanism(s) using laboratory data and theoretical considerations may yield constraints on stress and strain rates over a broader range of pressures, temperatures, and strain rates in nature. In particular, combining laboratory experiments with field studies of natural fault systems near the brittle-to-viscous transition may give us information on rate dependence of combined frictional and viscous deformation.

Field areas with a well-constrained thermo-barometric history are best suited as natural laboratories for this kind of work. Advances in geological dating techniques are allowing us to place closer constraints on the episodicity of slip (cf. Chapter 10). Also required is a better understanding of the deformation mechanisms during transient behavior. Do ductile fault rocks deform episodically—we know that brittle deformation can be episodic; is ductile deformation also episodic? If so, are there microstructures that can diagnose episodicity? This calls for a closer experimental investigation of transients—how they develop and what signature they leave in the rock record (see Chapters 5 and 7).

We should continue to develop tools for sampling Earth's response to sudden loading and unloading events, such as large earthquakes. These may provide important information about the time dependence of fault processes in the crust and upper mantle that are not accessible to direct observation.

Related to the timescales of localized faulting is a broader issue of the how large-scale deformation of the crust and mantle behaves with time. For example, large areas of distributed deformation in the Aegean Sea, the Basin and Range Province (western U.S.A.), and Tibet reflect homogeneous flow of the lithosphere. The mechanism by which such flow initiated or accelerated is largely unknown. If we knew this, we would have additional quantitative constraints on the interaction of faults and shear zones with the broadly deforming area.

LOCALIZATION OF FAULTS THROUGHOUT THE LITHOSPHERE

We have very few constraints or examples of the actual structural evolution of fault zones from inception to maturity. How do fault systems localize? Are spatial or temporal patterns of localization characteristic of the processes involved in their nucleation and growth, or do these patterns also reflect the fault type and tectonic setting?

The most complete set of observations of fault evolution and localization in the brittle upper crust come from extensional fault systems (Chapter 3). However, a basic open question is how to determine the pattern of fault localization for strike-slip and thrust fault systems. Progress has been made by studying large, deeply eroded Pre-Cambrian and Early Paleozoic oblique-slip fault systems in shield areas (e.g., southern Madagascar; Martelat et al. 2000). In many cases, however, the late stages of faulting tend to overprint the record of nucleation and early growth, making it difficult to discern the mechanisms active at the onset of faulting.

For example, although we have very good constraints on the rate at which the San Andreas fault system has lengthened during northwestward migration of the Mendocino triple junction, we have only a rudimentary understanding of how the primary fault strands formed (e.g., Burgmann and Freed 2004). One strategy that can be applied along actively propagating fault systems is to exploit the idea that their lateral propagation in map view preserves progressively older stages of their evolution along their strike. In a study of the Red Sea Rift system, Favre and Stampfli (1992) showed that progressively younger stages of rifting are exposed in the direction of present-day rift propagation. This space-for-time substitution has also been applied to the San Andreas fault system, whose age decreases toward the Mendocino triple junction. Therefore, we can study the early stages of fault system development near the triple junction and the more mature stages further to the south (e.g., in the vicinity of San Francisco). This approach mitigates somewhat the problem that arises when faulting erases traces of the early deformational phases.

A second strategy is to study finite strain gradients within fossil fault zones that have undergone differential exhumation and therefore expose different paleodepths of the fault at the surface. This allows the study of fault mechanisms and their propagation both along their strike and down dip. A good example of this is the Periadriatic fault system, an array of late orogenic transpressional faults in the Tertiary European Alps (Figure 4.3). The structures

