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A perspective on the emergence of modern structural geology: Celebrating the feedbacks between historical-based and process-based approaches

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ABSTRACT

Structural geology has emerged as an integrative, synthetic science in the past 50 years, focused on deciphering the history preserved in the rock record and determining the processes of rock deformation. Owing to the nature of structural geology, studies focus on historical elements, such as structural inheritance and tectonic history, and increasingly involve theoretical, process-based approaches. The strength of the field is that it uses these historical- and process-based approaches simultaneously in order to determine the three-dimensional architecture, kinematic evolution, and dynamic conditions of lithospheric deformation over a wide range of spatial and temporal scales.

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In this contribution we focus on significant progress made in understanding shear zones, fault zones, intrusions, and migmatites, both as individual features and as systems. Intrinsic to these advances are insights into the strain history, specifically through the temporal evolution of geologic structures. Increasingly sophisticated geochronological techniques have advanced the field of modern structural geology by allowing age determinations to be linked to rock microstructure and deformational fabrics, from which displacement rates and strain rates can be estimated in some settings. Structural studies involving new approaches (e.g., trenching), and integrated with geomorphology and geodesy, have been applied to study active geologic structures in near surface settings. Finally, significant progress has been made in constraining the rheology of naturally deformed rocks. These studies generally rely on results of experimental deformation, with microstructural analyses providing the connection between naturally deformed rocks and results of experiments. Integration of field-based observations, laboratory-derived rheological information, and numerical models provide significant opportunities for future work, and continues the tradition of simultaneously using historical- and process-based approaches.

INTRODUCTION: HISTORY AND PROCESS

Structural geology has progressed from principally documenting geometrical and historical relations via geologic maps to a diverse science that embraces and advances continuum mechanics approaches, regularly utilizes state-of-the-art analytical and imaging instrumentation, and increasingly relies on sophisticated numerical methods. Practitioners are intensely interested in the physical and chemical processes that result in geological structures and structural fabrics. Yet, structural geologists still make geologic maps, sometimes even with pencil and paper, and are interested in the tectonic evolution (i.e., history) of an area. Structural geology, as it is practiced currently, draws upon and interweaves both historical-based and process-based approaches to answer questions about how the Earth deforms. The historical approach is primarily concerned with the relative or absolute chronology of events; the process approach is mainly interested in mechanisms and models, including deformation mechanisms and the mechanics of deformation. The root of the difference between historical- and process-based approaches in structural geology lies in the broader empirical versus theoretical division apparent in many fields of science. To understand the state of the field, and appreciate the unique perspective of this discipline, it is helpful to understand how these historical and process-based approaches productively coexist.

Three principal goals motivate many structural geology studies: Understanding the three-dimensional architecture, kinematic evolution, and dynamic conditions of geological structures (Fig. 1). Either an empirical or theoretical approach is employed to understand geological structures; the choice depends on the viewpoint of the structural geologist and the tools that he or she uses. An empiricist typically chooses the *inductive* approach, proceeding from examples to general rules or principles. Consequently, for example, she or he evaluates three-dimensional geometry, then derives the kinematic evolution from a detailed analysis of the geometry, and finally moves to a dynamic analysis. All the analyses are ultimately based on a description of a rock's deformed state—the accessible reality of the three-dimensional architecture. In contrast, theoreticians typically choose a *deductive* approach, using rules to understand examples. He or she approaches problems with a rigorous construct of how the world works; for those interested in mechanics, it involves equations of strain compatibility, force equilibrium, constitutive

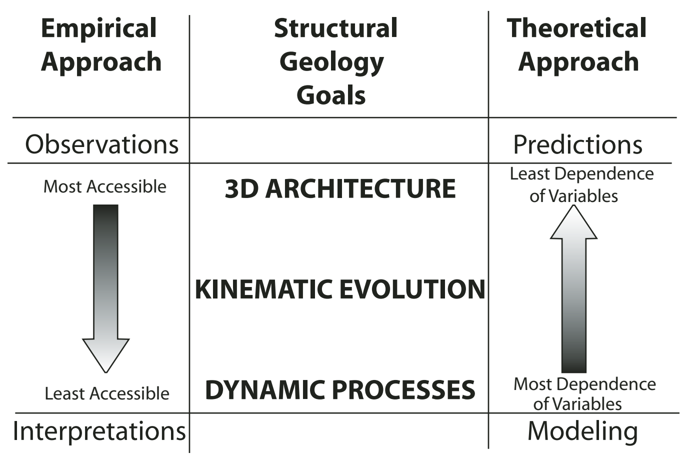


Figure 1. Understanding the three-dimensional architecture, kinematic evolution, and dynamic conditions of geological structures are three major goals in structural geology and can be approached from an empirical or theoretical viewpoint. Empiricists, using *inductive* logic, move from observations to interpretations; hence they start with three-dimensional architecture, model kinematic evolution consistent with their observations (accurate inferences or interpretations), and finally interpret the dynamics (admissible interpretations). A theoretician, using *deductive* logic, conducts dynamic analysis, testing that analysis with predictions about the inferable kinematic evolution and observable three-dimensional architecture.

equations relating the two, etc. For structural geology, theoretical approaches typically involve the dynamics of deformation within a system subjected to specified boundary conditions, from which increasingly advanced predictions about the kinematic evolution and three-dimensional geometry are made. These predictions are then compared to the geometry of natural structures. The mathematical nature of many theoretical analyses requires that they are quantitative, yet they are simultaneously more interpretive because the analyses are based on a construct (the scientist has decided on the important variables or processes and specified the boundary conditions) rather than a physical reality. The above discussion outlines the end-member cases; in practice, any individual scientist often employs both approaches, particularly within structural geology.

These distinctions between empirical and theoretical approaches occur throughout science. Empirical and theoretical approaches are often in tension, and more than once have produced severe scientific disagreements (e.g., Oreskes, 1999). This tension, however, also raises significant new questions, illuminates misconceptions, and generally moves the science forward. Further, each approach has limitations, so they are best used in tandem. The weakness of empirical approaches includes determining the generality of any specific result and the difficulty in distinguishing between correlation and causation. The problem with theoretical approach is the applicability to reality. Theoreticians sometimes assume that a model is correct if it recreates or explains reality in some sense: This assumption is a logical error. In fact, recreating some aspect of "reality" only means that the model is one plausible explanation. While these are real limitations, they do not decrease the importance of quantitative models. Models are extremely useful when (1) they make predictions for what may be empirically observed, and/or (2) they allow one to constrain the importance of a single variable (e.g., sensitivity analysis) within a structural system that can range from microscopic to orogenic scale. The significance of modeling deformation is particularly critical for structural geology because experiments cannot recreate the same spatial scales, temporal scales, and external conditions of rock deformation in orogenic systems.

There is one aspect of geological data, however, that is not present in many other fields of science: its historical nature. Many authors have commented about the importance of time in geological thinking, and the difference of historical (versus ahistorical) or timebound (versus time-independent) reasoning in science (Dott, 1998; Manduca and Kastens, 2012). The essential gist is that any historical science will inherently rely more on the empirical rather than the theoretical approach; the geological time scale, for instance, is fundamentally a result of the empirical approach.

Historical approaches tend to be empirical, because they attempt to infer a time development that must be gathered from observations. In contrast, process-based models typically define a dominant cause of formation and a temporal development for the geological structure, which is similar to a theoretical approach to a problem. There is a significant emphasis on "process" in almost all modern structural geology studies, which commonly rely on or develop a model for the formation of a geological structure or tectonic feature. For process-oriented approaches, a field area is chosen carefully that minimizes the variables influencing the process. There are weaknesses with process-based approaches. Most importantly, process-based approaches may lead an investigator to not recognize or to ignore observations or data that are incompatible with the model; this is poor science, but it is also the reality of human psychology that a person sees what she or he expects to see (e.g., Neisser and Becklen, 1975).

Once again, reality is messier than this easy division. First, most individual structural geologists employ a combination of both approaches for any particular problem. Second, there is not a strict one-for-one correlation between historical and empirical approaches. For example, process-based approaches can be based on non-quantitative models that were developed from empirically based observations. Third, the process of collecting structural data requires that we have to filter the external reality before we start collecting data. The difference between historical- and process-based approaches may result more from the type and degree of filtering than anything else. Open-mindedness to new ideas and unexpected observations, however, is clearly beneficial in all investigations.

The relevance of these philosophical musings is that anyone doing structural geology research works in a continuum between historical- and process-based approaches. Both approaches are essential to progress in our science. It is helpful to recognize where particular studies fall within the spectrum, and to allow for the validity of different approaches. In fact, many important structural geology studies use both approaches in very productive ways (e.g., Mitra, 1976; Rutter, 1976; Davis et al., 1983; Mitra, 1994; Sammis and Ben-Zion, 2008).

This chapter is organized into five parts, which reflect major advances in the field. The first part is a brief reflection on the relation of **plate tectonics** to structural geology. The second section on **three-dimensional architecture** reviews advances in understanding the three-dimensional geometry and the inferences about displacement, strain, and stress from analyses of geological structures. We focus on localization zones (faults, shear zones) in the crust, and the role of intrusions and melt, because of significant advances made on these subjects. We do not address folds and folding because excellent and comprehensive reviews exist on fold development (Hudleston and Treagus, 2010) and fluid flow in folds (Evans and Fischer, 2012): There is little that we could add to these recent contributions. The third section focuses on **application of geochronological techniques in structural geology studies,** which has allowed structural geologists to understand the timing within tectonic systems, estimate strain rates in particular settings, and directly date fault motion. The fourth section addresses the application of structural geology to the field of **active tectonics**, which uses geomorphic and geodetic information to address ongoing deformation. The final section focuses on **rheology** of naturally deformed rocks. Using the results of experimental deformation, structural geologists have used microstructures in naturally deformed rocks to make inferences about the dominant deformation mechanisms and hence the rheology. We conclude with rheology because it is inherently an interdisciplinary subject that likely reflects some of the future for structural geology.

Summarizing 50 years of progress in any field is nearly impossible. There is simply too much high-quality work, over a variety of subjects, to cover them all adequately. Not only is our choice of subjects biased, but even the organization of this chapter produces a bias. An additional limitation is that the organization of this book divides structural geology from tectonics (see Moores et al., this volume) into separate chapters. Although this is a useful division, many structural geologists attempt to work across this boundary, and consequently much work that integrates these two viewpoints is not addressed. We admit these failings here and hope that this chapter is none-the-less useful.

PLATE TECTONICS AND STRUCTURAL GEOLOGY

Plate tectonics theory significantly affected every subdiscipline within the geological sciences, but its effect on structural geology was particularly profound. The implications of plate tectonics were immediately recognized in orogenic belts, for understanding the origin of ancient orogens (e.g., Appalachian Mountains; Dewey and Bird, 1970), appreciating the role of terrane accretion in orogenesis (e.g., the North American Cordillera; Monger et al., 1972; Coney et al., 1980; Oldow et al., 1989), and unraveling the evolution of active tectonic belts (e.g., California margin; Atwater, 1970). A large part of structural geology done in the past 50 years—and arguably a majority of structural geology done in North America—has been an attempt to fit complex regional geology into the context of plate tectonics. This task was easiest in regions where past plate motions could be more easily inferred (e.g., western margin of North America in the Tertiary), increasingly difficult for more distant times in the geological past, and most difficult where no modern analogues are present (e.g., parts of the Precambrian).

Having the conceptual framework of plate tectonics for understanding geological structures has had two major impacts. First, it resulted in a more process-based approach. Prior to this, understanding process-based rock deformation could really only be done on small scales for structural geology research (outcrop to regional), where the boundary conditions of deformation could be reasonably inferred. Deformation at the scale of mountain belts could not really be understood before the advent of plate tectonics, as the ultimate cause of the large-scale motions was not understood. With the advent of the plate tectonic conceptual model, structural geology went from primarily an observationally based science (e.g., quadrangle mapping), to a synthetic science focused on understanding the genesis of geological structures.

Second, the plate tectonic paradigm has fostered integration across the disciplines within the geological science. Prior to 1970, there was significantly less interaction between the different subfields. Plate tectonics provided a framework where the information from different subfields became relevant to each other; structural geology became one of the tools used to look at the Earth system. As a result, structural geology is no longer done in isolation from other fields, but rather as a necessary complement to understanding the overall history. The first of these two effects emphasizes process-based understanding, whereas the second effect emphasizes historical-based understanding, as tectonic history became the common goal for integrated work. Consequently, the net effect of plate tectonics was to heighten that tension of "history" versus "process," but it also led to an increased sophistication of both. An additional contribution of plate tectonics that was specifically important for structural geology was the recognition of the importance of horizontal movements and forces, as opposed to vertical tectonics, in determining the behavior of Earth systems.

Geologists, and structural geologists in particular, played an important role in supporting continental drift hypotheses (e.g., Argand, 1924; du Toit, 1937; Griggs, 1939; Holmes, 1944) and facilitated the plate tectonics revolution (e.g., Dewey and Bird, 1970; Burke and Dewey, 1973). Their most significant contribution, however, was arguably in understanding continental deformation (and, now, also oceanic deformation) after the plate tectonic paradigm was developed primarily for oceanic lithosphere (e.g., Molnar, 1988). While that work is not completed, major reinterpretations of large regions are likely to become more infrequent. The core work of structural geology—the deformation of geological materials as recorded primarily by the rock record—predated the plate-tectonics paradigm and remains a vibrant source of research. Consequently, this contribution focuses on what has been discovered about rock deformation for the past 50 years, and does not explicitly cover the tectonic synthesis efforts undertaken by structural geologists.

THREE-DIMENSIONAL GEOMETRY

This section endeavors to describe advances in our understanding of the three-dimensional geometry of geological structures, as well as the evolution of those structures. The majority of these developments are products of direct observation; some resulted from the application of numerical models, and others were made through technological advances (such as 3D seismic images used for petroleum exploration).

Shear Zones

Individual Shear Zones

It is difficult to overestimate the influence of the analysis of shear zones by Ramsay and Graham (1970). This paper used a straightforward kinematic model of a spatially varying displacement field to predict variations in finite strain quantitatively. This linkage of the offsets of markers across zones of localized shearing with the magnitudes of strain within the zones underscored the relationships between displacement gradients and strain and the variations in fabric intensity observed in natural settings.

While the evaluation of finite strain did not start with Ramsay and Graham (1970)—antecedents include the Wisconsin School of Structural Geology: Leith (1937), Cloos (1947), Flinn (1956), Ramsay (1967), Elliott (1968), Dunnet (1969)—this was one of the first papers to outline methods of approaching inhomogeneous strain. The methods presented had sufficient generality to reproduce natural deformation gradients, yet sufficient simplicity that the methods were readily visualized and appreciated. This paper and Ramsay's structural geology textbook (Ramsay, 1967) solidified the reputation of the "Ramsay school of finite strain," which dominated progress in the field of structural geology for the next 20 years. At the heart of the Ramsay school was the careful documentation of naturally deformed rocks in the field, and the characterization of finite strain. The graduate students and post-doctoral fellows who worked with John Ramsay were most influential in Europe and North America, and continued work that combined mathematical analysis and careful observation of deformed rock.

The next major advance in analyzing the geometry of shear zones was also made in Europe. Structural geologists studying the South Armorican shear zone in France were among the early workers to recognize the significance of shear-sense indicators in determining the kinematic history of shear zones (Berthé et al., 1979). Law and Johnson (2010) show that B. Peach and J. Horne recognized and utilized what we now call shear-sense indicators to elucidate the structure of the Moine thrust zone (Scotland) in works published in 1888 and 1907 (Peach et al., 1888; 1907). Also, Eisbacher (1970) clearly recognized shear-sense indicators in the Cobequid shear zone of eastern Canada, including excellent sketches of asymmetric "mica fish." This valuable shear-sense indicator was subsequently emphasized in Lister and Snoke (1984). An *International Conference on Shear Zones in Rocks* held in Barcelona in 1979 (with conference papers appearing in volume 2, no. 1–2, of the *Journal of Structural Geology)* was instrumental in disseminating work on shear zone geometry, and shear-sense indicators were quickly applied to a variety of field settings (e.g., Burg et al., 1981) in Europe. This work was generally introduced in the United States during a Penrose Conference ("Significance and Petrogenesis of Mylonitic Rocks," convened by J. Tullis, A. Snoke, and V. Todd) in 1981. Simpson and Schmid (1983) were also influential in familiarizing North American structural geologists with these advances (as were Hanmer and Passchier, 1991). This approach was quickly utilized to make major tectonic contributions, such as the resolution of the extensional character of core complexes in the U.S. Cordillera (e.g., Davis et al., 1987) and the recognition of the normal-sense shear zone (central Tibetan detachment) at the base of Mount Everest (Burg and Chen, 1984).

Shear-Zone Evolution

In his review of shear-zone geometry, Ramsay (1979) demonstrated that the finite strain and offset marker patterns characteristic of shear zones need not result solely from simple shear deformation. C. Passchier calculated the relative percentage of pure shear to simple shear in zones of distributed deformation (Passchier, 1986, 1988). His approach was to characterize the deviation from simple shear using kinematic vorticity, based on the concept by Truesdell and Toupin (1960), and introduced in the structural geology literature by Means et al. (1980). Kinematic vorticity has subsequently become the major means for characterizing different, homogeneous strain paths, and methods include evaluating finite strain gradients (e.g., Fossen and Tikoff, 1993), forward and back rotation of clasts (e.g., Passchier, 1988), and the effect of tails on back-rotation and maximum-shear-strain-rate planes (Simpson and DePaor, 1993).

Kinematic modeling provided insight into three-dimensional flow within shear zones. This work initially involved transpressional-transtensional kinematics, first with finite-strain approaches (Sanderson and Marchini, 1984) and later with more continuum-mechanics–based approaches that emphasized evolution (e.g., Ramberg, 1975; Tikoff and Fossen, 1993; Iacopini et al., 2007; Davis and Titus, 2011). Increased sophistication of the models involved strain partitioning (Tikoff and Teyssier, 1994) and triclinic deformation (e.g., coaxial components not parallel to shear components of deformation; Jiang and Williams, 1998; Jones and Holdsworth, 1998; Lin et al., 1998; Jones et al., 2004) (Fig. 2). These models were used to understand shear zones with lineations parallel to the vorticity vector (the internal rotation axis controlled by the simple-shear component of deformation) (e.g., transpression; Tikoff and Greene, 1997), folds forming in transtensional settings (e.g., Krabbendam and Dewey, 1998), along-strike variations in fabric types in shear zones (Lin and Jiang, 2001), and major extrusion of shear zones (Vannay and Grasemann, 2001; Xypolias and Koukouvelas, 2001). In other cases, field work on natural shear zones with abundant structural data was not consistent with the existing kinematic models, and new conceptual models were created (e.g., "leaky transpression," Czeck and Hudleston, 2003). The net result of the kinematic modeling and field studies of shear zones was the recognition of shear zones that deviated significantly from the simple-shear model of Ramsay and Graham (1970). Figure 2 shows several of these types of shear zones, with the orientations of lineation, foliation, and vorticity vector.