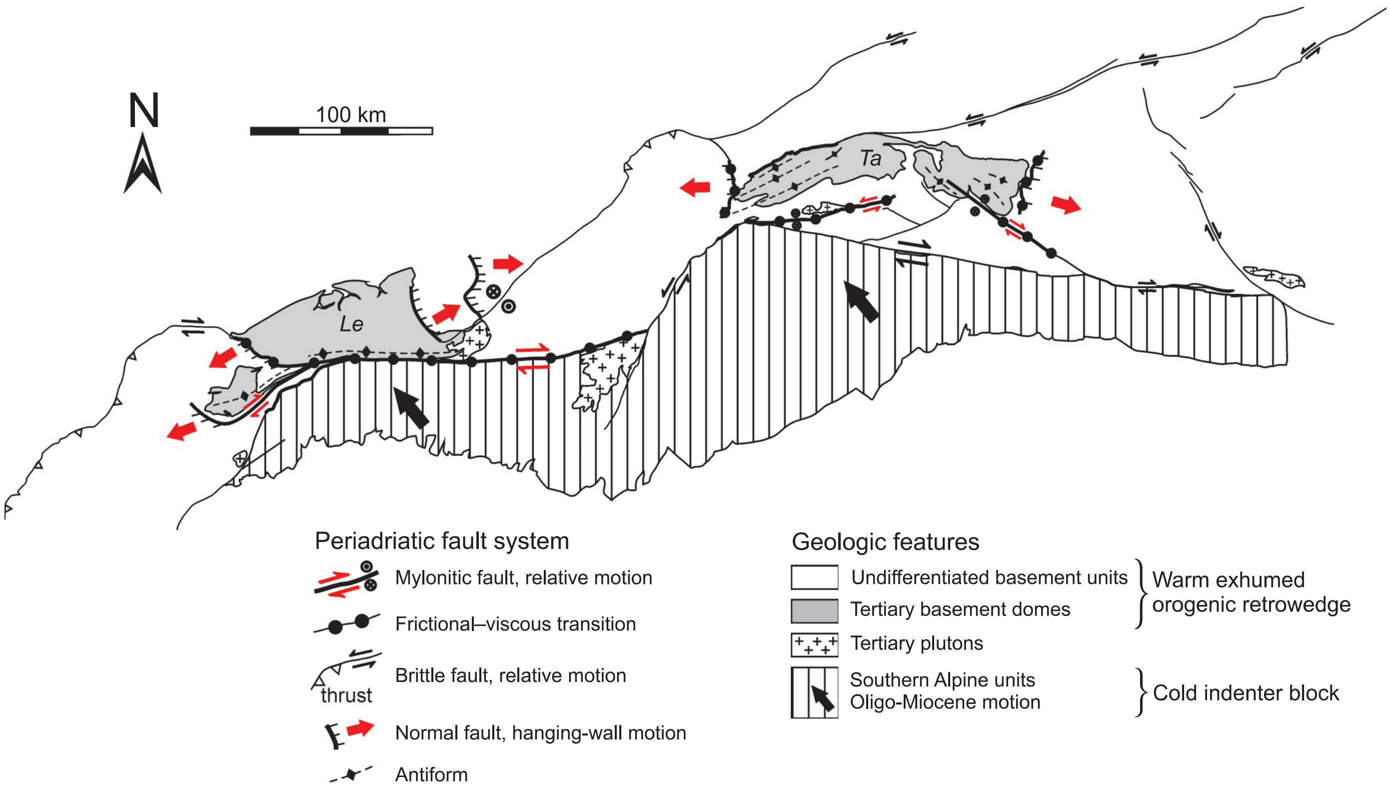

Figure 4.3 Map view of the Periadriatic fault system (PFS) modified from Handy et al. (2005). Le: Lepontine thermal dome; Ta: Tauern thermal dome.

preserved in this fossil fault zone indicate that localization in the upper, brittle crust involved fracturing and cataclasis, whereas in the lower crust, it involved buckling and viscous mylonitic shearing of existing mechanical anisotropies like compositional banding, schistosities, and even older mylonitic fault systems (see Box 4.1). Below the brittle-to-viscous transition, the transpressional shear zone terminates at its western end in a series of large constrictional folds and splayed, oblique extensional shear zones that deform the boundaries of previously formed basement nappes. This example underscores the important role of existing anisotropies on fault propagation and strain partitioning on the crustal scale.

Substituting space for time in the analysis of large fault systems to determine their evolution is not without problems. First, in the case of fossil fault systems, the interpretation of a finite strain gradient in terms of progressive localization is only valid if the strain gradient coincides with an isotopic record of progressive crystallization or cooling of synkinematic minerals (e.g., micas). The temporal resolution of the isotopic systems may not be sufficient to capture rapid localization events. Moreover, movement on fault systems with no, or only minor, dip-slip component is not recorded by differential cooling and closure of isotopic systems on either side of the fault. New *in situ* dating methods, preferably using highly retentive isotopic systems, are necessary to determine synkinematic formational ages over the complete spectrum of metamorphic temperatures. Second, the problem of selective microstructural memory is more acute in some tectonometmorphic settings than in others. Whereas fossil

Box 4.1 The Periadriatic fault system (PFS), European Alps.

The PFS delimits the retro-wedge of the Tertiary European Alps along a length of several hundred kilometers. During oblique convergence of the European and Adriatic Plates during the Oligo-Miocene, this fault system accommodated about 100–150 km of dextral strike-slip and several tens of km of shortening in a direction normal to the fault trace. In the vicinity of the Lepontine and Tauern thermal domes, it also accommodated up to 25 km of N-block-up exhumation. Due to differential exhumation and erosion, the PFS currently exposes fault segments that were active from near the surface to about 25 km depth. The fossil frictional-to-viscous transition during the Oligo-Miocene is marked with circles in Figure 4.3. Below this fossil frictional-viscous transition, strain localization involved buckling and shearing of preexisting anisotropies, assisted by the ingress of metamorphic fluids and melt into the mylonitic rock. Buckling was transitional, in both time and space, to passive folding, viscous mylonitic shearing, and strain localization in the fold limbs. The strain ultimately localized in a network of oblique, high-angle thrust faults and low-angle normal faults within the warm, exhuming orogenic block (see Figure 6.14). The PFS may be a good analogue for segments of large strike-slip faults at depth, like the San Andreas or North Anatolian fault systems.

extensional fault systems preserve the best record of fault evolution, thrust systems lose most structural and isotopic traces of fault nucleation and early growth due to thermal and deformational overprinting during prograde metamorphism. Third, structures in fossil fault zones are time-integrated products of the deformation history. Deformation at any given time during the activity of the shear zone may have been accommodated within only a part of the entire exposed shear zone before jumping to other parts of the same shear zone at a later time. Fourth, the structure of a fault reflects not only the finite strain, but also changes in kinematic boundary conditions. For example, the San Andreas fault system developed as the result of a change from subduction to lateral translation between the Pacific and North American Plates. Likewise, the Periadriatic fault system probably originated as a rift-related transform fault prior to becoming intra-orogenic strike-slip fault during the oblique convergence of Europe and Adria (Schmid et al. 1989). The transtensional setting at the western end of North Anatolian fault system in the northern Aegean Sea (Figure 4.4) is quite different from its transpressional environment in Anatolia during most of its formation and propagation. Determining whether there was a fundamental change in fault development across and along its trace is difficult in areas with a complex overprinting history (see Box 4.2).

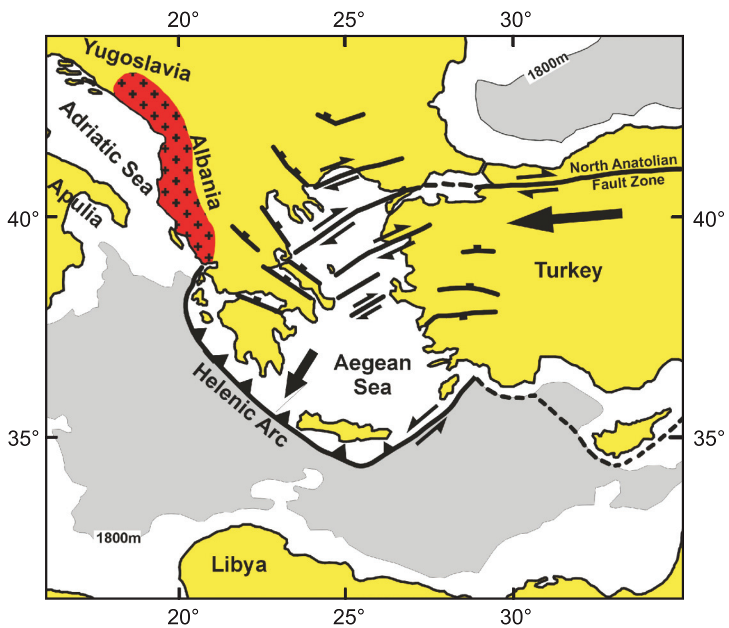

Figure 4.4 Tectonic overview of the Aegean–Anatolian region showing the relationship of Hellenic trench-arc system, Aegean extension, and motion of the North Anatolian fault (Taymaz et al. 1991).

Box 4.2 The North Anatolian fault zone (NAFZ).