Increased sophistication of field observations linked with kinematic modeling advanced a re-evaluation of the kinematic history of shear zones. A significant amount of work was done to constrain shear sense indicators following the recognition that not all shear sense indicators record the same direction of motion owing to the coaxial component of deformation (Simpson and DePaor, 1993; Passchier, 1997). Wallis (1992) was an influential early study of non-simple shear zones, combining crystallographic preferred orientations (CPO), finite strain, mesoscopic field fabric, and fields of shortening or elongation to constrain strain paths. Work on naturally deformed shear zones suggests that they deviate strongly from simple shearing (e.g., Law et al., 2004; Bailey et al., 2007; Jessup et al., 2007). The relative merits of the different criteria for determining shear sense and kinematic vorticity is an active source of research.

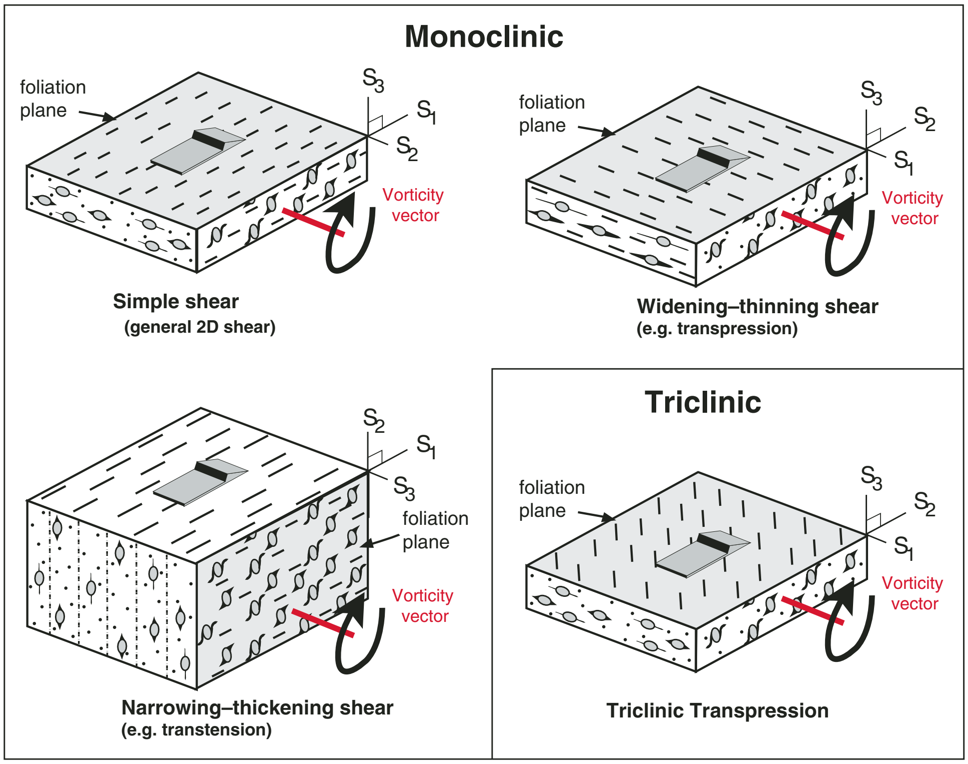


Figure 2. The orientation of foliation, lineation, and shear sense indicators within shear zones. Simple shear zones contain vorticity vectors (expressed in the rock by consistent sense-of-shear indicators) perpendicular to the lineation and within the plane of foliation. Kinematic modeling suggested the occurrence of other possibilities, including vorticity vectors parallel to the lineation (e.g., transpression), vorticity vectors perpendicular to the foliation (e.g., transtension), and vorticity vectors oblique to both lineation and foliation (e.g., triclinic transpression). The patterns predicted by transpression and triclinic transpression have been found in natural shear zones.

Crystallographic preferred orientations (CPO), initially determined using X-ray diffraction or optically with a universal stage, and now mainly by electron backscatter diffraction (EBSD) techniques (e.g., Prior et al., 1999), are useful to constrain deformation within shear zones. CPO was recognized in metals and deformed rocks and exploited nearly a century ago (see discussion in Weiss and Wenk, 1985). CPO analysis can constrain which deformation mechanisms accommodated deformation, deformation kinematics, and, in some cases, the temperature and pressure conditions of deformation of monomineralic rocks (e.g., ice, quartzites, carbonates, pyroxenites, and dunites (see also Wenk, 1985, and the Rheology section of this chapter). Naturally deformed rocks from shear zones associated with the Moine thrust belt (Scotland) have been repeatedly examined as our understanding of the development of CPO evolved (e.g., Phillips, 1937, 1945; Christie, 1960, 1963; Law et al., 1984; Schmid and Casey, 1986; Law and Johnson, 2010; Law et al., 2010; White, 2010). Other areas have also contributed to our understanding of CPO, including strongly sheared rocks from the thrust fault zones of the European Alps (e.g., Schmid et al., 1981; Schmid, 1982; Schmid and Casey, 1986) and deformed ultramafic rocks (e.g., Avé Lallemant and Carter, 1970; Nicolas, 1989). Studies such as these have provided important insight to our understanding of deformation kinematics, kinematic vorticity, and deformation conditions in these settings. Interpretation of CPO patterns is facilitated by theoretical modeling (using the Taylor-Bishop-Hill theory), which can simulate the observed CPO by specifying active slip systems (e.g., Lister et al., 1978; Lister and Hobbs, 1980). Significant improvements have occurred in the modeling of CPO development, including numerical modeling with visco-plastic self-consistent (VPSC) approach. The VPSC allows for the simulation of a CPO through intra-crystalline glide (Molinari et al., 1987; Wenk et al., 1989; Lebensohn and Tomé, 1993). The active slip systems must be input with the correct rheological parameters for the given system or mineral. This modeling has reproduced CPOs that are very similar to natural and experimental CPOs and has demonstrated that glide along basal planes is consistent with quartz deforming in most geological conditions in high strain environments (shear zones) (Morales et al., 2011). Similar use of VPSC modeling on ultramafic rocks has been able to reproduce a variety of fabrics observed in natural peridotites (e.g., Ben Ismail and Mainprice, 1998) in the upper mantle (e.g., Tommasi et al., 1999) or predict fabrics formed at mantle conditions (e.g., Mainprice et al., 2005).

A final issue in studies of shear zones is what role, if any, preexisting features—particularly fractures, veins, and faults—play in controlling the location and microstructural character of shear zones. One well-known example concerns the development of shear zones in granitic rocks in the Sierra Nevada (Segall and Simpson, 1986; Bürgmann and Pollard, 1994). Shear zones in core complexes are another example, as they are progressively overprinted by increasingly discontinuous ("brittle") deformation zones during exhumation (e.g., Davis et al., 1987). The opposite path (cataclastic deformation preceding crystal-plastic deformation) is also possible, but is not always as a result of burial. For example, Goodwin and Wenk (1995) demonstrated that cataclasis in narrow fault zones preceded mylonitization in the Santa Rosa mylonite zone, California. Cataclasis demonstrably occurred at higher temperatures than mylonitization (Goodwin and Renne, 1991), leading Goodwin and Wenk (1995) to speculate that brittle failure accommodated higher regional displacement rates than could be accomplished by ductile flow.

Shear Zone Networks

A related research direction has focused on the geometry of shear zone arrays (Fig. 3). Work focused on basement rocks below thrust sheets noted heterogeneous deformation with anastomosing shear zones surrounding lenses of weakly deformed rocks (e.g., Mitra, 1979; Ramsay and Allison, 1979). Work on the basement massifs of the French Alps confirmed the style of deformation, suggesting that the lozenges of deformed rocks could be used qualitatively as an indicator of the type and amount of strain (Fig. 3; Choukroune and Gapais, 1983; Gapais et al., 1987). The intersections of these anastomosing zones were particularly problematic in terms of understanding strain, and theoretical work has focused on strain compatibility at these intersections (e.g., Lamouroux et al., 1991; Pennacchioni and Mancktelow, 2007) and the temporal development of the arrays (Fusseis et al. 2006; Fusseis and Handy, 2008). Detailed kinematic analyses indicate that anastomosing shear zones can deviate strongly from simple shear deformation (e.g., Arbaret and Burg, 2003; Bhattacharyya and Hudleston, 2001; Baird and Hudleston, 2007) consistent with theoretical models for overall strain compatibility (Hudleston, 1999). Furthermore, shear

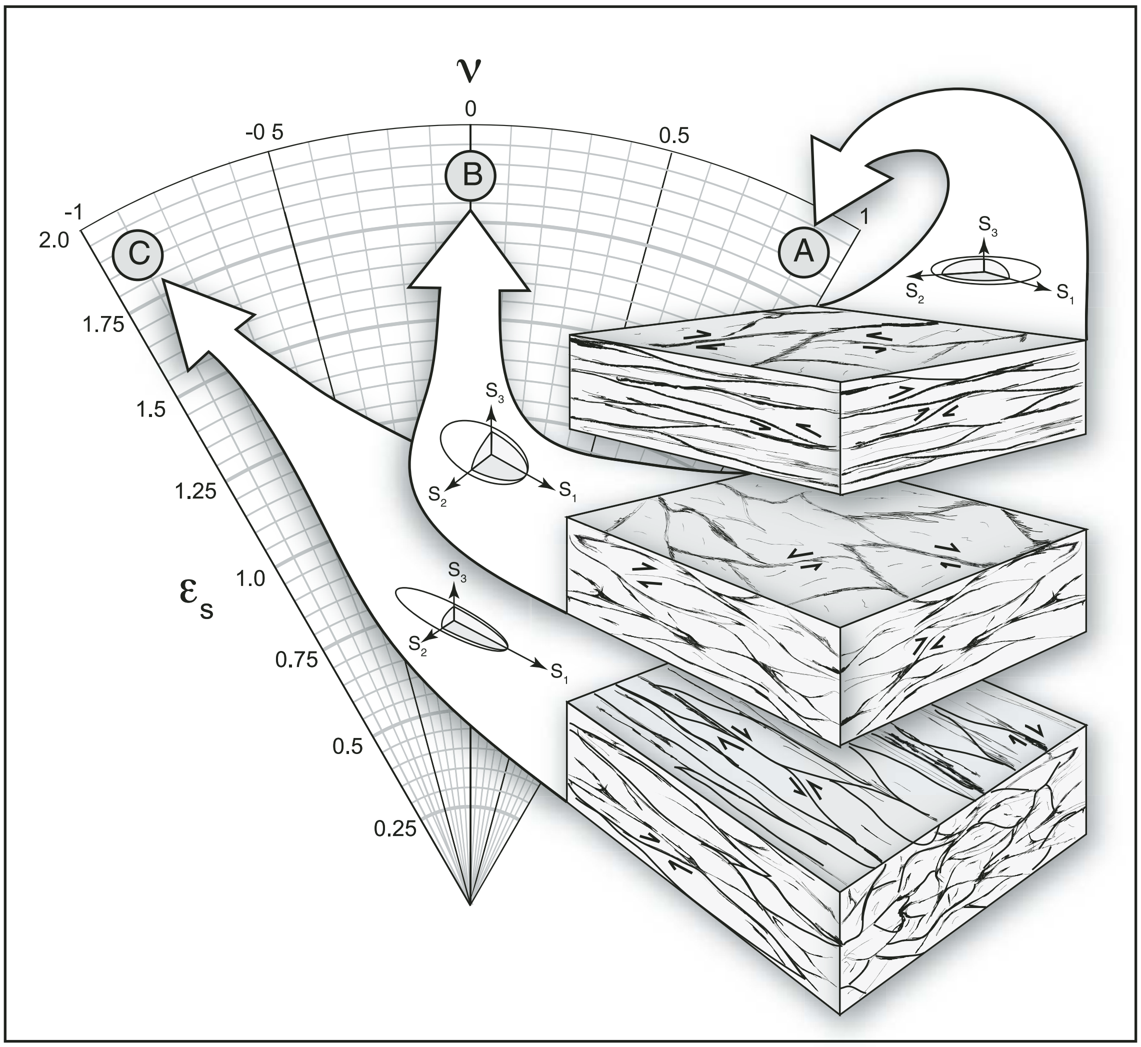


Figure 3. The three-dimensional geometry of anastomosing shear zones, separating lozenges of undeformed or less deformed rocks, shown on a background of a Hsü plot for finite strain (magnitude is distance from origin; shape is prolate to oblate across the top). The 3D pattern of the lozenges reflects the three-dimensional finite strain recorded by the rock (shown by cut-away ellipsoids), with prolate lozenges indicating prolate strain (e.g., Choukroune and Gapais, 1983). The same pattern is observed in biotite selvages in migmatites in the study of Kruckenberg et al. (2010), from which this figure was modified.

zones within networks form at multiple times during progressive deformation, and Carreras et al. (2010) developed criteria to determine the order of formation.

A major question that remains about shear zones is what causes them to form where they do? This issue remains a question for a variety of reasons, including the numerical modeling of viscous materials that cannot initiate shear zones (strain-localization zones) without a nucleation point. There are two dominant hypotheses. First, Mancktelow and Pennacchioni (2005) argue that shear zones are controlled by mechanical instabilities, and hence they support the concept that many shear zones have brittle precursors (for a specific example, see Fusseis and Handy, 2008). Second, it is possible that shear zones localize in areas of competency contrast (e.g., Goodwin and Wenk, 1995; Goodwin and Tikoff, 2002) as a result of strain incompatibility. This debate is unresolved.

Features Associated with Shear Zones

Shear zones generally display strain gradients, often visible in terms of increased fabric development, which can commonly be characterized in terms of finite strain (using deflected marker layers or strain markers). Strain gradients allow structural geologists to utilize space (or distance)-for-time substitutions. For example, fabric development in the highest strain part of a shear zone is assumed to have initially been similar to the lowest strain part of the same shear zone; consequently, the strain gradient is critical for interpreting progressive development of structures. Furlong et al. (2007) provide a good review of the limitations of such an approach, specifically when applied to large-offset faults. However, the same space-for-time substitutions are often used in experimental deformation. Simpson (1983) also demonstrated how strain gradients can be integrated to determine the overall offset of the shear zone.

Means (1995) envisioned three "end-member" types of shear zones, where shear zones thickened, maintained constant thickness, or thinned as offset across the zone accrued. These models tied the geometric evolution of the zone to the evolving rheology of the rock within the shear zone, so that shear zone thickening was in response to strain hardening, shear zone thinning was in response to strain softening, and no change in shear zone thickness was a result of steady-state flow (also see Hull, 1988). These models presumed steady deformation kinematics for the shear zone and the enveloping rock. More recently, Horsman and Tikoff (2007) and Vitale and Mazzoli (2008) used finite strain gradients within shear zones to constrain the kinematic vorticity and to evaluate whether the kinematic vorticity remained constant over time.

Sheath folds are commonly found in shear zones. Sheath folds were initially thought to result from high-strain deformation of perturbations with no significant competency differences (e.g., Cobbold and Quinquis, 1980; Vollmer, 1988). A more nuanced view of sheath folds has evolved, in which sheath folds are thought to reflect perturbations in the flow field (e.g., Alsop and Holdsworth, 2002).

Finally, vein arrays are common within and directly adjacent to shear zones. En echelon veins commonly form in low-strain shear zones in weakly metamorphosed rocks; en echelon arrays often occur in conjugate pairs that together accommodate bulk pure-shear displacement fields. Veins appear to form periodically within shear zones, often subsequently acting as rigid markers and displaying inverse drag (Hudleston, 1989). Cox and Knackstedt (1999) noted that most vein-rich deposits were within releasing bends of the regional-scale shear zones within the Yilgarn Province, Australia.

Relation to Metamorphism

Shear zones also display interesting interactions between the growth of new minerals and deformation. Among other important contributions, this interaction allows determination of the pressure-temperature-time (P–T–t) paths of shear zones (e.g., Spear and Peacock, 1989). This topic is not addressed here, except to note the somewhat contentious role of porphyroblasts in deformation analyses. Porphyroblast-inclusion relationships were originally documented by Zwart (1962) and Spry (1963), and allowed for the determination of the timing of thermal pulses (porphyroblast growth) versus the timing of deformation (rotation of inclusion trails). The growth of snowball garnets, which "wrap" the foliation during rotation, are perhaps the most obvious examples (Rosenfeld, 1970). These original relationships have been challenged and modified (Vernon, 1975, 1989; Bell and Rubenach, 1983; Bell et al., 1986; Passchier et al., 1992; Johnson, 1999). The principal controversy is whether or not porphyroblasts rotate during noncoaxial flow. The resolution to the controversy seems to be that porphyroblasts do rotate (e.g., Johnson, 2009), but probably not as much as is interpreted by using a Jeffreys rotation model that requires perfect coupling between a rigid ellipsoid and the matrix in a viscous flow (Jeffrey, 1922). This slow rotation results from strain localization on the edges of porphyroclasts and non-cohesion of the porphyroclasts to the matrix (Ildefonse and Mancktelow, 1993; Iacopini et al., 2011).

Faults

Individual Faults

The seminal paper by Watterson (1986), which compiled data on the dimensions and displacements of natural faults and proposed a geometric model for fault growth, led to an explosion of interest in fault geometry over the next twenty years. Earlier work on fault geometry drew upon maps and widely spaced, well-constrained cross sections (e.g., Elliott, 1976; Gudmundsson, 1980; Scholz, 1982; Muraoka and Kamata, 1983), but the advent of displacement profiles on faults from three-dimensional seismic reflection and coal mine data was a key technical development that provided detailed information on the dimensions and displacements of large numbers of faults. The growing body of data facilitated the refinement of Watterson's geometric model for the growth and evolution of individual faults. The observed relationships between fault length, fault width, and maximum displacement were used to argue that faults exhibit self-similarity or self-affinity (Watterson, 1986; Walsh and Watterson, 1988; Cowie and Scholz, 1992c); later work argued that vein arrays (Johnston and McCaffrey, 1996) and stylolites (Karcz and Scholz, 2003; Renard et al., 2004; Peacock and Azzam, 2006) also exhibited self-similarity or self-affinity. Fault/fracture self-similarity or self-affinity is evidence that these structures are fractal (Villemin and Sunwoo, 1987; Marrett and Allmendinger, 1991), a characteristic that can be exploited to understand the characteristics of fault/fracture populations, particularly the cumulative displacement or elongation accommodated by the population along transects (e.g., King, 1983; Marrett and Allmendinger, 1992; Walsh et al., 1991; Gross and Engelder, 1995).

Application of the space-for-time substitution, drawing on the geometric similarity implied by scaling relationships of Watterson (1986) and Walsh and Watterson (1987, 1988), supported fault evolution models in which (1) fault displacement near the center of the fault was envisioned as the product of repeated slip events (in some conceptual models, earthquakes) over the whole surface of the fault, and (2) each slip event was associated with an incremental lengthening of the fault surface (Fig. 4). A strong theme in the first of these studies was the growth of faults by seismic events (Sibson, 1989; Cowie and Scholz, 1992a), and some studies extended interpretation of the scaling laws to mechanics (e.g., Cowie and Scholz, 1992c). Subsequently, several ideas for how the total displacement profile of a fault might be built up from slip of parts of fault surfaces were conceived (e.g., Cladouhos and Marrett, 1996), partly inspired by observations of earthquake ruptures on active faults. Ultimately, alternative ideas were developed for fault growth, with rapid initial establishment of the final fault length followed by increasing accumulation of fault displacement (Walsh et al., 2002; Nicol et al., 2005).

Measured displacement profiles on faults (e.g., Rippon, 1984; Barnett et al., 1987), with maximum displacement magnitudes near the center of a fault and smooth decreases in fault displacement along radial paths toward the tip line of the fault, were key elements in developing models for the growth of individual faults. Actual displacement profiles commonly are asymmetric, owing to interactions between faults and lithologic boundaries, faults and Earth's surface, or faults with other, overlapping faults. Several research groups recognized that composite displacement profiles constructed for fault systems consisting of interacting splays approximated the displacement profiles of individual isolated faults (Barnett et al., 1987; Peacock and Sanderson, 1991, 1994; Dawers and Anders, 1995), whether the splays were "hard linked" (splays demonstrated to connect and form a continuous movement surface) or "soft linked" (faults overlapping, panels of inclined or folded rock separating splays) (Walsh and Watterson, 1991; Schlische, 1992; Anders and Schlische, 1994; Peacock and Sanderson, 1994; Nicol et al., 1996).

Fault Networks

For many years the interpretation of fault networks was dominated by the approach, explained fully in the influential

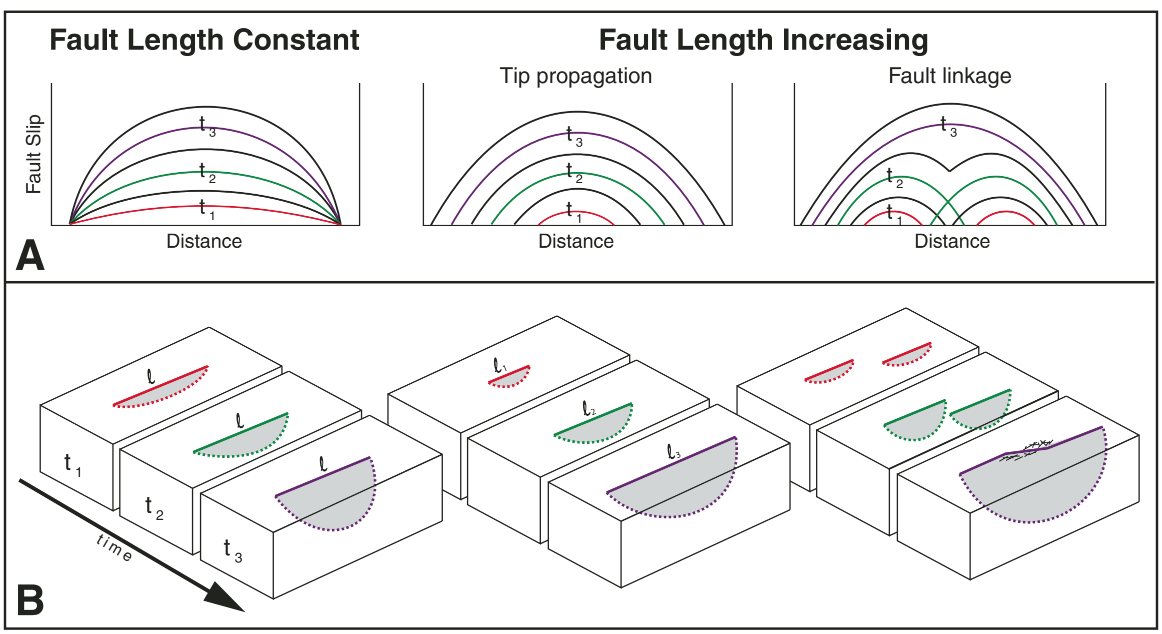


Figure 4. Schematic diagrams, showing models for the accumulation of fault slip during repeated earthquakes. (A) The graphs show offset versus distance for discrete seismic events, for constant length, and increasing length fault-growth models (e.g., Walsh and Watterson, 1988; Scholz, 1990; Walsh et al., 2002). These models apply to faults with cumulative displacements that decrease from a centrally located maximum and are end members in a continuum. (B) Three-dimensional block models of fault growth, corresponding to 1, 2, and 3 of above. The color of the faults in part B corresponds to the same color line in the graphs of part A.

book by Anderson (1951), that faults are products of Coulomb failure, controlled by the orientations of the most compressive and least compressive principal stress. In terms of the orientation of faults and the geometry of slip on them, the Coulomb approach is essentially a two-dimensional one, predicting that (1) individual fault surfaces parallel one of two conjugate planes whose intersection is perpendicular to the plane containing the most and least compressive stress directions, (2) conjugate planes meet at ~60° angles with the most compressive stress parallel to the acute bisector of the planes, and (3) fault slip vectors lie within the plane containing the most and least compressive stress directions. Some fault arrays do exhibit this geometry, but the faults in many natural arrays have neither the requisite conjugate sets whose intersections define a single direction nor the fault slip vectors confined to a plane perpendicular to fault intersections. Departures from the predicted geometry are expected in anisotropic rock or during fault reactivation (e.g., Donath, 1961). Even fault arrays formed in previously un-faulted, relatively isotropic rock regularly exhibit (1) a predominance of faults parallel to only one of the conjugate planes predicted by Anderson (1951), (2) intersections of faults with a variety of directions, or (3) slip vectors oblique to fault intersections. Orthorhombic fault sets, where faults form with four or more orientations, are common in nature and in laboratory experiments (Oertel, 1965; Aydin and Reches, 1982; Reches, 1983; Krantz, 1988; Healy et al., 2006). Reches and Dieterich (1983) derived a formalism that relates three-dimensional fault arrays exhibiting orthorhombic symmetry (and accommodating bulk, irrotational strain) to triaxial stress states. More recently, polygonal fault systems have been described from mudrock or shale sequences (Cartwright and Lonergan, 1996; Watterson et al., 2000) and were attributed to volume changes or density inversions below the faulted sequence.

The Anderson (1951) model emphasized interpreting faults in terms of stresses to the exclusion of considering the displacements or finite strain recorded by groups of faults. Some workers (e.g., Ode, 1960; Kehle, 1970) conceived of faults as elements in a displacement or velocity field, but most analyses interpreted faults and fault arrays in terms of the stresses inferred to be responsible for their formation. Arthaud (1969) and Arthaud and Mattauer (1969) pioneered an approach that envisioned fracture and fault arrays from the perspective of the cumulative displacements they accomplish in aggregate, and they developed techniques to determine principal directions of incremental shortening and elongation. This approach was further developed and automated by Carey and Brunier (1974), Carey et al. (1974), Etchecopar et al. (1981), Angelier (1984), and Michael (1984), who interpreted the calculated incremental shortening and elongation directions as principal directions of a "paleostress" tensor. Equating incremental strain principal directions to incremental stress principal directions is justified for isotropic materials undergoing small strains. If the bulk deformation is coaxial, limitation to small strain deformations can be relaxed. In this sense, paleostress analyses are comparable to analyses of deformation twinning in carbonates, where early work focused on constraining incremental strain directions (e.g., Groshong, 1972; although Turner, 1953, focused on stress orientations), and later studies used results to infer principal stress directions (Jamison and Spang, 1976; Craddock and van der Pluijm, 1999; Lacombe, 2007). Analyses of deformation twinning are widely presumed to be reliable in situations where strain magnitudes are small, strain paths do not include bulk rotation, or where rocks are not subjected to multiple deformation episodes. Under comparable conditions, incremental strain directions inferred from fault arrays are thought to be as reliable as those inferred from deformation twins.

An important justification for interpreting fault slip data in terms of stresses ("dynamic analyses") is the "Wallace-Bott" hypothesis that slip on a fault plane will occur in the direction of maximum resolved shear stress (Wallace, 1951; Bott, 1959), which underpins several classic methods of paleostress analysis such as the right dihedra (Angelier and Mechler, 1977), right trihedra (Lisle, 1987), and inversion methods (e.g., Angelier, 1984, 1994; Gephart and Forsyth, 1984; Lisle et al., 2001). The assumption seems justified on the basis of numerical modeling (Dupin et al., 1993; Pollard et al., 1993) and field data (Lisle and Srivastava, 2004). However, a significant difficulty for both kinematic and dynamic analyses of fault slip data is posed by collections of fault slip data that may belong to different deformation events. Vigorous research has been directed toward this problem in the last decade or so (e.g., Nemcok et al., 1999; Yamaji, 2000, 2003; Shan et al., 2003, 2004; Liesa and Lisle, 2004; Shan and Fry, 2005; Yamaji et al., 2006; Otsubo et al., 2006), leading to some promising methods for detecting the existence of different stress tensors within a collection of fault slip data, which commonly lack evidence for age relationships between faults. None of the previous paleostress analyses take account of the influence of the fault itself on the stress state: they are "faultless." Kaven et al. (2011) introduced a mechanical analysis for paleostress inversion that overcomes this problem. The results of this method agree in some circumstances with previous methods, but there may also be discrepancies. Arguably, paleostress analyses can be used to draw significant conclusions on a global scale about crustal stress states (e.g., Lisle et al., 2006).

Molnar (1983), Jamison (1989), Wojtal (1989), and Marrett and Allmendinger (1990) outlined different techniques that compile displacements on faults in arrays to calculate finite strain principal directions and magnitudes. These displacement field techniques are not restricted to small strain magnitudes, although it is not clear if they capture the rotational component of the finite strain. D. Rouby and colleagues used least-squares palinspastic restoration inverse methods to show how displacements on individual faults can be combined to produce displacement fields (Rouby et al., 1993, 1996). Twiss and Gefell (1990) and Twiss et al. (1991, 1993) formalized a technique capable of analyzing 3D fault arrays, even with monoclinic symmetry. Their analyses extend the treatment of fault arrays to include general, rotational strain states. Twiss et al. (1991, 1993) and Twiss and Unruh (1998) used the symmetry of tangent lineations, geometric elements that reflect both the orientation of a fault and the orientation of the slip on it, to infer the general character of the velocity or displacement field accommodated by faulting (though their analysis does not constrain strain magnitudes). Twiss's analysis is based on the inference that the velocity gradient tensor, not the stress tensor, constrains the geometry of slip on faults in an array. These advances are perhaps nearing an apogee with attempts to combine kinematic and dynamic analyses (Žalohar and Vrabec, 2008), and the introduction of Cosserat continuum mechanics to describe the deformation of faulted crust, starting from a kinematic approach to faults (Žalohar and Vrabec, 2010).

Structures Associated with Fault Zones

In an influential series of papers published almost 50 years ago, Chinnery (1964, 1966a, 1966b) used calculations of the stresses around a dislocation embedded in an elastic solid to argue that fault slip (1) relieves shear stresses along the trace of a vertical strike-slip fault, and (2) concentrates stresses at the termination of the fault. His predictions for the orientations of secondary faults at fault terminations were a significant improvement over previous analyses. The observation of characteristic geometries of fractures, veins, and stylolites at fault terminations (e.g., Rispoli, 1981) was another step toward wider understanding of stresses around fractures and faults. Computational models were a critical factor in the rapid maturing of the use of linear elastic fracture mechanics (LEFM) models to predict stresses and displacements adjacent to faults (e.g., Segall and Pollard, 1983; Pollard and Segall, 1987). The concepts of fracture modes I, II, and III, originally developed to describe types of fracture in relation to loading conditions (e.g., Lawn and Wilshaw, 1975), were found to be applicable to structural geology studies (Fig. 5A), with mode I fractures corresponding to extension fractures, and modes II and III corresponding to parts of faults where displacements are perpendicular and parallel to fault tip lines, respectively. These relations facilitated the application of LEFM to faulting, with faults envisioned to form in an elastic, isotropic medium. LEFM models have been particularly successful at reproducing the orientation of secondary fractures at fault tips (Hori and Nemat-Nasser, 1987; Jeyakumaran and Rudniki, 1995; Willemse and Pollard, 1998; de Joussineau et al., 2007), fracture orientations in stepovers (Soliva et al., 2010), and the origin of bends in faults (Martel, 1999).

LEFM models have identified one contributor to the development of damage zones, the regions of structures indicative of non-elastic deformation that surround many faults. Calculations and numerical models indicate that stresses in the vicinity of the tips of propagating fractures regularly exceed the elastic limit of rock, giving rise to a process zone, i.e., regions of nonlinear material behavior that develop in front of the propagating tips of fractures or faults (Blenkinsop and Drury, 1988; Cowie and Scholz, 1992b; Martel, 1997; Vermilye and Scholz, 1998). LEFM model predictions of the relative sizes of process zones, and the stress levels within them, compare favorably with microstructural evidence for plastic deformation near fault tips for relatively small faults. Further, predictions of the role of plastic deformation and fracture linkage inform our understanding of fault propagation (e.g., Cowie and Shipton, 1998). Thus, process zone deformation contributes to the development of damage zones.

A number of workers have documented structures other than fractures in the damage zones of faults in relatively weak, high porosity geological materials. Cataclastic shear deformation bands were first documented by Aydin (1978) and Aydin and Johnson (1978) in eolian sandstones; they have been subsequently described in poorly lithified sediments (Heynekamp et al., 1999; Cashman and Cashman, 2000) and non-welded ignimbrites (Wilson et al., 2003b). Fossen et al. (2007) review types of deformation bands (including compaction and dilation bands), and Davis (1999) documents the geometries of deformation bands in fault zones on the Colorado Plateau. Cataclasis in deformation bands in sandstone (Aydin, 1978) and poorly lithified sediments (Cashman and Cashman, 2000; Rawling and Goodwin, 2003) has received attention because it reduces permeability (e.g., Antonellini et al., 1994). Other deformation mechanisms, such as particulate flow, grain boundary sliding on clays, and chemical compaction (Rawling and Goodwin, 2003; Antonellini et al., 1994; Gibson, 1998, respectively) have been shown to operate instead of, or in concert with, cataclasis. Deformation bands can reduce permeability by two or more orders of magnitude with respect to the host rock, significantly impacting fluid flow (e.g., Jamison and Stearns, 1982; Antonellini and Aydin, 1994) (Fig. 6).

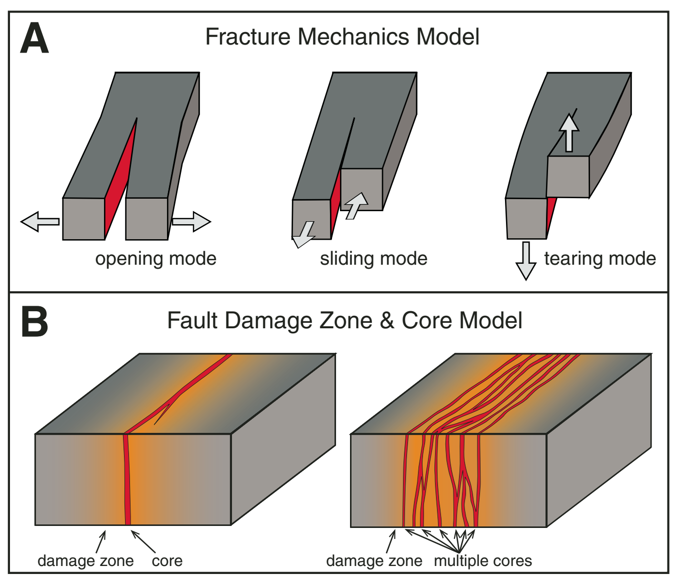


Figure 5. Fracture mechanics model versus core-and-damage zone model of fault geometry. Mode I (displacement parallel), mode II (displacement parallel to the fracture edge), and mode III (displacement perpendicular to the fracture edge) are the conceptualizations of fault geometry from fracture mechanics. In contrast, the core-and-damage zone model suggests that a central area(s) of the core is surrounded by a damage zone of increased fracturing. The displacement gradient associated with these faults is most abrupt in the core and decreases outward into the fracture zone.

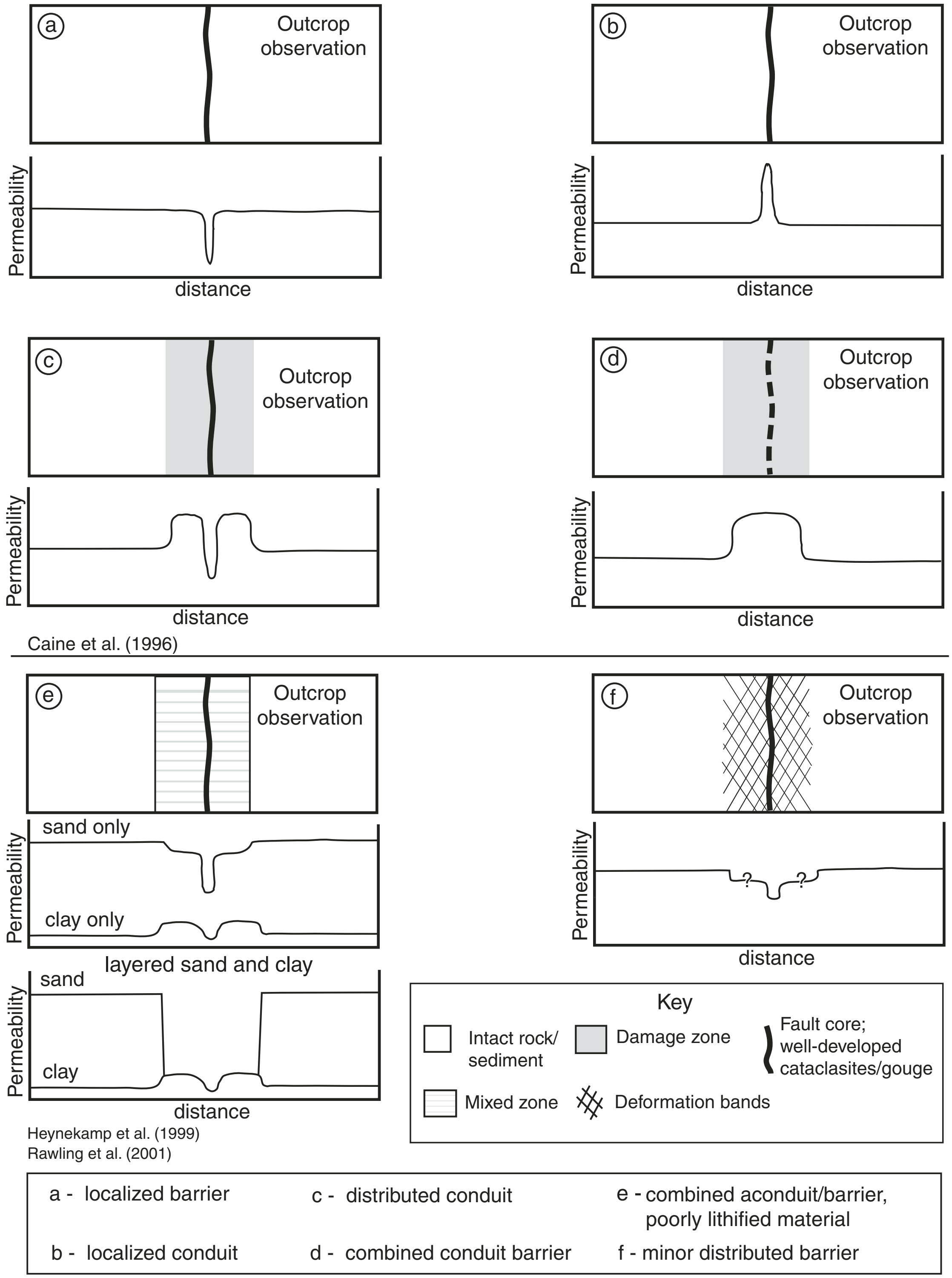


Figure 6. Fluid flow in faults, using the models of Caine et al. (1996). (A–D) Heynekamp et al. (1999) and Rawling et al. (2001). (E) Results from permeability analyses associated with the fault architecture. (A) and (B) reflect the opposite cases, where the fault acts as a conduit or a barrier for fluid flow (localized conduit and localized barrier, respectively). (C) Combined conduit-barrier, in which the damage zone has increased permeability, and the fault acts as a barrier. (D) Distributed conduit, in which the damage zone has increased permeability, and the fault has no effect. (E) Permeability within a fault zone in unconsolidated and layered sand and clay, which become increasingly mixed as the fault is approached. The mixed sand and clay of the damage zone ("mixed zone") increases permeability slightly relative to the clay, but lowers it dramatically relative to the sandstone. The fault core has very low permeability, similar to the undeformed clay layer. (F) In areas where the damage zone is dominated by deformation bands (porous starting material, e.g., sandstone), permeability is likely to decrease in the damage zone and will decrease within the fault core.