NAFZ is an approximately 1500 km long, broadly arcuate, dextral strike-slip fault system that extends from Greece to eastern Turkey. It is predominantly a single zone of a few hundred meters to 40 km width. Along much of its length, this fault zone consists of a few shorter, subparallel fault strands that sometimes anastomose. The age and cause of dextral motion along NAFZ is controversial, and there are basically four different views:

1. The right-lateral motion commenced by the middle Miocene and resulted from westward lateral extrusion of Anatolia away from the collisional zone between the Arabian indenter and Eurasian Plate.
2. The NAFZ did not initiate until the latest Miocene or Early Pliocene.
3. The NAFZ initiated in eastern Anatolia during the Late Miocene and propagated westwards, reaching the Sea of Marmara area during the Pliocene.
4. The NAFZ initiated at ~16 Ma or more in the east, but at less than 3 Ma in the west.

Geologic mapping of offset markers along the NAFZ yields total displacements ranging from a maximum of 85 ±5 km to as little as 20–25 km. Geological data supports the view that the NAFZ slips at rates ranging from 5–10 to 17 ±2 mm yr^{-1}, whereas plate motions and seismological data suggest rates of 30–40 mm yr^{-1}. This discrepancy arises from the exaggerated slip rate obtained by treating the intense seismicity on the NAFZ during 1939–1967 as typical of all time. Recent GPS data indicate present-day rates of about 15–25 mm yr^{-1}. Extrapolating these rates back to the Early Pliocene yields a total displacement of 75–125 km, which is in close agreement with the maximum 85 ±5 km estimate based on offset markers.

Over the past 60 years, the earthquakes along different segments of the NAFZ are atypical of long faults. Beginning with the 1939 Erzincan earthquake (M = 7.9 to 8.0), which produced about 350 km of ground rupture, the NAFZ ruptured eleven times during moderate to large earthquakes (M > 6.7), forming a surface rupture trace more than 1000 km long. Most of the earthquakes occurred sequentially, from east to west.

Therefore, space-for-time substitution only works in field laboratories where kinematics have not varied significantly during the time span of fault evolution. To distinguish whether all faults follow similar movement paths during their evolution, we need to obtain similar data sets from various fault types and settings. In addition, we need to analyze fault systems of different sizes, because it is not clear whether fault systems behave and evolve in a self-similar fashion. For example, should we expect that lithosphere-scale extensional fault systems, such as the Basin and Range Province, have similar deformation histories to the much smaller extensional faults in regions such as the volcanic tablelands of eastern California?

Implicit in many models of fault zone evolution and localization is the assumption that similar mechanisms drive localization, irrespective of scale. Is localization at shallow levels within the brittle (seismogenic) crust simply a response to localized shearing at depth? Or does localization in the brittle crust drive localization in the deeper ductile shear zones of the lower crust and lithospheric mantle? Numerical modeling has shown that both processes are viable. However, significant feedbacks are likely between the shallow, dominantly brittle faults and the deeper, dominantly ductile shear zones of the lower crust and lithospheric mantle. One system may lead or lag behind the other on short a timescale; however, on longer timescales they likely develop together as strain compatibility is maintained.

The Blackwater fault in the Mojave Desert of southeastern California is part of a broad fault network (Chapter 3). The displacement along individual faults in the array varies, even though these faults are crustal-scale structures some tens of kilometers long. Geodetic data show that strain has accumulated across the entire Mojave Desert fault array, but the data lack the spatial and temporal resolution to show how individual faults are being loaded. They could be loaded by localized shear at depth, although this has not been documented for most faults in the array (cf. discussion of the Blackwater fault in Chapter 3). Alternatively they could be loaded horizontally by stress transfer between adjacent faults. Yet another possibility is that the loading of these faults is strongly influenced by active deformation to the north in Owen's Valley, such that deformation is propagating from north to south (G. King, pers. comm.). Recent earthquake activity in this area occupies a narrow zone within the array (e.g., Landers and Hector Mine events). The small spacing of these events has led some authors to suggest that the fault array reflects a stage of incipient localization. Thus far, however, there is no clear evidence to suggest that the recent seismic activity is due to localized strain accumulation at depth. How can we determine the degree of strain localization reached within this fault array? If it is in a state of incipient localization, then what observations or measurements should be made to constrain the driving mechanism of localization?

The mechanism of fault-loading obviously has a direct bearing on seismicity. For example, the theory of rate- and state-dependent friction indicates that the loading rate is as important as the load magnitude for producing failure (Tse and Rice 1986). Still, many seismologists argue that strike-slip faults are loaded from below by aseismic creep within a deep-seated shear zone (see Chapter 7). Indeed, Handy et al. (Chapter 6) have argued that accelerated creep associated with shearing instabilities (shear bands) within ductile shear zones potentially trigger earthquakes in the brittle, upper crust. The loading history of any given fault is expected to vary due to its interaction with nearby faults (Stein 1999) as well as to time-dependent gradients in the pore fluid pressure (e.g., Miller 2002).