Approaching faults from the point of view of permeability, a conceptual fault zone model consisting of a damage zone surrounding a fault core (the portion of a fault zone where most of the relative movement occurs) has become widespread (Fig. 5B) (e.g., Chester et al., 1993; Caine et al., 1996). New insights from the model are (1) where damage zones are dominated by fractures, fluid flow is localized in the damage zone (Fig. 5B) rather than in the fault core, which can be impermeable owing to fine-grained fault products, precipitation of secondary minerals, and/or significant alteration of minerals; and (2) fault zone permeability is typically anisotropic and sensitive to the detailed structure of the fault zone (e.g., Odling et al., 2004). Combining this information with the work of Rawling et al. (2001) and Balsamo et al. (2010), one can arrive at an integrative picture of how fluid flow potentially moves through fault zones (Fig. 6). These models of fault zone architecture and permeability structure have guided a number of detailed studies of fault zone development, including both static and dynamic effects (Kim et al., 2004; Mitchell and Faulkner, 2009; Savage and Cooke, 2010). The damage zone and fault core model of fault zones share some geometric and structural characteristics with the LEFM model with a process zone forming at the tip of a propagating fracture. These two approaches are quite different, however. The LEFM conceptualization treats faults as planar cracks in elastic media (Fig. 5), whereas the damage zone–fault core models require no assumptions a priori about fault surface shape or rock properties. Still, some versions of the damage zone–fault core model also rest on generalizations, and will benefit from studies aimed at refining the model and applying it to specific examples (Wibberley and Shimamoto, 2003; Billi et al., 2003; Faulkner et al., 2010).

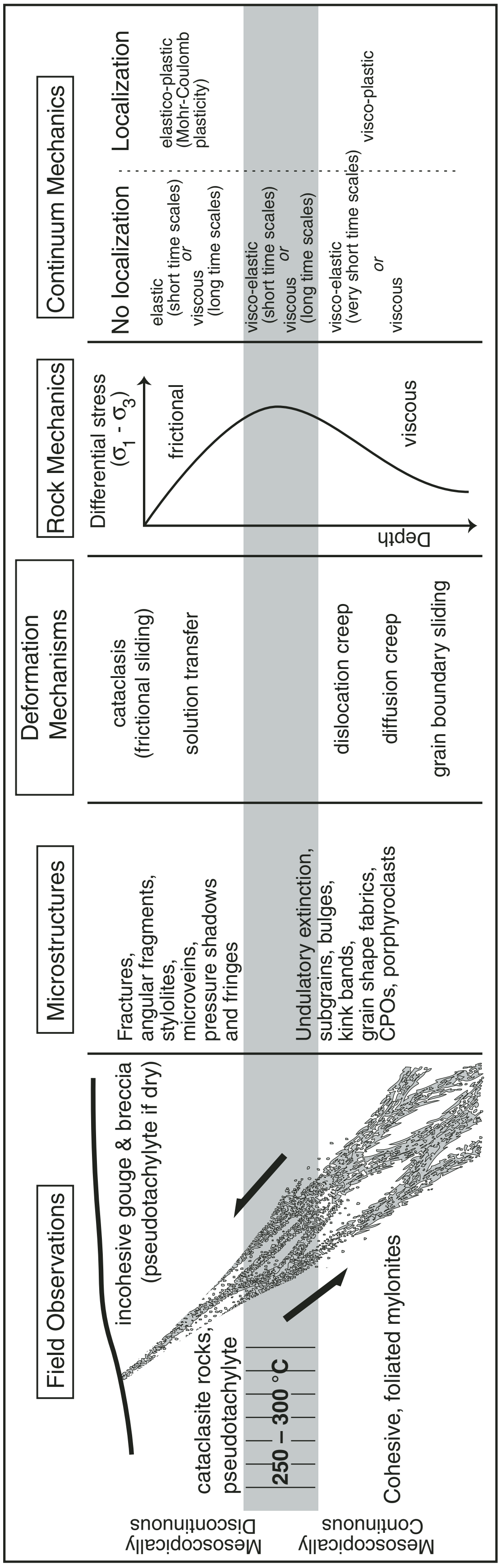
We address the microstructures recognized in fault rocks in greater detail in the Rheology section below, but we mention here studies that compare microstructures in different parts of fault zones. Engelder (1974) and Anders and Wiltschko (1994) demonstrated that microfracture densities increase toward the fault and that microstructures have orientations consistent with the stress concentrations about the tip of a fault zone. These findings were used to test alternative hypotheses for the formation of damage zones, such as their origin as process zones or around asperities (Wilson et al., 2003a; Mitchell and Faulkner, 2009). Microstructural studies have also revealed the significance of solution transfer (pressure solution) in fault zones (Hadizadeh et al., 2012) and have drawn attention to the importance of fabrics in cataclastic rocks (Chester and Logan, 1987).

The Fault Zone–Shear Zone Connection

There are a number of parallels between the way in which research in fault and shear zones has evolved. In the early 1970s, shear zone geometry was described in terms of displacement fields and strain: This step was taken for faults in the 1980s, when fault zones were related to strain (Jamison, 1989; Wojtal, 1989; Marrett and Allmendinger, 1991), and the first detailed data were collected on fault displacement fields for individual faults and for populations of faults. It is also interesting to note that determining the orientation of stresses—an obsession of many geoscientists in regard to fault zones—is not typically reported for shear zones. This connection is not made even though the orientation of the principal infinitesimal strain axes is often well constrained by vorticity studies, which would define the principal stress axes if the material was approximated as isotropic.

The connection between faults and shear zones was reinforced in one of the most important conceptual diagrams of the past 50 years (Fig. 7 [from Sibson, 1977]; see also Ramsay, 1979), which proposed an idealized model of a major fault zone. This visionary paper also introduced the basis for classification of fault rocks (sensu lato), considered the mechanics and energy budget of faults, discussed seismic versus aseismic faulting modes, and elaborated on deformation mechanisms in faults using a variety of inductive and deductive approaches. A paper as wide-ranging and generalizing as this one would be very hard to publish today.

Another parallel between shear zones and faults is the use of displacement-thickness diagrams for both features. These relations were plotted for shear zones by Mitra (1979) and for fault zones by Robertson (1983) and Scholz (1987). Power et al. (1988) and Power and Tullis (1991) measured the roughness of fault surfaces, and argued that the fractal nature of fault surfaces conformed with Scholz's (1987) hypothesis that the fault gouge thickness-displacement relationship derived from the frictional wear of asperities on rough fault surfaces. The displacement-thickness relation may apply to shear zones (Hull, 1988), but applying it to fault zones has been especially controversial (Evans, 1990). Using a space-for-time substitution, compilations of fault thickness versus length data and direct observations have been used to argue for widening of fault zones with time (Watterson et al., 1998; Hull, 1988), but this approach has been criticized because fault zone widths are highly variable locally, even along one fault (Woodward et al., 1988; Blenkinsop, 1989; Newman and Mitra, 1993). Mitra (1984) and Wojtal and Mitra (1986, 1988) showed that the thickness of the active portion of the fault zone changes over time due to changes in the dominant deformation mechanism and, for large displacement faults, with changes in the ambient conditions at which deformation occurs. Childs et al. (1996) described bifurcations of faults along lines parallel to the propagation direction or at asperities; either situation could lead to considerable complexity in the evolution of fault zone thickness with time. The roughness of large faults is less than small faults (e.g., Candela et al., 2012): Applying the space-for-time substitution, this implies that faults become smoother with time and slip events (Sagy et al., 2007). Further, Shipton et al. (2006) suggest that different scaling relationships should be used for different components of the fault zone (e.g., damage zone, fault core). In an analogous manner, deformation bands exhibit displacement-length relationships, although different relationships apparently hold depending on the dominant deformation mechanism (e.g., Fossen et al., 2007).



A clear distinction between shear zones and fault zones makes increasingly less sense from a microstructural-deformation mechanism perspective. Originally, it was conceived that fault zones were controlled nearly exclusively by cataclastic deformation mechanisms, and shear zones were controlled dominantly by intracrystalline plasticity deformation mechanisms (plus a contribution from diffusive mass transfer). The recognition of foliated cataclasites suggested that deformation mechanisms besides cataclasis (perhaps dislocation glide within micas) occur within fault zones (e.g., Chester et al., 1985). In addition, identification of the importance of solution transfer of quartz and calcite in fine-grained fault rocks (Mitra, 1984; Wojtal and Mitra, 1986, 1988; Kennedy and Logan, 1997, 1998) led to the hypothesis that resistance to fault slip would decrease and fault zones would narrow as slip accrued. The operation of solution transfer in faults is comparable to Coble creep in higher temperature shear zones (Elliott, 1973), indicating that diffusional mass transfer is significant in both shear zones and fault zones. Kennedy and White (2001) documented dislocation glide and low-temperature dynamic recrystallization in vein calcite that originated by solution transfer, further blurring the distinction between fault zones and shear zones. While the significance of metamorphic reactions in shear zones was recognized in the 1970s (e.g., Mitra, 1978; White and Knipe, 1978), the documentation of low-temperature mineralogical reactions in thrust fault zones (Vrolijk and van der Pluijm, 1999) and the SAFOD drill hole (Lockner et al., 2011) indicates that phase transformations also occur within fault zones. Further, metamorphic phase transformations may exert an important control on pore fluid pressure, leading to the occurrence of localized slip (i.e., faulting) in shear zones (Axen et al., 2001).

Finally, it is unlikely that the strain rate is constant in any deformation zone over time. It has long been recognized that strain rate transients are likely to occur at all lithospheric levels (e.g., Wise et al., 1984; Knipe, 1989); direct evidence is now being observed on active faults (e.g., Shelly, 2010). Other evidence for variations in strain rate include (1) transient earthquakes (with presumably cataclastic deformation mechanisms that may occur at elevated pressure and temperatures (e.g., Hobbs et al.,

Figure 7. Rheology of crustal faults in quartz-rich rocks, in the context of the Sibson (1977) diagram, as described by different approaches to rheology. Field studies differentiate between structural features that are mesoscopically discontinuous or continuous, while also noting displacements and the nature of the geological structures. Microstructural studies distinguish between observed microscopic features, some of which are found at different crustal levels. Deformation mechanisms are dominantly cataclastic and solutional (solution transfer) in the upper crust, and dislocation creep and diffusional creep dominate in the lower crust. The rock mechanics studies would distinguish a frictional (Byerlee's) upper crust and a viscous (quartz, power-law flow) lower crust. Modeling studies might distinguish an elasto-plastic upper crust from a visco-plastic lower crust. An important point is that these are simplifications that apply to a narrow range of strain rates. CPOs—crystallographic preferred orientations.

1986), and (2) the alternation of the dominance of diffusional mass transfer and cataclastic deformation mechanisms in fault zones (e.g., Gratier and Gamond, 1990; Hadizadeh et al., 2012); dominance is used because diffusional processes are active even at low temperature. Consequently, it is likely that tools used to study shear zones and those used to study faults may increasingly overlap in the future as the division between the two archetypal deformation zones diminishes.

Fold-and-Thrust Belts: Fault Systems

Elucidation of the geometry of thrust faults and the evolution of fold-and-thrust belts is one of the major accomplishments of structural geology in the past 50 years (Fig. 8). In addition, the work on thrust faults completed in the 1960s to 1980s was the first thorough investigation of *systems* of faults that accommodate regional shortening, thickening, and elongation of large masses of rock. Echer von der Linth first mapped thrust faults in the Alps in the mid-nineteenth century, and geologists working in other belts were profoundly influenced by his work and subsequent contributions by Alpine geologists such as Heim, Bertrand, Buxdorf, and Argand (see Bailey, 1935, for an engaging review of the discovery of thrust faults). Surprisingly modern work on fold-and-thrust belts appeared at the beginning of the 1900s, with influential studies on the Scottish Highlands (e.g., Peach et al., 1907) and the Appalachians (Chamberlin, 1910). More relevant to the time frame of this contribution, significant advances in understanding the geometry of thrusting came from the integration of seismic studies in the Canadian Cordillera (e.g., Bally et al., 1966; Dahlstrom, 1970; Price and Mountjoy, 1970), which constrained subsurface geometry, and sedimentological studies in the North American Cordillera, and also the evolution to "in-sequence" or "break-forward" thrust faulting (Oriel and Armstrong, 1966; Royse et al., 1975; Jordan 1981). Boyer and Elliott (1982) and Butler (1982) compiled and rationalized existing definitions for the geometry of individual faults (i.e., fault terminations or tip lines), provided tools for understanding the patterns exhibited by fault arrays (i.e., splays, branch lines, imbricates, and duplexes), and outlined models for the temporal development of fault systems (i.e., duplexes vs. truncated imbricate fans). In addition, they extended the discussion of slip distributions on fault surfaces begun by Elliott (1976: the bow-and-arrow rule) and used an analysis of distribution of displacement among faults in a system to understand the relationships between folded thrusts and windows or culminations, antiformal stacks, and downward-facing structures in foreland-dipping duplexes. Although these models were conceptually simple (they did not incorporate studies of deformation mechanisms or rock rheology), they were extremely useful in interpreting the kinematics of fold and thrust belts. Another highly applicable approach was J. Suppe's analysis of the geometry of the folds associated with thrust faults (Suppe, 1983; Suppe and Medwedeff, 1990), which provided a foundation for analyses of fault-bend and fault-termination folds.

Understanding the geometry of these systems enabled structural geologists to develop cross sections that were both kinematically accurate (consistent with the data) and admissible (geometrically possible). Bally et al. (1966) and Dahlstrom (1969) used the conceptual model of hanging wall deformation above a step-shaped décollement (Rich, 1934) to restore the sedimentary layers within the thrust sheets to an originally horizontal position using their bed length. Increasing sophistication of these methods includes the use of multiple restored sections (Affolter and Gapais, 2004; Affolter et al., 2008) and area-depth-strain restorations (Groshong et al., 2012).

All of the above work attempts to reconstruct the movement (or translational component) of thrust sheets, while not addressing either the internal strain or rotational components of the displacement field. Workers have recognized the internal deformation that occurs within deeper thrust sheets, such as the Moine thrust, and attempted to interpret these signals in terms of simple kinematic models (Coward and Kim, 1981; Sanderson, 1982). Large rotations within individual thrust sheets have been detected using paleomagnetic studies (e.g., McCaig and McClelland, 1992); paleomagnetism also revealed vertical axis rotations on an orogenic scale in fold-and-thrust belts (Ries et al., 1980). More recent work has integrated translation, internal deformation, and rotation in an attempt to understand fully the three-dimensional deformation associated with thrust systems (Fig. 8; Weil et al., 2010).

The timing of foreland fault movements was also evaluated by structural geologists using synsedimentary deposition. Work in northeast Utah and southwest Wyoming used syntectonic deposition to reconstruct the evolution of foreland deformation (DeCelles and Mitra, 1995). In the Spanish Pyrenees, Holl and Anastasio (1993) used synsedimentary deposits to constrain the timing of rotation and thrust fault movement. By integrating paleomagnetic studies with sedimentary constraints on timing, highly detailed reconstructions of rotation rates can be made (Mochales et al., 2012).

A recent development is a broader recognition of the role of salt in fault systems. Earlier studies focused on understanding diapiric salt structures (e.g., Jackson and Talbot, 1986, 1989), with some workers recognizing the need to account for large-scale movement of salt masses, in some cases entirely out of cross-section, in regional-scale reconstructions (e.g., Worrall and Snelson, 1989). Recent compilations document the importance of plugs and allochthonous sheets of salt in the development of fold-and-thrust structures in some settings (Hudec and Jackson, 2006, 2007).

In addition to the improved knowledge of the structure and evolution of thrust belts, there were two theoretical breakthroughs, one kinematic and one dynamic, that had major impact on the interpretation of fold-and-thrust belts. Erslev (1991) proposed trishear, a kinematic model which combined translation of thrust sheets with internal deformation to make accurate predictions of geological structures involved in thrust movement (also Allmendinger, 1998). The trishear model suggests that a

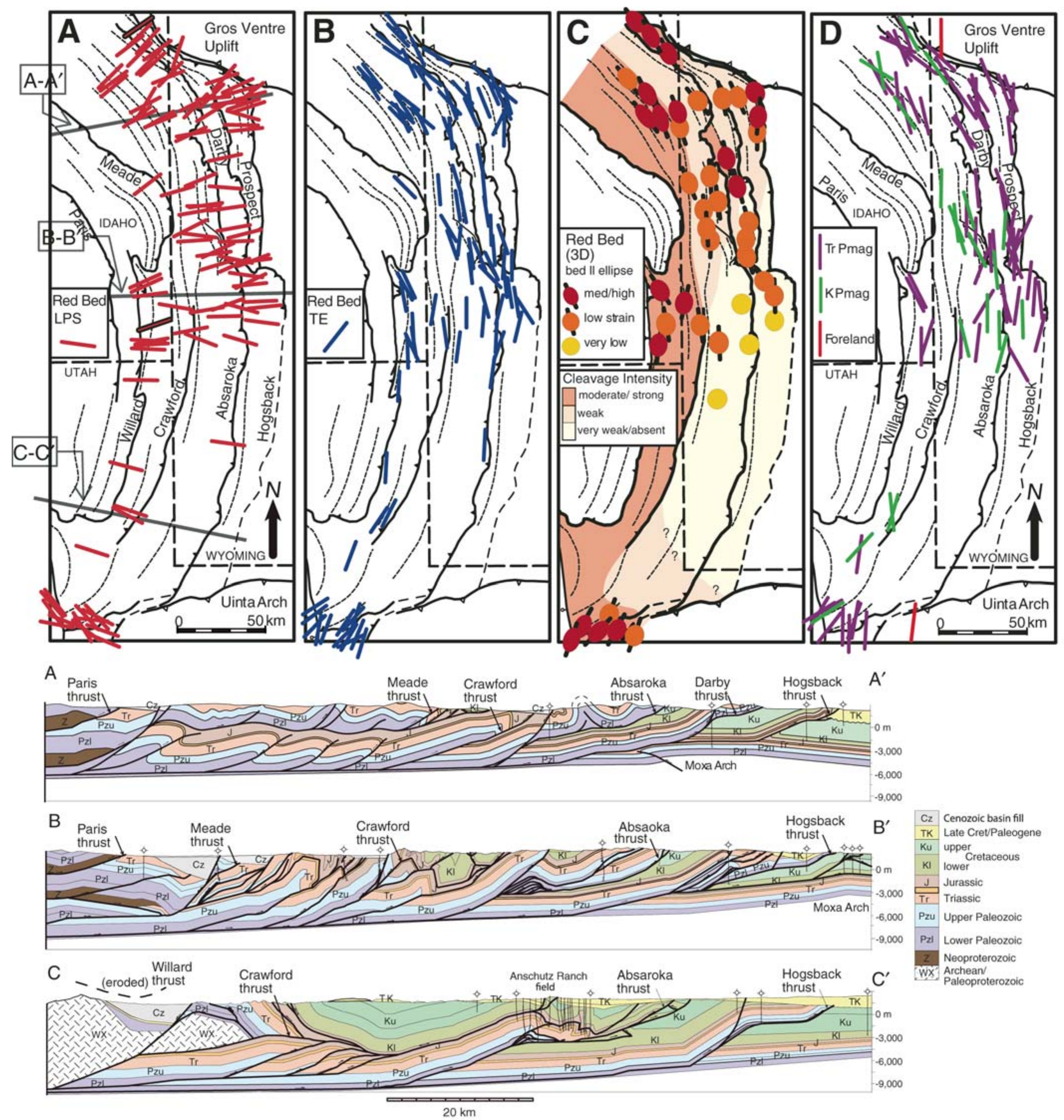


Figure 8. From Weil et al. (2010), Yonkee and Weil (2010), and Yonkee and Weil (2011). Four panels from the Utah-Idaho-Wyoming thrust belt, showing (A) layer parallel shortening (LPS); (B) tangential extension (TE); (C) finite strain measurements (shown in plan view as ellipses); (D) paleomagnetic declinations relative to North of both Triassic and Cretaceous aged magnetizations. LPE and TE were measured using cleavage, fracture, and vein networks, plus minor folds and minor faults. These data constrain the internal strain and rotation component of displacement, while balanced cross sections constrain the translational component of deformation. (E) Three cross sections through the Utah-Idaho-Wyoming thrust belt. Cross section locations are highlighted in frame (A).

triangle-shaped zone of deformation extends outward from the tip of an active thrust fault. The trishear model was particularly good at predicting the timing and geometry of folding and minor faulting associated with reverse fault motion. Cristallini et al. (2004) recently expanded trishear analysis into three dimensions.