We should not expect that the processes driving localization are the same on all length scales. Considered on the lithospheric scale, fault development is driven largely by mechanical instabilities on the order of kilometers to hundreds of kilometers in length. However, as one moves down to the scale of grains and grain aggregates, the determinants of strain localization are metamorphic fluids, cracks opened by local decreases in the effective pressure, or preexisting structures like grain boundaries. Localized domains on the granular scale can then coalesce to form large-scale fault zones ("upscaling" shown in Figure 6.10b), such that the change in length scale is associated with a change in the factors and mechanism of localization.

Speculation on how these various factors may influence localization in different depths is as follows: If the lower crust and/or upper mantle are nominally dry and/or melt-free, then strain localization below the seismogenic zone primarily reflects the effects of the temperature field and the spatial distribution of lithologies on rheology. The length scale of this localization will be long compared to that of hydrous and/or partially melted crust, where fluids enhance weakening due to embrittlement and the formation of intrinsically weak, hydrous minerals (e.g., micas) (Chapter 11). Brittle faults will propagate ephemerally (coseismically) downward from the seismogenic zone (Chapter 6), thereby introducing fluids to the top of the viscous crust and shortening the length scale of localization there. The localization pattern may then be controlled by stresses and fluid feedback systems in the upper, brittle crust; the rate control on localization is therefore "from above."

If the lower crust and/or upper mantle are hydrous or contain a melt (either intruded from below or derived by partial melting), then this can induce hydrofracturing (e.g., Davidson et al. 1994) and will likely shorten the length scale of localization in the viscous lithosphere with respect to the dry condition (above). Episodic hydrofracturing in the presence of a metamorphic fluid or melt is very fast, occurring at rates of meters to kilometers per second (Handy et al. 2001, and references therein). Localization will therefore propagate upward, probably sporadically, at the rate of deformationally induced, upward fluid advection. The frequency of these upwardly penetrating localization events is controlled by the overall fluid production rate at depth; the rate control on localization is therefore "from below."

In both of the above scenarios, surface processes (erosion, transport, sedimentation) only affect faulting on short length scales to superficial depths (say, 1–2 km: Chapter 8), i.e., to levels at which topographic stresses dominate the composite stress field (tectonic stress + lithostatic stress + topographic stress). The faulting geometries at these shallow depths are important inasmuch as they affect the mode of faulting and hence, the effective surface area exposed to erosion. They will govern erosion rates (Chapter 9). On longer time- and length scales, high erosion rates can influence the dynamic stability of the lithosphere and thus enhance exhumation rates.

In this conceptual model, there are three main controlling factors on fault zone evolution: (a) the rate of fluid production and location of fluid reservoirs; (b) the kinematics of faulting (thrusting, extension, or strike-slip); and (c) the length scales and geometry of preexisting anisotropies (faults, lithologies, foliations) with respect to size of the elastic stress field.

The question whether strain localization progresses top-down or bottom-up probably does not have a single, simple answer. Different fault systems behave differently and change through time as feedbacks modify the system. The style of localization also potentially affects the behavior of brittle faults. One can imagine that the manner in which a fault is loaded—from below where a shear zone has developed beneath the brittle fault, or from the side if the fault is significantly more localized than its substratum—will affect its seismogenic character.

To distinguish the relative roles of these factors during localization, it would be useful to compare continental faults with faults in oceanic environments (oceanic transforms, fracture zones), where localization evidently occurs even in the absence of fluids. The same is true of continental strike-slip fault systems, where rocks are neither significantly buried nor exhumed and therefore do not dehydrate or hydrate. In contrast, continental thrust faults and normal faults experience a continuous flux of fluids as rocks are subjected to changing *P–T* conditions during burial and exhumation. It is likely that strain localization is achieved in different ways in hydrous and anhydrous environments.

INTERACTION OF FAULT ZONES WITH LITHOSPHERIC AND ASTHENOSPHERIC LAYERING

Do different types of faults (strike-slip, thrust, normal) interact with thermal and lithological layering in the lithosphere in fundamentally different ways? A common assumption in geodynamic modeling is that lithological layering is horizontal. Likewise, isotherms are considered to be subhorizontal in a steady-state thermal regime. Together, these thermal and lithological boundaries form a predominantly subhorizontal mechanical layering, commonly termed rheological stratification. Faults with a dip-slip component of motion obviously interact with such layering differently than subvertical strike-slip faults. Thrust and normal faults reflect the rheologic stratification, at least in the upper crust, by adopting a ramp-flat geometry. Still, we do not fully understand the mechanics of the interaction between fault geometry, kinematics, and mechanical layering, particularly at depth, within the ductile regime.

During thrusting and extension, changes in crustal thickness are accompanied by temperature changes, leading to changes in the rheology of the lithosphere. The thermo-mechanical feedbacks that characterize thrusting and normal faulting at the lithospheric scale are far less important in strike-slip fault zones, where fault displacement is parallel to horizontal isotherms and overall lithological layering.

Of course, lithological contacts are not always horizontal and, especially in orogens, they have been multiply folded during earlier deformation. Likewise, deformation and exhumation rates are usually high compared to the thermal equilibration rate of the orogenic crust, as manifest by clockwise *P–T* loops derived from petrological studies of metamorphic mineral parageneses. In such cases, faults are expected to evolve in thickness, length, and style.

Temperature in the lower crust and lithospheric mantle almost certainly has a major influence on the width and orientation of highly deformed zones in the lithosphere. Fossil prograde amphibolite- to granulite-facies shear zones in deeply eroded Precambrian crust show that localization occurs at the scale of tens to hundreds of kilometers (Hoffman 1987; Martelat et al. 2000; Vauchez and Tommasi 2003). At Moho temperatures less than 800°C, the strength of the upper mantle exceeds that of the middle and/or lower crust. The ductile lower crust becomes a large-scale décollement horizon between the upper crust and the upper mantle (Figure 4.5). Analogue and numerical modeling have shown that the amount of coupling between the upper brittle crust and the ductile décollement layer strongly controls the dynamics of upper crust faulting, with considerable variations in the duration, vergence, sequence, and spacing of the primary thrust and normal faults. Deformation below the Moho remains poorly understood, in part due to the lack of high-resolution geophysical imaging and the paucity of large tracts (= 100 km^2) of well-exposed, exhumed mantle rocks at Earth's surface.

Petrological data and numerical modeling indicate that the formation of metamorphic core complexes in large extensional domains, such as the Basin and Range Province or the Aegean Sea, occurs at Moho temperatures of 1000°C or more. There, the lithospheric mantle was probably replaced by the asthenosphere. However, the modes of faulting and ductile shearing leading to such a replacement remain unclear.

Geologic data shows that the Death Valley region within the Basin and Range Province has accommodated the greatest amount of highly localized, upper crustal extension (150 km) in this province. Seismic imaging captures the trace

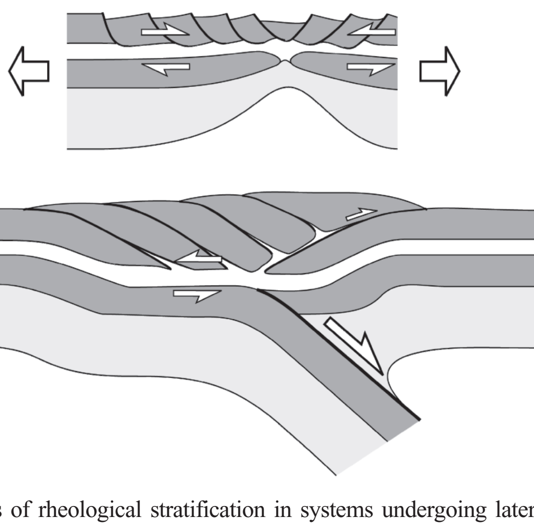

Figure 4.5 Models of rheological stratification in systems undergoing lateral extension (above) and thickening (below), showing weak a lower crustal layer that decouples deformation in the upper mantle from the faulted upper crust (J.-P. Brun, unpublished data).

of a "mylonite front" exposed at the surface in the extended area to a depth of 20 km beneath the Sierra Nevada and demonstrates that the Moho is subhorizontal across the lateral gradient in upper crustal extensional strain between the Death Valley and the Sierra Nevada. The mylonite front is interpreted to reflect the counterflow of "subcrustal asthenosphere" from beneath the Sierra Nevada toward the center of the extended terrain. The compensating layer must have the properties of a weak, possibly molten or partially molten, Newtonian fluid that is unable to maintain horizontal pressure gradients for significant lengths of time. Seismic imaging and sampling of mantle xenoliths from the Sierra Nevada shows that while rifting was occurring in the Death Valley region (between 12 and 3 Ma), a large part of the dense, Sierran mantle lithosphere detached and began to sink into the asthenosphere. At present, very little mantle lithosphere is present, with a localized region of pronounced velocity contrast (>4%) in a cylindrical area that persists to a depth of >200 km beneath the central Sierra Nevada/Great Valley block. These observations suggest that, under at least some circumstances, fault localization occurs in an environment where the lateral flow of extremely weak crustal layers and buoyancy-driven removal of the mantle lithosphere are likely important mechanical controls.