The second major breakthrough was the critical taper theory, which addressed the significant question of what drives the emplacement of thrust sheets, masses of rock that are several kilometers to tens of kilometers thick, several tens of kilometers wide, and hundreds of kilometers long, that move laterally great distances while also moving from structurally lower to higher positions. The most widely pursued approach built upon the work of Hubbert and Rubey (1959), which (1) analyzed the motion of a thrust sheet driven by a push from behind and accommodated by Coulomb sliding and (2) concluded that significant pore fluid pressures are required to enable thrust sheets of the sizes regularly found in orogenic belts to move. The key modification of this approach is the application of critical taper modeling to the foreland fold-thrust belts (Davis et al., 1983; Stockmal, 1983; Emerman and Turcotte, 1983; Platt, 2000; Simpson, 2011). Critical taper theory proved to be highly successful because it linked the orogenic "push from behind" with a second major causal mechanism, a forward-dipping upper surface, to account for the observable deformation and fault-fold geometries in fold-thrust belts. Chapple (1978) proposed that a perfectly plastic mass of rock (i.e., the thrust wedge) sitting on a linearly viscous décollement would acquire a critical taper during deformation. The more widely adopted tapered-wedge model envisioned that the interior of the wedge exhibits Coulomb plasticity and deforms internally until the wedge attains a particular combination of surface slope and basal décollement dip, termed the *critical taper value* (Davis et al., 1983; Dahlen, 1984, 1990; Dahlen et al., 1984; Dahlen, 1990). The critical taper value is dependent upon the pore fluid ratio within the deforming wedge. Once that wedge meets this criterion, it is able to slide stably across the basal décollement. Davis et al. (1983) used this model to examine the evolution of western Taiwan and a number of other subaqueous and subaerial orogenic wedges. Several authors (Woodward, 1987; Boyer, 1995; DeCelles and Mitra, 1995) analyzed the geological implications of critical wedge theory, in terms of the relative ages of faults in a deforming wedge, the effect of the original taper of the sedimentary wedge, and the effects of regional erosion and deposition patterns on wedge development.

Fault Systems in Other Tectonic Settings

Many of the techniques (balancing, restoration, mechanical modeling, etc.) that were initially developed for thrust fault systems were sufficiently successful to have been subsequently applied to other tectonic environments (e.g., extensional and wrench settings). Fault restoration, for example, was applied to normal fault systems, in conjunction with seismic data associated with petroleum exploration (Gibbs, 1984). Quantification of this work continued with Gratier et al. (1991) and Rouby et al. (1996, 2000).

Following the initial insight that low-angle normal faults existed (Proffett, 1977), it was recognized that systems of normal faults can occur associated with major detachment surfaces (e.g., Wernicke, 1981; Wernicke and Burchfiel, 1982). This insight led to the discovery of metamorphic core complexes, exhumed beneath major normal detachments (e.g., Davis, 1980). Further, the detachment model afforded an alternative model for pervasive mid- to lower-crustal deformation below upper-crustal faults. Interestingly, it appears that both the detachment and lower-crustal flow models are applicable to the Basin and Range Province of the United States, albeit at different times. Core complexes were subsequently recognized in many orogens around the world (e.g., Lister et al., 1984; Gibson et al., 1988; Dalziel and Brown, 1989; Verdel et al., 2007), in Archean belts (Kisters et al., 2003), and even in mid-ocean ridge settings (Cann et al., 1997; Tucholke et al., 1998). Some early studies of normal detachments and core complexes took an essentially historical approach, a theme that continued with efforts to constrain the exhumation path of core complexes through innovative and integrated low-temperature geochronology (e.g., Carter et al., 2004; Foster et al., 2010). More process-oriented studies looked at deformation mechanisms, rheology, and mechanics (e.g., Davis, 1983; Wickham et al., 1993, Rosenbaum et al., 2005), and the involvement of melt in the extensional process (e.g., Vanderhaeghe and Teyssier, 2001b; Rey et al., 2009). Low-angle detachments were imaged on mid-ocean ridges (Tucholke et al., 1998) and were hypothesized on Venus on a larger scale than occur in terrestrial settings (Spencer, 2001).

Coincident with the burgeoning core complex studies and the discovery of low-angle detachments, a debate started about the geometry of major crustal normal faults at depth (e.g., Jackson and McKenzie, 1983). Models with deep crustal listric geometries were rapidly adopted for core complexes (e.g., Lister and Davis, 1989) and rift-bounding normal faults (e.g., Bosworth, 1987; Morley, 1989), and this concept was extended to passive margins (Lister et al., 1986). However, a lack of seismicity on gently dipping normal faults proved a challenge at first, and a case for relatively planar normal faults to the base of the crust was argued (Jackson and White, 1989), partly based on focal mechanism studies (cf. Collettini and Sibson, 2001). Increasingly plausible cases have been put forward for seismicity on detachments (Rietbrock et al., 1996), but the case for the existence of crustal-penetrating, low-angle normal faults at passive margins has not been made convincingly.

The mechanics of listric and low-angle normal faults present another challenge for Andersonian fault mechanics, which suggests that low-angle normal faults are in a highly unfavorable orientation for slip. Several solutions to this paradox have been suggested. Detachment faulting may occur on very shallowly dipping or horizontal surfaces or zones because of changes in rheology within the crust (Wernicke, 1981; Lister and Davis, 1989), specifically the change from "brittle" to "ductile" behavior between the upper and lower crust. In support of a low-angle origin for normal faults, paleotemperature estimates for rocks exposed in the footwalls of the faults show that the initial dips of the normal faults were low (e.g., Foster and John, 1999). Furthermore, the presence of pseudotachylite on these faults demonstrates that they were seismically active. One reason for initial formation of low-angle normal faults could be the perturbation of an Andersonian stress by intrusions (Parsons and Thompson, 1993; Morley, 1999). However, alternative geological arguments have been advanced that low-angle normal faults did not form with low dip angles. These include the observations that bedding is cut off by the faults at high angles (inconsistent with the faults being initially at low angles and therefore subparallel to bedding), that metamorphic gradients in the hanging wall are only consistent with a high-angle fault, and that large gradients in paleotemperature of the footwall imply that the footwalls have been tilted (e.g., Wong and Gans, 2003). Hence low-angle normal faults may have formed at conventional high angles and rotated into their present low-angle orientation (e.g., Buck, 1988). Such fault rotation is seen in the flexural cantilever model for normal faults (Kusznir and Ziegler, 1992), and in models where flow of the lower crust from unloading leads to rotation of the faults (Yin, 1989; Gessner et al., 2007). At large extensional strains, some models predict a "rolling hinge," where a high-angle fault rotates to a low angle near the surface (e.g., Axen et al., 1995; Lavier et al., 1999), and field evidence has been advanced to support this model (e.g., Fletcher and Spelz, 2009). Both inductive and deductive evidence has been put forward for and against the formation of normal faults with low dip angles. The contradictions between the various types of empirical evidence could suggest that there are at least some low-angle faults that form in this orientation, while others are rotated. It is unlikely that a single model or set of observations applies to all situations.

Studies of strike-slip fault systems in the last 50 years shared an initial similarity with their dip-slip counterparts in focusing on geometry. The Riedel model for R, R′, and Y shears and T fractures in strike-slip fault zones was highly influential (Tchalenko, 1970) and widely—possibly too widely—applied. Three-dimensional aspects of strike-slip faults were investigated through sandbox studies (Naylor et al., 1986), leading to the appreciation that Riedel fractures could have helicoidal geometries; these geometries are inferred in neotectonics settings (Carne and Little, 2012). Field studies also revealed changing dips of strike-slip faults near the surface, encapsulated by terms such as *flower, palm tree,* and *tulip structure* (Wilcox et al., 1973; Sylvester and Smith, 1976; Naylor et al., 1986). The diverging upward geometry of flower structures was taken as diagnostic of strike-slip faults, especially for seismic interpretation. Alternatively, inversion of normal faults can lead to the same geometry, which led to an interesting controversy about the role of strike-slip tectonism versus inversion tectonics in South America (Amilibia et al., 2008). Non-planar fault geometries result in vertical movements associated with the dominant horizontal displacements of a strike-slip fault zone, and pull-apart basins are another consequence of non-planar fault geometries (e.g., Crowell, 1982; Aydin and Nur, 1982; Sylvester, 1988).

The concept of duplexes, first appreciated for thrust systems, was applied to strike-slip systems (e.g., Woodcock and Fischer, 1986; Swanson, 1988). A surprising discovery was the existence of vertical axis rotations revealed by paleomagnetism in the San Andreas fault zone of southern California (Luyendyk et al., 1980), accommodated by movement on antithetic faults. The block rotation model was conceived for crustal scale blocks, but the concept of rotations accommodated by fault systems is more general and could apply at much smaller scales (Gapais et al., 2000). Dewey et al. (1998), for example, considers that fault systems pervasively "shatter" regions undergoing transtensional deformation, resulting in many independent blocks (and adjacent faults) that undergo local rotation. Rotation of faults is, in fact, an inevitable consequence of simple shear accommodated by slip on antithetic faults (Nur et al., 1986), and it is an integral aspect of Cosserat continuum modeling (Žalohar and Vrabek, 2010). The question of how much a fault can rotate and remain active in a strike-slip system has interesting similarities with the problem of slip on low-angle faults in normal fault systems. Scholz et al. (2010) argued that a new fault will form as strike slip faults rotate when a stress state is reached that will no longer reactivate the rotated fault.

Intrusions and Migmatites

The generation and emplacement of igneous rocks into the Earth's crust has always been of importance to structural geology and is a source of ongoing research. Intrusions and migmatites (i.e., partially molten rocks and magmas) are included in this contribution because of fundamental advances made in a number of key areas: the geometry of intrusions and melt migration networks, magma ascent and emplacement mechanisms, and flow processes in intrusions and migmatites. Integration of structural and petrologic studies, aided by increasingly sophisticated geochemical and geochronological methods, has also led to significant advances in our understanding of the petrogenesis of partially molten rocks and magmas, which are beyond the scope of this chapter (see reviews by Brown, 2001b; Sawyer and Brown, 2008; and Vanderhaeghe, 2009).

Gilbert (1877) was probably one of the first to recognize on his expeditions to the Henry Mountains of southern Utah that intrusions exist in the shallow crust and that magmas could deform and uplift their wall rocks. This prescient work is emblematic of key questions that the structural geology community would tackle in earnest over the last few decades, including (1) the ongoing question of how crustal material has been displaced to allow for magma emplacement, i.e., the "space problem" debate, which has focused on how tectonic structures aid the space-making process versus how much of the space the magma can make for itself (forceful emplacement); (2) the processes by which melt segregates from its source region through ascent conduits to the site of emplacement, manifested as the "dikes versus diapirs" debate; and (3) interpreting the deformational history of plutons through analysis of strain-induced fabrics, rock microstructure, and thermal evolution.

Early work on plutons focused on internal fabrics, with well-known work in Germany by Hans Cloos and brought to North America (e.g., Balk, 1937). Pitcher and Berger's (1972) seminal work on the Donegal intrusions emphasized the connection between internal fabrics and structures within the wall rocks. They modeled the intrusions by a combination of syntectonic forceful emplacement of magma pulses and stoping of the wall rocks. These models are still widely used today, although the effectiveness of stoping as an emplacement mechanism remains controversial (e.g., Marsh, 1982; Clarke et al., 1998; Dumond et al., 2005; Glazner and Bartley, 2006). Observing the association between plutons and major regional tectonic structures, Hutton (1982), also working on the Main Donegal Granite, developed a model whereby space to accommodate pluton emplacement was created by a strain gradient across a shear zone on the margins of the pluton. To make crustal-scale "space," other authors proposed pull-apart models associated with transpressional tectonics (e.g., McCaffrey, 1992; Tikoff and Teyssier, 1992; Vigneresse, 1995). However, many plutons—even when associated with shear zones or faults—show evidence for deforming the surrounding rocks. Brun et al. (1990), for example, provide a compelling example of forceful emplacement of an intrusion associated with a major structure. Tommassi et al. (1994) pointed out that magmatism also facilitates tectonic movement, and vice versa, and that it may be impossible to separate interactions of shear zone motion and plutonism.

Studies of the internal structures within plutons continued (e.g., Castro, 1986), facilitated by microstructural observations (e.g., Vernon, 2000) and the distinction between magmatic and solid-state fabrics (Marre, 1986; Blumenfeld and Bouchez, 1988; Paterson et al., 1989). These studies were aided by the systematic application of the anisotropy of magnetic susceptibility (AMS) (e.g., Henry et al., 1988; Bouillin et al., 1993; Bouchez, 1997), which proved to be of tremendous advantage for capturing the subtle internal fabrics (particularly lineations) within intrusions (e.g., St. Blanquat and Tikoff, 1997; Launeau and Cruden, 1998). Several classic plutons with structures originally interpreted to have formed by forceful emplacement (e.g., the Ardara in Ireland, Akaad, 1956; Papoose Flat in California, USA, Sylvester et al., 1978) were re-interpreted to have formed as a result of regional deformation postdating or synchronous with emplacement (e.g., Paterson et al., 1991). This interpretation was supported by the observation that most fabrics in plutons appear to have formed late in the emplacement history (Paterson et al., 1998), the hypothesis that magma (liquid plus crystals) pressure is not great enough to produce solid-state and gneissic foliations (Vernon and Paterson, 1993), and calculations that the strain produced in the country rocks was insufficient to accommodate the volume of the plutons (Paterson and Fowler, 1993). These arguments were countered with new strain analyses (Molyneux and Hutton, 2000), by making room with floor subsidence (Cruden 1998; Cruden and McCaffrey, 2001), or by making room by upward and/or outward translation of the wall-roof rocks during ballooning, which resolves the space problem in certain cases (Ramsay, 1989; Morgan et al., 1998; Tikoff et al., 1999; Johnson et al., 2003). In other cases, however, the space problem still exists, especially for plutons in which there are no clear deflections or offsets of regional structures and minimal aureole strain (Bilodeau and Nelson, 1993; Paterson et al., 1996).

Experimental deformation studies have repeatedly shown that only a few percent of melt are needed to achieve connectivity of the melt phase at larger grain scales (e.g., Bulau et al., 1979; Jurewicz and Watson, 1984, 1985) and therefore to facilitate the segregation and migration of melt from the source into a permeable melt-migration network (Vigneresse et al., 1996; Vanderhaeghe, 1999; Sawyer, 2001; Brown, 2001a). Consequently, the mechanisms of melt transport in the crust from their zone of formation at depth to the upper crust where they are intruded as plutons has been the focus of vigorous debate and active research. Marsh (1982) and Mahon et al. (1988) calculated the thermal-mechanical-temporal requirements necessary for a magma diapir to soften the middle and upper crust to allow itself to rise (hot Stokes flow). Their analyses indicated that in order to soften the crust, the heat loss from the magma would rapidly lead to crystallization and locking up. These thermal constraints on diapirism led to a model based on transport of magma through fractures (Clemens and Mawer, 1992; Rubin, 1993). Work by Brown (1994), Weinberg (1996), and Weinberg and Searle (1998) further advocated for the role of pervasive magma flow and its likely importance for controlling whether diking or diapirism controls later magma ascent. Large diameter plutons, initially believed to be vertically extensive, were effectively modeled as tabular or sheet-like (Vigneresse, 1988), often with root zones occurring above localized vertical lineations within the intrusion (Guillet et al., 1985; Cruden et al., 1999). The latter was significant because it finally eliminated the concept that all upper-crustal plutons are diapirs. In fact, diapiric rise of magma is the exception, rather than the rule, for upper crustal emplacement of granitic magmas (e.g., Petford et al., 1993).

One interesting aspect of this outcome, however, was that the rate of tectonic movement was generally insufficient relative to the rates required for pluton emplacement (e.g., Paterson and Tobisch, 1992; Petford et al., 2000; St. Blanquat et al., 2011). The sheet-like or tabular geometry of many plutons (e.g., Hogan and Gilbert, 1995; Cruden, 1998) led some workers to model plutons as laccoliths (e.g., Morgan et al., 1998; Rocchi et al., 2002) and also to suggest that shapes of plutons/laccoliths were scale invariant (McCaffrey and Petford, 1997). These sheet-like geometries are well exposed in, once again, the Henry Mountains (e.g., Horsman et al., 2005; Morgan et al., 2008), but they are increasingly recognized—albeit cryptic—in larger plutons (Glazner et al., 2004). The geophysical evidence for tabular or sheet-like geometries further supported the increasingly strong evidence for pulsed-pluton construction (Pitcher, 1970; Hutton, 1982; Brown and McClelland, 2000), especially considering the geochronologic evidence for long-lived (between 5 and 8 Ma) intrusive suites (Coleman et al., 2004). Discrete pulses of magmas transported through dikes (Clemens and Mawer, 1992) aid the space problem and are more compatible with crustal strain rates and with evidence of long-lived magma chambers. In contrast, some large tabular intrusions have been emplaced within tens of thousands of years (Michel et al., 2008), consistent with more recently determined felsic to intermediate magma viscosities that are significantly lower than previously believed (Clemens and Petford, 1999). Given these lower viscosities and higher magma ascent rates, the rate-limiting step in granite production seems to be the longer time scales required for melt production (Petford et al., 2000).