Throughout our discussions of localization, the question arose as to whether we have the tools to measure the degree of strain localization at depth. Are there limits to our ability to resolve the length scale of deforming zones within the lithosphere and asthenosphere?

Shear zones are commonly perceived to broaden with depth. Yet, the resolution of geophysical imaging techniques decreases with depth such that the length scale (e.g., width) of resolvable features also grows with depth (Figure 4.6). This raises the disturbing possibility that conceptual models of deformation broadening with depth may be partly or largely an artifact of limitations in developing high-resolution images of the deformation field.

Studies of fossil fault zones that expose shallow to lower crustal levels have provided such high-resolution views of how deformation varies with depth, as noted above. Drilling of active fault zones (e.g., the SAFOD project on the San Andreas Fault) provide *in situ* information complementary to that obtained from exhumed fault zones. We must, however, continue to refine our tools to provide high-precision geophysical imaging of fault systems in the crust and mantle. The application of seismic techniques such as "Double-Difference Earthquake locations" reveal that microseismicity along several active strands of the San Andreas fault system are restricted to extremely narrow domains (e.g., Schaff et al. 2002). In some cases, the active trace of these fault strands is only 5–10 m wide, and certainly less than 50–100 m wide at depths of 5–10 km in the crust.

In analyzing the interaction of fault systems, it is useful to consider whether deformation occurs in an open or closed system at the lithospheric scale: Is mass conserved or removed from the system? Clearly, the prevalence of tectonic

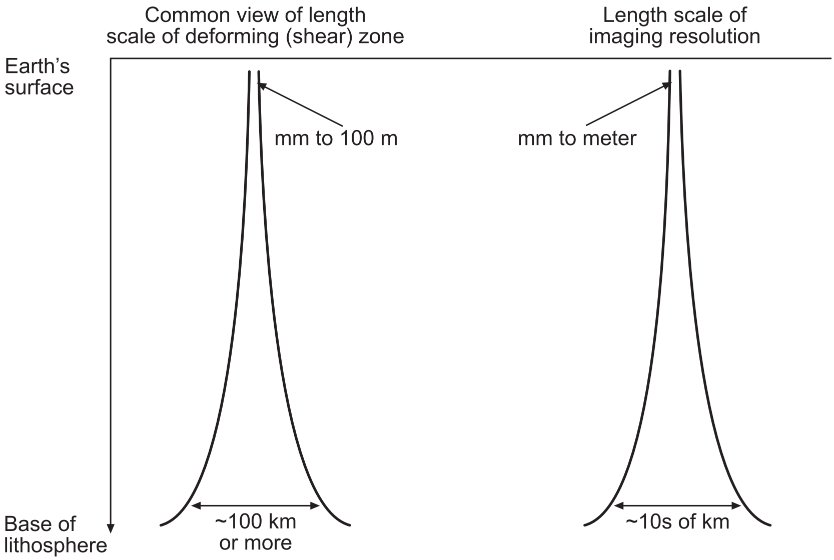

Figure 4.6 Length scale of fault zone versus depth in comparison to the depth dependence of spatial resolution in geophysical imaging (from G. Beroza).

erosion or accretion at convergent margins indicates that subduction zones are open systems from a deformational perspective. But are orogenic systems closed? The Death Valley–Sierra Nevada coupling may reflect a situation in which faulting and its localization has driven a previously closed deformational system into an open mode in which substantial mass (the entire lithospheric mantle?) is removed from the system.