A resurgence of petrological research on migmatites, showing that migmatites are a product of anatexis rather than solid state differentiation (cf. Brown, 1973), engendered renewed interest in the structural studies that aimed to understand the grain- to orogen-scale connections between melting and deformation processes. Since the influential work of Mehnert (1968), our understanding of the structure of migmatites has advanced significantly. Classical interpretations of migmatitic structure (e.g., Campbell, 1980; Hopgood, 1980) that emphasized overprinting and solid-state structural successions are now understood in the context of syndeformational melt flow and evolving mechanical anisotropy during progressive melting (e.g., Hasalová et al., 2008; Schulmann et al., 2009). The efforts of many researchers (e.g., Blumenfeld and Bouchez, 1988; Vernon and Collins, 1988; Brown and Solar, 1998a 1998b, 1999; Sawyer, 2001; Marchildon and Brown, 2003; Holness, 2008) have identified microstructures in migmatites indicative of partial melting (e.g., phenocrysts of feldspar with interstitial quartz, myrmekitic intergrowths of quartz and plagioclase, geometrically distinct crystal facets and aggregates) or melt-rock reaction (e.g., symplectites, coronas, replacement textures). Rock deformation studies, such as those of Jurewicz and Watson (1985), further established criteria for recognizing the former presence of melt in migmatites by comparison of experimentally formed microstructures and documentation of the geometries of melt-solid interfaces at the grain scale (e.g., van der Molen and Paterson, 1979; Dell'Angelo and Tullis, 1988; Rutter and Neumann, 1995; Rosenberg and Handy, 2005).

Studies of the spatial distributions of melt-bearing structures in migmatites at mesoscopic and microstructural scales (e.g., Sawyer, 2001; Marchildon and Brown, 2002) document synmigmatitic layering and patterns of leucosome distribution developed within dilatancy structures (e.g., extension fractures, shear bands, boudin necks) that are thought to be the signature of pervasive melt flow, compaction, and dilatancy-pumping during syntectonic deformation (e.g., Brown, 1994; Collins and Sawyer, 1996; Vanderhaeghe, 1999). Indeed, grain-boundary flow, melt migration along structural fabrics, intergrain or intragrain tensile microcracking, and pervasive flow have been demonstrated as viable melt transfer mechanisms in migmatites (e.g., Brown and Solar, 1999; Weinberg, 1999; Sawyer, 2001; Marchildon and Brown, 2002; Hasalová et al., 2008; Weinberg and Mark, 2008). Accordingly, modern classifications of migmatites correlate the continuity of the solid gneissic framework within migmatites and the inferred former melt fraction with the transition between partially molten rocks (i.e., metatexites characterized by a continuous solid framework) and magmas (i.e., diatexites characterized by melt with solids and/or crystals in suspension) (cf. Brown, 1973; Wickham, 1987; Burg and Vanderhaeghe, 1993; Vanderhaeghe, 2001, 2009, and references therein). In recent years, the anisotropy of magnetic susceptibility has proven to be a useful technique in the structural analysis of migmatites, capable of recovering planar and linear fabrics developed during flow (Ferré et al., 2003, 2004; Kruckenberg et al., 2010) and thereby aiding tectonic studies of partially molten terranes (Teyssier et al., 2005; Charles et al., 2009; Kruckenberg et al., 2011).

Feedbacks between deformation and melting in orogenic belts have been an area of active research throughout the past few decades (e.g., shear zones, convergent orogens: Brown and Solar, 1998a, 1998b) and remain a vibrant (and heavily debated) area of research today. Motivations to study migmatite-granite systems include understanding the processes of melt migration and deformation partitioning in the lithosphere (e.g., McKenzie et al., 2000), channel flow (Block and Royden, 1990; Wdowinski and Axen, 1992; Clark and Royden, 2000; Beaumont et al., 2001), orogenic collapse (e.g., Vanderhaeghe and Teyssier, 2001a, 2001b), and gneiss dome formation (e.g., Teyssier and Whitney, 2002).

APPLICATION OF GEOCHRONOLOGICAL TECHNIQUES IN STRUCTURAL GEOLOGY STUDIES

For the structural geologist, time is a critical parameter for both historical- and process-based approaches. Accordingly, the integration of geochronologic techniques and structural geology studies has been one of the most important and profound changes in the field. In the last few decades there has been a proliferation of new geochronological techniques that are used to study deformation over a large range of temperatures (Fig. 9). Prior to the advent of geochronology, most age constraints were based on stratigraphic relations. While useful for determining relative chronologies, the introduction of isotopic methods allowed absolute ages to bracket specific events. This advance provided the structural geologist with critical information on the timing of deformational, magmatic, or metamorphic events. It also supplied the ability to estimate the rates and durations of tectonic processes, albeit with some limitations as to the accuracy of ages imposed by the large sample volumes needed for analysis. Geochronologic methods now routinely allow geologists to date individual accessory mineral phases and—with the development of in situ dating methods—domains within minerals.

The subject of geochronology is addressed in detail by J. Mattinson in this volume (Mattinson, 2013). Consequently, we highlight only a handful of examples that illustrate new ways in which geochronological techniques are advancing the field of modern structural geology through the direct dating of geological structures (e.g., faults) and characterization of the duration and/or rates of tectonic processes (e.g., strain rates). $^{40}Ar/^{39}Ar$ dating of fault gouge clay (e.g., Vrolijk and van der Pluijm, 1999)

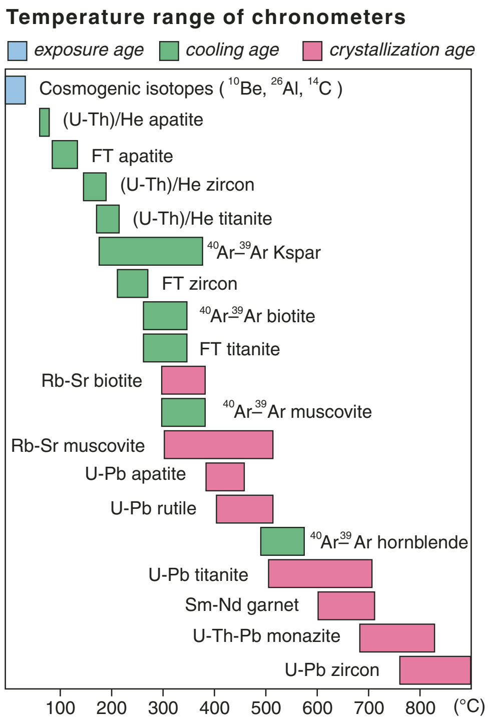


Figure 9. A list of temperature ranges of chronometers currently used for structural geology studies.

is one example of a new geochronologic technique available for structural geologists. The key to this approach is that illite grows during deformation. The Ar signal from the newly formed illite is distinguished from detrital illite by quantifying illite polytypes (2M1 vs. 1Md, detrital and authigenic, respectively) in different size fractions with X-ray diffraction techniques (van der Pluijm et al., 2001; Haines and van der Pluijm, 2008), and by dating each of these fractions with the ^{40}Ar/^{39}Ar illite vacuum-encapsulated dating method (Dong et al., 1995). Faults dated in this way have been used to constrain the age of regional faulting (van der Pluijm et al., 2001), infer the age of early orogenic fault movements (Duvall et al., 2011), and estimate strain rates on faults (Haines and van der Pluijm, 2008).

Determining strain rates for geological structures constrains an important aspect of strain history. This information is critical for linking different orogenic levels, constraining tectonic processes, and inferring rheological information (Fig. 10). Pfiffner and Ramsay (1982) evaluated strain rates in orogenic belts, ranging from 10^{-14} to 10^{-16} 1/s, using finite-strain estimates and approximating the duration of orogenesis. Tectonically oriented studies typically estimate strain rates by measuring cross-cutting relations, which is a useful, albeit indirect, way of constraining the problem. Direct measurement of strain rate is also possible; the most direct dating of deformation analyzed fringes on pyrites, which were used extensively for determining strain histories

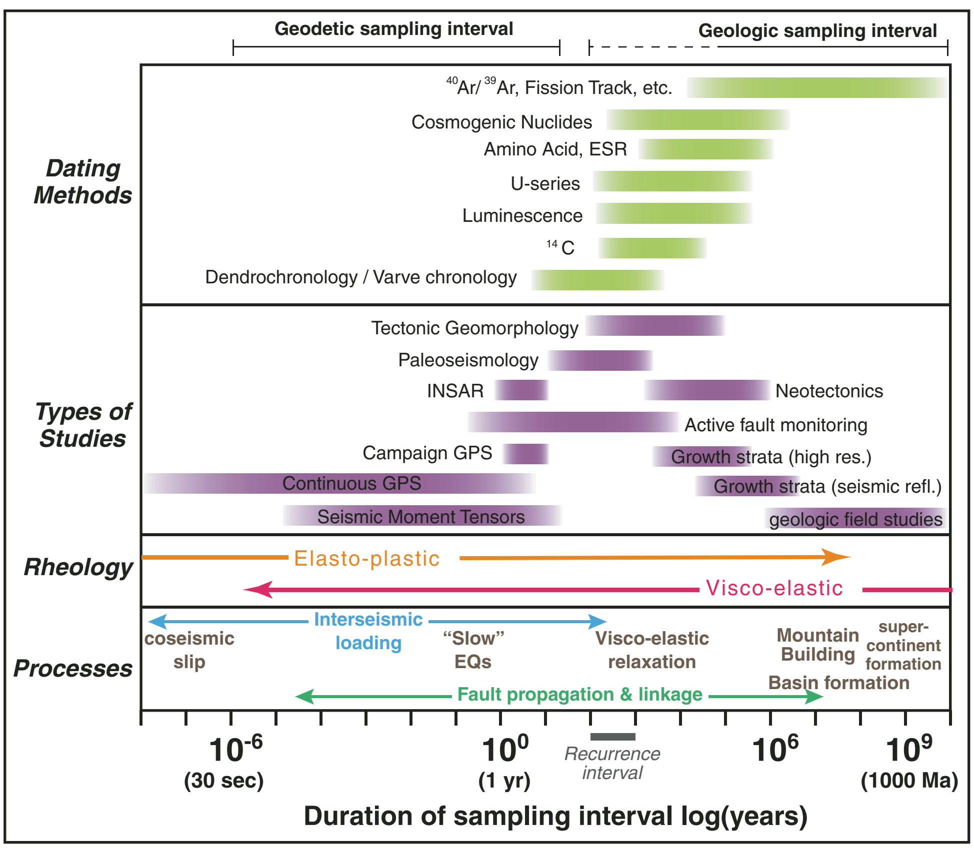


Figure 10. From R. Allmendinger (*in* Pollard et al., 2002). Graph showing the 15 orders of temporal magnitude used in studying lithospheric deformation. The geochronological methods, geodetic and structural studies, rheology, and processes exhibit wide variation. The time band from decades to tens of thousands of years is the most poorly understood, although it is critical to understand how short-term geodetic behavior results in long-term orogenic behavior. ESR—electron spin resonance; INSAR—interferometric synthetic aperture radar.

(e.g., Elliott, 1972; Choukroune, 1971; Etchecopar and Malavieille, 1987). Muller et al. (2000) combined the strain histories inferred by Aerden (1996) with detailed Rb/Sr analyses on large strain fringes to calculate rates of rotation. Their strain-rate estimates range from 1×10^{-15} to 8×10^{-15} 1/s. The $^{40}Ar/^{39}Ar$ method has also been a powerful means to date deformation by obtaining ages on minerals (e.g., phengite) that grow *below* their closure temperature; hence the age of mineral growth directly dates the time of deformation. This method was used by Dunlap et al. (1991) to date thrust sheets in the Alice Springs orogeny in central Australia (Fig. 11). Because the minerals dated were within quartzites and directly comparable with the results of experimental studies (e.g., Hirth and Tullis, 1992), Dunlap et al. (1997) were able to integrate the age information and rock microstructures to derive the strain rate through time within this mid-crustal thrust-fault duplex.

The implications of quantitative age control for structural geology studies are profound, but too broad to cover here. The development of in situ spatial dating methods (e.g., LA-ICPMS [laser ablation, inductively coupled plasma mass spectrometry], SHRIMP [sensitive high-resolution ion microprobe], and microprobe dating of monazite) now allow ages from specific crystallographic, structural, and/or compositional domains within minerals to be directly linked to their microstructural context. These integrated microtextural, isotopic, and chemical approaches are reshaping modern structural analysis by linking structural and metamorphic fabrics to infer more complete pressure-temperature-deformation histories (e.g., Baxter et al., 2002; Williams and Jercinovic, 2002; Kelly and Harley, 2005; Reddy et al., 2009; Timms et al., 2011).

STRUCTURAL GEOLOGY STUDIES IN ACTIVE TECTONIC SETTINGS

It is worth initially addressing what is meant, for us, by the terms *neotectonics* and *active tectonics.* Neotectonics is an inclusive term used for scientists studying the recent geological past: Neotectonics is the study of young tectonic events that have occurred or are still occurring after a previous tectonic event (a paraphrase of Pavlides, 1989). Although opinions vary, for some workers, neotectonics includes deformation up to post-Miocene (e.g., Slemmons, 1991). In academic geology, the definition of *active tectonics* typically refers to deformation occurring after ca. 130 ka. Note, however, that Wallace (1986)—who coined the term *active tectonics*—described it as 500 ka and younger. In the United States, "active" has a legal definition (which varies in different governmental organizations) of a certain time scale, including as low as any post-Holocene (~12,000 yr B.P.) movement. For example, the State of California defines an active fault as one that has ruptured in the last 11,000 years (California Department of Conservation website); the State of Nevada distinguishes between Holocene active faults (less than 10,000 yr B.P. motion), late Quaternary active faults (less than 130,000 yr b.p. motion), and Quaternary active faults (less than 1.6 Ma motion) (Nevada Earthquake Safety Council); and the U.S. Bureau of Reclamation defines active faults as showing displacement in deposits younger than 130,000 years. For this publication, we will use 130 ka as the reference time interval for active tectonics. This definition is practical: Use of Holocene (ca. 12 ka) is too restrictive to employ in slowly deforming regions, whereas deformation of ca. 150 ka and older can be studied in more traditional ways.

In an analogous manner, there is no encompassing term for the type of structural geology that is practiced in these settings (active structural geology?). Neotectonics is similar to tectonics insofar as there are many disciplines within the geological sciences that study neotectonics, including structural geology, geophysics, geomorphology, sedimentology, and volcanology. The role of structural geology in active tectonics is most similar to its role played in "regular" tectonics, and the same skill sets are often used (e.g., cross-section drawing and balancing, fault reconstruction, producing structure-contour maps of important stratigraphic horizons, microstructural analysis). Structural geologists are, however, engaged in different, recognized fields associated with active tectonics. *Earthquake geology* (Yeats et al., 1997) is the study of the structural geology of individual or successive earthquake ruptures taking place along an active fault. It also includes the pattern of deformation, such as warping or folding, that affects the Earth's surface as a result of slip on buried (blind) earthquake ruptures. *Paleoseismology* is a subdiscipline of earthquake geology that focuses on describing and interpreting the distribution and chronology of prehistoric earthquakes in space and time, with special application to seismic hazard evaluation (e.g., McCalpin et al., 2009). Much of the work in these fields requires structural geology because of their emphasis on understanding deformation, although there are also a considerable amount of geophysical and engineering techniques involved in earthquake geology. *Tectonic geomorphology* evaluates the interplay between deformational and surface processes that shape the landscape (Burbank and Anderson, 2001). Because of the preeminent importance of the landscape surface as a recorder or marker of recent deformation, many structural geologists work primarily in the field of tectonic geomorphology. In addition to the geomorphology, structural geologists working in these settings are focused on the characterization of the geological structures and their development, including grain-scale processes.

Much of the major progress in structural geology of active regions was done in two main areas: California (and directly adjacent Nevada) and New Zealand. There is no question that high-quality work has been done in a variety of other actively deforming regions, including the Himalayas, Japan, Cascadia subduction zone, U.S. Intermountain West, Andes, and eastern Mediterranean region; the Himalayas, in particular, are well studied in terms of tectonic geomorphology. We concentrate on the San Andreas fault system, the Walker Lane system, and the Alpine Fault of New Zealand for brevity, and because they have had sufficient focus and resources (e.g., the Southern California Earthquake Center) to allow in-depth questions to be answered.

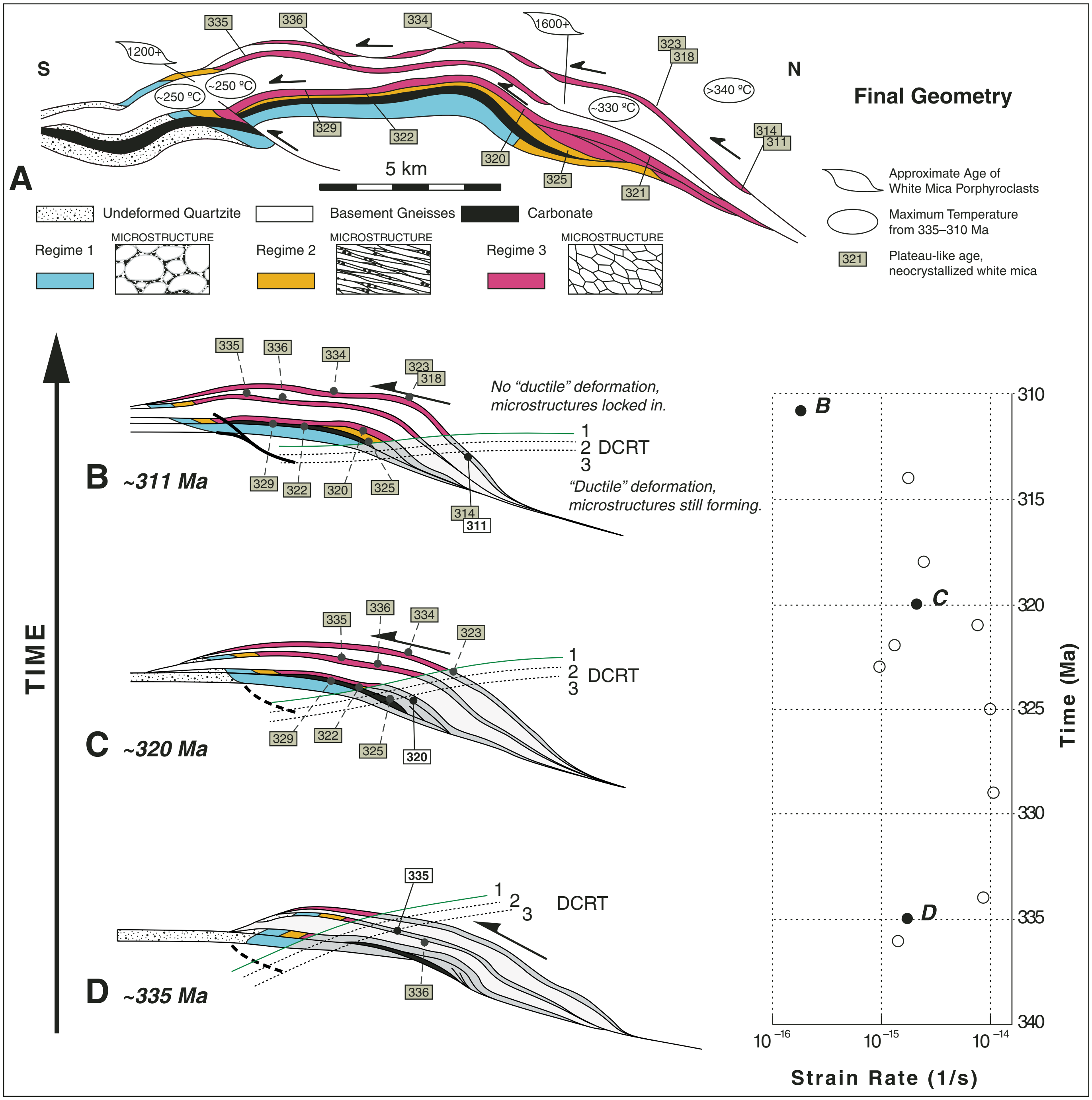


Figure 11. A synoptic diagram based on the study of Dunlap et al. (1997). The top diagram shows the recorded microstructural regimes (after Hirth and Tullis, 1992), and the boxes show the age of white mica grown below the closure temperature, which thus date the deformation. The lower diagrams show the evolution of the thrust system. The green line on the diagram separates continuously deforming material below, and the colors indicate rocks where the microstructures are "locked in" at any time slice because they cooled below ~325 °C. The right plot shows the evolution of strain rate; the black dot represents the strain rate for the time. DCRT—Dislocation creep regime transition zone for wet quartzite.

Paleoseismology and Tectonic Geomorphology

Historically, one might argue that paleoseismology started with Strabo (64 BCE to 24 CE), a Greek geographer from the south side of the Black Sea, who inferred that earthquakes are due to movement on faults. In the United States in the 1900s, a famous early example of earthquake geology is the Lawson report (Lawson and Reid, 1908), which documented deformation associated with the 1906 San Francisco earthquake. In New Zealand, Wellman (1953, 1983) led the way in characterizing offset on active faults, although the timing of the non-historical offsets was largely constrained by geological inference. From a twentieth century perspective on paleoseismology, G.K. Gilbert deserves mention. In a posthumously published report on the Basin and Range, he discussed fault scarps, segmentation, and recurrence intervals (Gilbert, 1928). Yeats (2012) succinctly summarizes the initiation of the field of active tectonics in the post–World War II era. He particularly emphasizes the role of R. Wallace of the U.S. Geological Survey and the field-based observation of active geological structures (e.g., Wallace, 1949, 1968).

By the early 1980s, tectonic geomorphology started to become an active area of research, owing largely to the ability to date landscapes. For example, Muhs (1983) was able to date uplifted marine terraces off San Clemente Island and correlate them to Quaternary high sea-level stands. By conducting this research along the west coast of the United States, and looking at the current relative position of the marine terraces, it was possible to estimate uplift rates (e.g., Muhs et al., 1992). Typically, the New Zealand geologists seem to have started estimating uplift rates with marine terraces earlier (e.g., Bull and Cooper, 1986), but were sufficiently understated about their findings that they did not receive adequate attention. Coral marine terraces can be used effectively to constrain seismic uplift, because the growth of carbonate allows them to be well dated (e.g., Ota et al., 1993). Although these marine terraces are commonly assumed to rise incrementally during seismic events, it is actually unclear whether the uplift process is always seismic. A recent example from New Zealand suggests that this process can occur entirely aseismically (Wilson et al., 2007).

The 1980s also saw the integration of geomorphology, structural geology, dating methods, and the use of trenching. K. Sieh has two highly significant publications during this period. Sieh (1984) utilized trenching across the active San Andreas at Pallett Creek to work out the movement history. This study was based on his dissertation work (Sieh, 1978), but he provided better age constraints and more earthquake cycles. The first use of trenching seems to have been by M. Clark and colleagues at the U.S. Geological Survey in the Salton Trough in California (Clark et al., 1972; R. Yeats, 2012, personal commun.). Second, in the Carrizo Plain of central California, Sieh and Jahns (1984) documented right-lateral offset associated with the San Andreas Fault at Wallace Creek (named for R. Wallace of the U.S. Geological Survey). In this locality, these authors identified and dated several large past seismic events and used them to constrain the slip rate for this segment of the San Andreas Fault. Recent studies by Ludwig et al. (2010) and Zielke et al. (2010) took account of the rate of channel incision to refine estimates of offsets on the large earthquakes on this segment and to demonstrate that the characteristic slip model is applicable to this segment. These studies underscore the need for deeper understanding of the linkages between landscape evolution and structural processes.

The linkages between structural geology and landscape evolution continued to be explored. Medwedeff (1992) documented a growing anticline above a blind thrust at Wheeler Ridge, which propagated laterally as a result of increased throw on the fault. The study was groundbreaking for its use of structural models and seismic data to make sense of the observed geomorphology and geology. Wheeler Ridge also is a well-studied system for evaluating the interaction of geomorphology and structural geology (e.g., Burbank et al., 1996; Keller et al., 1998), including estimating uplift and lateral propagation rates.

The use of geomorphology has also been extended to river terraces. This work has arguably been done most notably in New Zealand, where a series of river terraces formed from glacial outwash have been cut down through time. This situation has allowed the quantification of deformation of these surfaces over time (e.g., Lensen, 1968; Mason et al., 2006). This approach has been conducted in a variety of other settings, including the Himalayan collision (e.g., Cowgill, 2007). A particularly integrative study was done on the fluvial terraces in the foreland thrust belt of central Nepal (Lavé and Avouac, 2000). This study documented the folding of these terraces, which was then traced to fault movement using kinematic models of fault-bend folding. The result suggested that nearly all of the convergent motion, imposed by plate motion in the India-Asia collision, is currently accommodated by the Main Frontal Thrust.

The Wrightwood study of Weldon et al. (2004) is a particularly well-constrained and accessible example of paleoseismology and earthquake geology. Using 45 trenches along the San Andreas Fault, this study documented when faults were active and the amount of movement on each fault. The data set allowed the authors to test whether faults are predictable at all and, if so, to distinguish between three distinct models of earthquake recurrence. The three models test whether (1) the length of seismic quiescence depends on the size of the prior earthquake (time-predictable), (2) the size of an earthquake depends on the amount of time since the previous earthquake (slip-predictable), or (3) the likelihood of an earthquake depends on the deviation of cumulative slip from an averaged long-term slip rate (strain-predictable) (Fig. 12). The Wrightwood data are not consistent with the slip-predictable and time-predictable models, but they appear consistent with the strain-predictable model.

One of the most important advances in tectonic geomorphology is how mass distributions associated with surface process affect the growth of mountain belts. Koons (1990) and Beaumont et al. (1992) postulated the coupled nature of climate, erosion, and mountain building. The basic idea is that rapid erosion and rapid uplift are coupled: In this way, the crust/lithosphere could be exhumed in a mountain belt through rapid erosion caused by climatic variables.

With time, new rich and detailed data sets have become available because of GPS surveying capabilities and light detection and ranging (LiDAR) technologies that allow the construction of digital elevation models (DEMs), which reveal fault traces and allow displacements of landforms to be measured in 3D with unprecedented detail and accuracy. Many of the above studies rely on the presence of geodetic data as a standard for comparison. As a result, the problem of a mismatch between geological versus geodetic data is a recurring theme in several recent publications (e.g., Allmendinger et al., 2009). For example, Oskin et al. (2007) document a mismatch in theWalker Lane shear zone (Eastern California shear zone) between geologic and geodetic rates, but suggest that this situation indicates that the geodetic rates are not constant over time (strain transients are occurring). Farther north in the same zone, Wesnousky et al. (2012) note a similar problem, but suggest that active deformation may be occurring by distributed, aseismic shearing. Similar comparisons of geodetic–interferometric synthetic (InSAR) images and geological observations have been conducted in a variety of tectonic environments (e.g., Jackson, 1999; Peltzer et al., 2001; Friedrich et al., 2004) and often show a disparity.

A good example of how earthquakes are analyzed in geodetically well-instrumented areas—and combined with other advances in earthquake geology—is demonstrated by the 2010 Darfield earthquake near Canterbury, New Zealand (Beavan et al., 2010; Holden et al., 2011; Quigley et al., 2012). The **M** 7.1 Darfield earthquake occurred on the unmapped Greendale fault, had a 5.3 m right-lateral displacement on a 29.5-km-long fault, and occurred in a low-strain region of New Zealand. Quigley et al. (2012) evaluate how well fault-scaling relationships developed by Wells and Coppersmith (1994) and Wesnousky (2008) addressed the magnitude of this earthquake. The result was that the geologically based estimates for magnitude were lower than those calculated from the observed seismic movement. Another interesting outcome from this study is that one can investigate the seismic displacement field in significant detail. It appears that one-half of the slip was 100 m from the fault, indicating a large component of distributed, off-fault deformation.

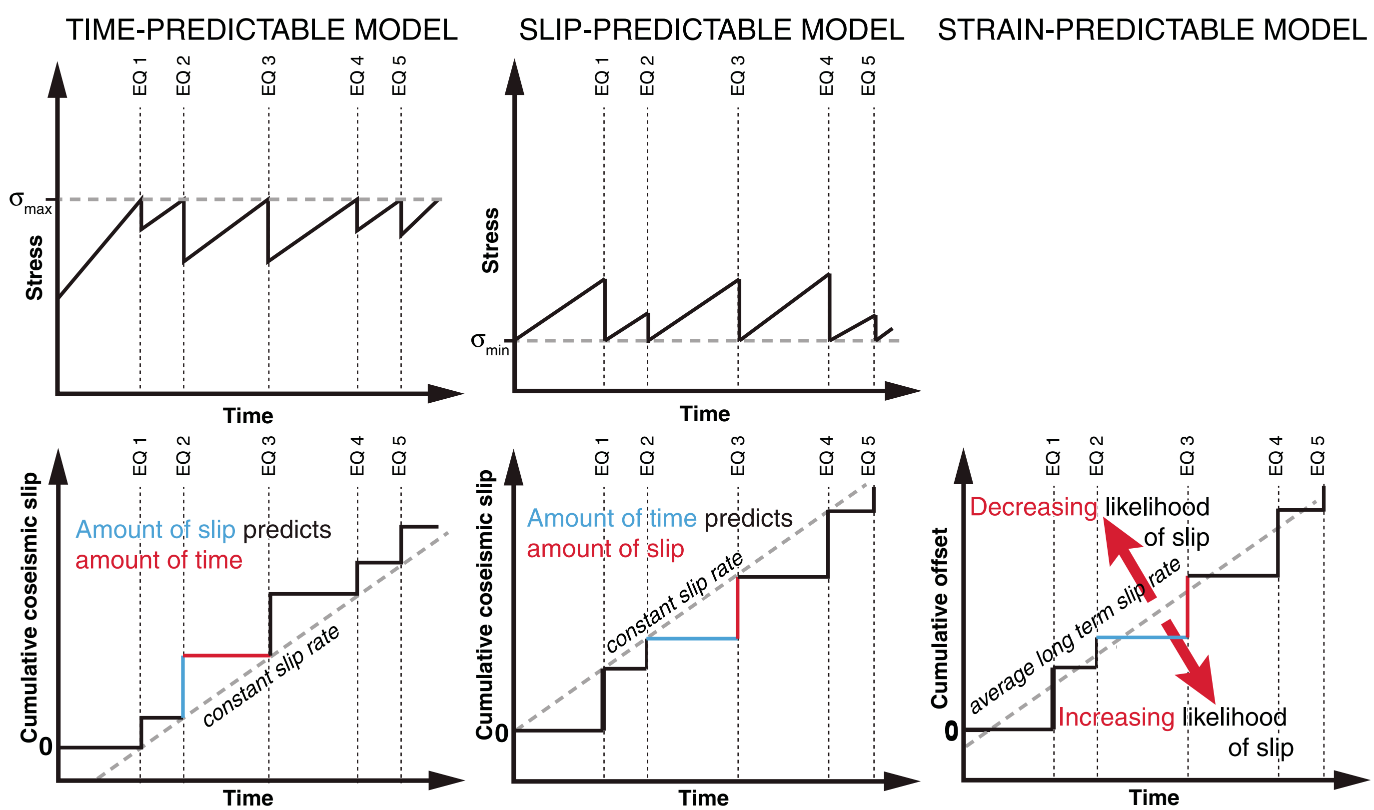


Figure 12. (A) Time-predictable, (B) slip-predictable, and (C) strain-predictable models for earthquake prediction. (A) The time-predictable model predicts that the fault will slip when an upper stress magnitude is reached. Therefore, the amount of slip on the fault predicts the time until the next slip event (but not the amount of slip). (B) The slip-predictable model suggests that the fault will slip until it reaches a low level of stress. Therefore, the amount of time since the last seismic event predicts the amount of slip (the longer one waits, the larger the earthquake). (C) The strain-predictable model notes the difference between the cumulative departure from mean displacement and where the fault is in a long-term cycle. There is an increased likelihood of seismic slip if the latest slip events have not kept up with long-term movement. Plots A and B are based on Shimazaki and Nakata (1980). EQ—earthquake.

More recently it has become possible to try to investigate the relative effects of tectonic uplift versus geomorphic development. Using a restraining bend in the Carrizo Plain in California, Hilley and Arrowsmith (2008) documented the geomorphic evolution of a ridge (Dragon's Back) as it moved through a restraining bend. The authors were able to document that change in rock uplift rate occurred on the thousand year cycle, but hillslope processes took greater than an order of magnitude more time to adjust to uplift rates.

An underutilized approach is the documentation of microstructures along active geological structures. Cashman et al. (2007), for example, noted a difference in microstructures between the central creeping and locked segments of the San Andreas Fault. One problem is obtaining materials without the cost of trenching (or coring, in the case of the San Andreas Observatory at Depth; e.g., Zoback et al., 2011). Direct fault observation is commonly difficult for very young deformations, as only normal faults get self-exhumed. One spectacular example of fault exhumation is made available by the new shuttle imagery on newly forming extensional complexes (e.g., Daymon Dome, Papua New Guinea; Spencer, 2010). With the advent of ground-based LiDAR, structural geologists have also been documenting the detailed geometries of exposed fault surfaces. This information is also being obtained and utilized by scientists interested in the mechanics of earthquakes (e.g., Sagy et al., 2007; Brodsky et al., 2011).

The Seismic Cycle

Allmendinger et al. (2009) attempt to elucidate the links between geodetic information and finite strain measurements typically collected in studying ancient orogens. The difficulty is that the methods and time scales of geological and geodetic data largely do not overlap (Figs. 10, 13). A particular problem is that the seismic cycle (Fig. 13) is thought to dominate the geodetic signal. The "classic" seismic cycle (e.g., Reid, 1910; Fig. 13A) describes the time and deformation buildup between earthquake events; the only deformation that accumulates is elastic strain (Fig. 13A2), which is completely released when the earthquake occurs (Fig. 13A3). Many workers use models for seismic behavior that are built on this assumption and that utilize dislocation theory (e.g., Savage and Prescott, 1978; Savage, 1983). While there is no doubt that the dominant signal in the geodetic studies results from elastic strain buildup associated with the seismic cycle, some permanent non-recoverable strain occurs adjacent to large faults, which is inconsistent with the "classic" seismic cycle model (Fig. 13A). This permanent deformation is recorded using geodetic data (e.g., Hyndman and Wang, 1995) and fault formation (e.g., Loveless et al., 2009) in subduction-zone settings, distributed wrenching in strike-slip settings (e.g., Titus et al., 2007a), and modeled in analogue studies (Cooke et al., 2013). Thus, one can consider a "progressive"

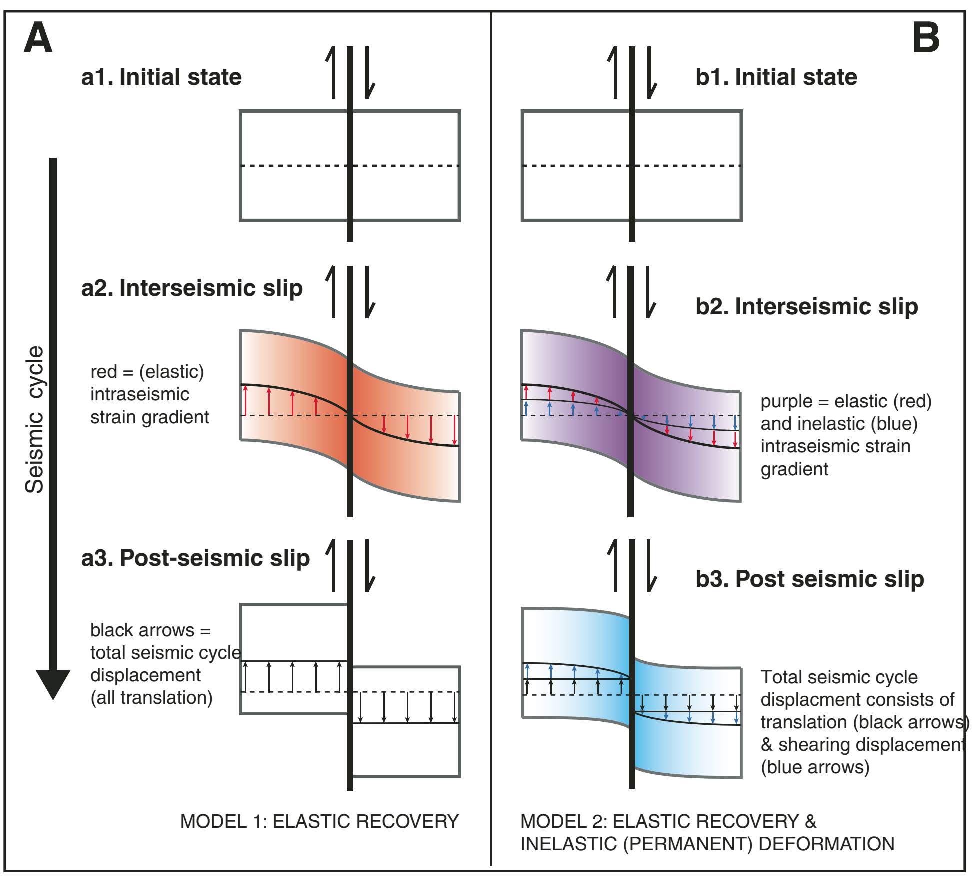


Figure 13. Two plots of the seismic cycle. (A) The "classic" seismic cycle, in which all the deformation adjacent to a major fault is elastic and is recovered at the end of the seismic cycle. Despite the fact that this model is incorrect, it is still the most commonly used. (B) A "progressive" seismic cycle, in which deformation adjacent to a major fault is a combination of recoverable (elastic) strain and permanent strain. The relative magnitudes of recoverable versus permanent strain is unknown, although the recoverable strain is likely >80%.

seismic cycle—insofar as progressive deformation occurs adjacent to faults—as shown in Figure 13B.