Most major fault zones accommodate motion that is oblique to their traces at the surface. In the shallow parts of large transpressional fault systems, this oblique motion is often partitioned into strike-slip faults and separate convergent structures (thrust faults, folds). For example, along the Alpine fault system of South Island, New Zealand, oblique plate motion, with respect to the orientation of the primary upper crustal fault (the Alpine fault), leads to the juxtaposition of strike-slip motion on the fault with substantial crustal shortening and thickening of crust of the adjacent Pacific Plate (Stern et al. 2000; Sutherland et al. 2000; Walcott 1998). The ~50–100 km wide upper crustal deformational zone that produced the Southern Alps of New Zealand contrasts with the ~1 km wide mylonite zone in the middle crust that is exposed along the exhumed, eastern side of the Alpine fault. Deformation in the middle crust appears to have been much more localized than in the brittle upper crust, where strain partitioning is very pronounced. The component of displacement normal to the plate boundary has increased over the past 5–12 Ma as the motion between the Pacific and Australian Plates has become increasingly oblique with respect to their mutual boundary. What is not clear, however, is how the lithosphere, particularly the subcrustal lithosphere, responds to this increasingly transpressional regime. Does the ductile part of the fault also deform in a transpressional manner, producing a broad deformation zone that mimics the width of the upper crustal convergence zone? Or does the ductile shear zone change its orientation in response to changing plate motions, allowing it to accommodate simple shear while linking the subduction boundaries at both of its ends?

Similarly, the nature of deformation in the mantle lithosphere within large-scale orogenic and intracontinental extension/rift systems is not well understood. Space problems can develop in such settings unless the lithospheric mantle deforms substantially or is removed from the system. Few constraints currently exist to provide definitive answers to this question. What observations are needed, what tools need to be developed, and what disciplines linked in order to address the question of how fault zones behave in different levels of the lithosphere?

There is a definite need for improved geophysical images of mantle deformation. Currently the primary imaging tool is passive, teleseismic information, especially the direction of split shear waves, which provide information on the bulk fabric orientation of upper mantle rocks beneath the observational site (e.g., Tommasi et al. 1999). Less commonly employed, but also effective, are approaches that exploit the pronounced P-wave anisotropy of upper mantle rocks. Combining electrical conductivity anisotropy measurements (e.g., magnetotelluric, MT) with seismic anisotropy measurements is a powerful approach (e.g., Maercklin et al., in press). The frequency-depth kernels for MT measurements have been used to evaluate the depth dependence of any observed anisotropy. These observations are critical to defining patterns and amounts of deformation but are not without limitations. Interpreting seismic wave velocity anisotropy in terms of deformation involves making assumptions about deformation mechanisms and the conditions of deformation. As mentioned above, the volume of rock sampled by the seismic waves broadens with depth (Figure 4.6), and as an integral over the ray path, it is often difficult to isolate the depth interval at which the imaged fabric exists. To complicate matters, the superposition of deformational events at depth may either obscure earlier events or produce a composite structure that reflects none of the individual deformational events.

Other seismological tools include mantle tomography, receiver function analyses, and active source imaging (see Chapter 2, and references therein). All of these can provide constraints on both the three-dimensional structure and the nature of thermal and/or compositional character of the crust and mantle. Improvements in spatial resolution and coverage of all of these tools are needed. Innovative combinations of measured properties (e.g., the ratio of P- and S-wave velocities, $V\text{p}/V\text{s}$) are already used to map the temperature distribution in the lithosphere. Probably the most useful "advances" will be the integration of results from all of these geophysical techniques with detailed field and laboratory observations of structures in active and fossil fault zones.

SUMMARY

Our deliberations on how large fault systems nucleate and grow led to a consensus that progress will entail identifying the determinants of transient motion during faulting and, more generally, regarding structure and rheology in a broad kinematic context, from the Earth's surface down into its asthenosphere. Episodicity associated both with seismic and aseismic faulting ("slow" or "silent" earthquakes) is not restricted to the upper brittle crust and, indeed, may originate in the viscous lower crust and upper mantle. The origins of episodic fault activity are still not known, and several mechanisms were considered, from viscous instabilities nucleating on existing anisotropies to volume changes and fluid flux associated with phase transformations. The consideration of how stress fields interact with existing structures and thermal regimes at depth, and with erosional and depositional regimes at the surface, is crucial to understanding how faults link to form a throughgoing network that weakens the entire lithosphere. Progress will require a multifaceted approach, in which high-resolution imaging of active faults (involving satellite-based and deep geophysical sounding methods) is combined with detailed study of "frozen-in" structures in exhumed, fossil fault zones. Although these disparate methods are individually well established, using them in concert is a challenge that will require adaptation from each specialized part of the Earth Science community. Therein lies the future of research on fault dynamics.

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