One place where it is currently possible to directly evaluate permanent versus elastic strain accumulation is the central creeping segment of the San Andreas Fault (Fig. 14). At this location, the fault creeps nearly as fast as the long-term slip rate inferred from adjacent locked segments (Titus et al., 2006). The 1906 San Francisco earthquake ruptured down to the creeping segment from the north, and the 1857 Fort Tejon earthquake ruptured up to it from the south. Recent earthquakes in the area, including the 1983 Coalinga, 1985 Kettleman Hills, and 2003 San Simeon earthquakes, all occurred on off-fault structures, as opposed to the creeping segment. Thus, it is possible that little elastic strain occurs on this segment, although there is some weak historical evidence of earthquakes along this section (Toppozada et al., 2002), and recent trenching may indicate that some seismic faulting has occurred (N. Toke, 2012, personal commun.). The geodetically derived strain gradient parallel to the fault is distinct along this segment compared to other places along the San Andreas Fault in central California (Fig. 14A; Rolandone et al.,

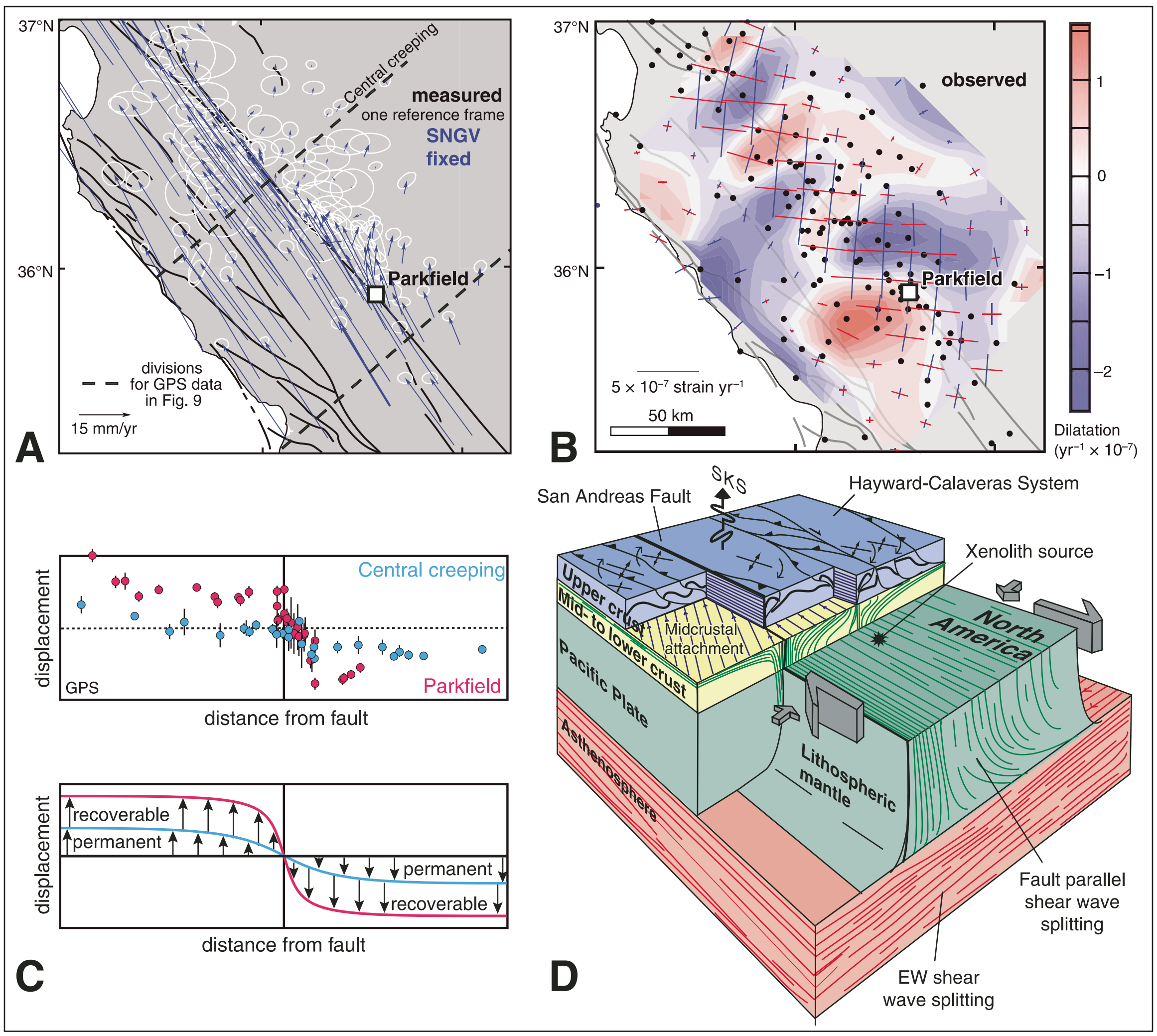


Figure 14. The creeping section of the San Andreas Fault, California. Because the central creeping segment moves at nearly plate-motion rates (~28 mm/yr), it is the one place along the San Andreas Fault where it may be possible to look through the elastic deformation and observe the permanent deformation in real time. (A) The relative motions of the two sides of the fault, with the E (Sierra Nevada block) side fixed, as recorded by a network of GPS stations. (B) Infinitesimal strain axes (blue, shortening; red, elongation) along the same section of the San Andreas Fault. (C) The fault-parallel motion of GPS stations near Parkfield (red) and in the central creeping segment (blue), with zero "pinned" to the fault. The Parkfield stations move significantly more because they also record elastic strain. The central creeping segment is interpreted to record dominantly permanent deformation. (D) A three-dimensional block diagram of California, with mantle deformation inferred by shear-wave studies and mantle xenoliths exhumed from underneath the Calaveras Fault. Modified from Titus et al. (2011). SNGV—Sierra Nevada–Great Valley.

2008; Titus et al., 2011), presumably because of the lack of elastic strain buildup (Fig. 13B). This strain is thought to be permanent, rather than elastic; this amount of permanent deformation is broadly consistent with the observed geology.

Studies of near-surface deformation associated with the San Andreas Fault can be combined with other data to constrain the character and kinematics of the fault or shear zone in the lower crust and upper mantle (Fig. 14D). Seismic imaging across the northern California Coast ranges clearly indicates that the San Andreas and subsidiary faults are nearly vertical, cut through the entire crust, and offset the Moho (e.g., Hole et al., 1998; Henstock and Levander, 2000). Seismic anisotropy data from western California (Özalaybey and Savage, 1994) and data from mantle xenoliths (Titus et al., 2007b) indicate distributed deformation in the uppermost mantle that allows one to estimate the lithospheric structure associated with deformation (Fig. 14D). Given the lithospheric scale of the San Andreas and other major strike-slip faults (e.g., Mooney et al., 2007), future work will likely concentrate on considering what lithospheric layer(s) dominate the mechanical response of this system. It remains possible that even these earthquakes might be generated in the lower ("viscous") parts of these systems (Handy et al., 2007).

There have been many attempts to model the geodetic signal in areas where locked faults occur. Block models are one potentially useful method (e.g., Matsu'ura et al., 1986; Meade and Loveless, 2009). Block models describe relative motion across discrete structures as well as the smooth velocity gradients produced by accumulation of elastic strain on these structures by integrating seismic cycle models with microplate rotations. Using this approach, assumptions have to be made about the location and extent of the locked faults. Increasingly sophisticated models that include viscous layers below an elastic crust (e.g., Johnson and Segall, 2005), as well as geodetic data through multiple seismic cycles on specific faults, may allow us to more fully understand the exact nature of seismic-related deformation.

RHEOLOGY

Rheology describes the relationship between stresses and strain rates (or strain, for elastic material) for any particular material. Consequently, it is through rheological analyses that the kinematics can be linked to the dynamics (or vice versa). Experimental studies of the rock-dependent rheologies of geological materials started 50 years ago with analyses of data from axial-compression, laboratory deformation experiments (e.g., Griggs and Handin, 1960). At that time, experimentalists examined the deformation of single crystals of common rock-forming minerals and began what we now know are pioneering studies of monomineralic and polymineralic rocks under a range of physical conditions (e.g., Heard et al., 1972). At present, our understanding of the rheology of geological materials and their deformation mechanisms are derived from increasingly sophisticated experimental studies, numerical modeling of fundamental deformation processes (e.g., intracrystalline plasticity by dislocation movement, mineral fracture), and field-based studies of naturally deformed rocks (to infer deformation mechanisms using microstructural evidence and deformation conditions using mineral chemistry). In addition, studies are no longer limited to the analysis of steady deformation under uniform conditions; workers regularly examine how changes in temperature, pressure, differential stress, pore fluid pressure, and deformation rate affect which deformation mechanisms predominate and the constitutive properties of rocks under different conditions.

Recent work has focused on the recognition, examination, and significance of rheological layering in Earth. In this context, we note a profound proliferation of the terms *brittle* and *ductile* in referring to the rheological behavior of rocks at different places in Earth. Both *brittle* and *ductile* have been used to connote such a wide range of geometrical and rheological characteristics that they are as likely to be confusing as illuminating, a point made by Rutter (1986) and Snoke et al. (1998). The terms *brittle* and *ductile* were initially used to describe the characteristics of experimentally deformed rocks. Heard (1960) defined ductility as a material's ability to withstand more than 5% permanent strain before failure. Part of the confusion in defining *brittle* and *ductile* resulted from early experiments in which the microscopic-scale deformation mechanisms corresponded well with mesoscopic changes in displacement fields. However, more recent experimental studies show several types of transitions, including *ductile faulting,* accommodated by plastic deformation (Post, 1977), as well as *delocalized cataclasis,* homogeneous deformation characterized by microscopically brittle deformation (Wong and Baud, 2012).

Field geologists typically use these terms in two different ways: (1) to distinguish between deformation zones in which the material has lost cohesion (brittle) or not (ductile) (e.g., definition of van der Pluijm and Marshak, 2004), or (2) to distinguish the manner by which deformation occurs; e.g., *brittle deformation* comprises fracturing and faulting, both of which generate abrupt local changes in displacements, whereas *ductile deformation* has smoothly varying displacements associated with folding and fabric formation (e.g., Twiss and Moores, 2007). Both of these field-based definitions explicitly depend on the scale at which the observations are made (Snoke et al., 1998). Further, deformation mechanisms may change during deformation, so that any individual rock sample or geologic structure may have characteristics of both brittle and ductile deformation. Moreover, neither of these common usages is consistent with the definitions derived from experimental work, nor are the characteristics of deformation (amount of strain incurred before failure or formation of a fracture or fault) required to apply these experimental definitions ever known about any geological structure observed in the field.

Brittle and *ductile* can also be used to distinguish different deformation mechanisms or processes at the microstructural scale. Brittle deformation is defined precisely by the degree of lattice distortion accompanying fracture (e.g., Lawn, 1993). Ductile processes include twinning and intracrystalline slip and diffusional mass transfer (e.g., Burkhard, 1990; Williams et al., 1994). Finally, geophysicists typically use these terms to distinguish bulk behaviors of different parts of the lithosphere, such as "brittle" upper crust and "ductile" lower crust. We are sensitive to the sentiments of Tullis et al. (1982, p. 230): "As structural geologists, we are interested in the processes and conditions of deformation; we need to determine criteria for the recognition of faults and shear zones; rather than arguing about terminology." It is clear, however, that the currently used terminology is now an impediment to our understanding.

Climbing Out of a Nomenclature Tangle

In an attempt to distinguish and illuminate the different aspects of rheology as used by different communities in the geological sciences, we employ the framework given in Figure 7 (based on Sibson, 1977; see Acknowledgments for other contributions). Field geologists assume that rocks are mesoscopically discontinuous at shallow crustal levels and mesoscopically continuous at deeper (>12 km in a "typical" geothermal gradient) crustal levels. In fact, this dichotomy is what most field-based geologists mean when they use *brittle* and *ductile* (e.g., van der Pluijm and Marshak, 1997). If microstructures are added to the framework (Fig. 7), there is likely to be wider agreement on the difference between deformation with limited recovery (cracks, "brittle") versus deformation with recovery at elevated temperatures, with no evidence of dilatant processes. Rutter (1986) defines three fundamental deformation mechanisms: cataclasis, intracrystalline plasticity, and diffusive mass transfer (solution transfer at low temperatures). Cataclasis (elastic distortion, grain-scale fracture, and frictional sliding) and solution transfer dominate at lower temperatures, while dislocation and diffusion creep dominate at higher temperatures. Thus, one could argue broadly that the upper lithosphere deforms generally by cataclastic and/or solution transfer deformation mechanisms (cataclastic-solutional?; Gratier et al., 1999), whereas the lower lithosphere deforms by dislocation creep and/or diffusion creep (dislocational-diffusional?) (Fig. 7).

The breakthrough made by Sibson (1977) was to recognize that field observations of material continuity (faults versus shear zones) generally correlate to a change in deformation mechanisms (cataclastic-solutional versus dislocational-diffusional) at the "brittle-ductile" transition. This approach, however, has some shortcomings. The presence of foliated gouge and local areas of distributed-displacement deformation (e.g., folds) in fault zones (Mitra, 1984; Chester et al., 1985; Rutter, 1986) suggests that there is not a one-to-one correspondence between deformation mechanisms and material continuity. The recognition of "brittle-ductile" structures (Ramsay, 1980) is another argument that field-based inferences of material behavior do not correspond directly to deformation mechanisms and/or that different deformation mechanisms can operate under similar conditions (e.g., in different rock types). Further, evidence compatible with earthquake generation at temperatures where dislocational-diffusional deformation mechanisms are inferred (e.g., Passchier, 1982; Hobbs et al., 1986) suggests that behaviors may be transient and depend on local external conditions.

The traditional "brittle versus ductile transition" is also emphasized by lithospheric strength profiles, based on flow laws, that require an assumption of differences in material behavior at different lithospheric levels (Goetze and Evans, 1979; Brace and Kohlstedt, 1980). The upper crust in these diagrams is assumed to follow a constitutive relationship for friction, based on Byerlees' law, which describes a well-defined critical (time-independent) shear stress that depends on normal stress and thus depth. The middle-lower crust behaves according to the power-law constitutive relationship of thermally activated deformation of quartz. In this simplification, the upper crust is frictional (pressure-dependent plastic), while the middle-lower crust is viscous (e.g., Fusseis and Handy, 2008). This distinction does not allow for the influence of diffusion and other non-frictional processes in the upper crust, but is a reasonable simplification of the assumptions implicit in lithospheric strength profiles. As experimentally derived flow laws become available for other minerals, deforming by different mechanisms, wet and dry, and with varying grain sizes, further models of the strength of the lithosphere are being explored (e.g., Vissers et al., 1996; Jackson, 2002; Bürgmann and Dresen, 2008).

A final division is that used by numerical modelers in attempting to simulate and predict lithospheric deformation. In zones where localization does not occur, the upper crust is typically simulated as elastic, whereas the lower crust is considered viscous. If deformation is localized, deformation in the upper crust is assumed to be elastoplastic. The use of *plasticity* here is in a continuum mechanics sense, meaning that strength is not dependent on strain rate (i.e., characterized by flow at a specific yield stress). Our use of plasticity has no implication of (1) a particular deformation mechanism or collection of deformation mechanisms, and (2) no implication of "crystal-plasticity," which refers to low-temperature dislocation slip (and twinning) and is described by highly nonlinear bulk viscous material behavior. At upper-crustal conditions, Mohr-Coloumb behavior or frictional behavior is typically assumed, although this assumption has difficulties, as outlined above. Localized deformation at lower lithospheric levels is assumed to be visco-plastic, with von Mises plasticity controlling the orientation of the localization zones while viscosity controls the deformation.

Following from the discussion above, we contend that in referring to "the brittle upper crust," different structural geologists might intend to indicate that the upper crust (1) is dominantly faulted, (2) lacks microstructures associated with recrystallization and recovery, (3) exhibits mainly cataclastic and diffusional mass transfer deformation mechanisms, (4) is characterized by frictional behavior, or (5) follows elastic or elastoplastic flow laws. They are all simplifications of reality, and they are not mutually compatible. Further, any model is a simplification of reality, even for a given rock type. As examples, diffusive mass transfer and metamorphic-fluid-assisted reactions at upper crustal conditions may result in continuous deformation, whereas sufficiently fast strain rates or high-fluid pressures may result in discontinuous deformation at higher pressure and temperature conditions. Taking care to describe deformation in detail at different scales of observation and to specify deformation mechanisms and strain distribution will result in less ambiguity and will enhance communication among geoscientists with a wide variety of backgrounds.

The important issue, however, is to note the astounding progress that has been made on understanding rheology of the lithosphere in the past 50 years. The work on this topic is truly the integration of field observations (e.g., Sibson, 1977), experimental deformation (e.g., Brace and Kohlstedt, 1980; Hirth and Tullis, 1994), microstructural investigations informed by experimental deformation (e.g., Dunlap et al., 1997), and the feedback to numerical models to simulate regional- to orogenic-scale deformation. A recent article by Bürgmann and Dresen (2008) provides a comprehensive review of the rheology of the crust and upper mantle with a focus on laboratory results. Hence, we will concentrate here on the contributions of structural geologists to understanding the role of rheology in strain localization.

Rheology of Shear Zones

The most significant advance in understanding the rheology of shear zones was the resolution, in the 1970s, of the 100-year-old controversy regarding the deformation mechanisms responsible for the formation of mylonites (e.g., Nicolas et al., 1971; Bell and Etheridge, 1973; White, 1976). Mylonites have long been identified as fine-grained, foliated and lineated, coherent rocks found within narrow zones that accommodated displacement (e.g., Bell and Etheridge, 1973; White, 2010). Decades of discussion focused on which processes led to the grain-size reduction: cataclastic versus crystal-plastic processes (e.g., Higgins, 1971; Bell and Etheridge, 1973). Bell and Etheridge (1973) showed that dislocation creep and diffusional mass transfer are the dominant deformation mechanisms operating in many mylonite zones, and observations of experimentally deformed rocks confirmed this conclusion (e.g., Tullis et al., 1973). However, as noted above, observations of both naturally and experimentally deformed rocks demonstrate that assigning a single mechanism is not straightforward because (1) different minerals deform by different mechanisms under the same conditions, and (2) individual shear zones exhibit evidence for cataclastic, plastic, and diffusive mechanisms that occurred coevally (e.g., Mitra, 1978).

Inferences on the rheology of shear zones derive directly from (1) recognition of characteristic microstructures that indicate a dominant deformation mechanism, and (2) experimental calibration of the constitutive relations for those deformation mechanisms. This view of rheology was largely derived from the metallurgical literature, which is such a fundamentally different way of viewing rheology that Nicolas and Poirier (1976) distinguish between the "continuum mechanics method" and "physical metallurgical method," which they note are complementary and operate on different spatial scales. Early experimental work provided data on crystallographic preferred orientations and determined slip systems for deformed aggregates (e.g., Turner, 1953). Later experimental work added detailed microstructural characterization (e.g., Tullis et al. 1973) to samples deformed under a variety of experimental conditions. This microstructural work was combined with the insight of Voll (1961), who, by noting similarities with better constrained work on metals, made an explicit comparison between the observed microstructures and the deformation mechanisms in geological material. Using all these advances simultaneously, it became possible to make overt links between microstuctures, deformation mechanisms, and flow laws. Ashby (1972) was the first to describe deformation mechanism maps, a similar concept was presented for diffusion creep in metamorphic rocks in Elliott (1973) (Fig. 15). The result of this confluence was a proliferation of papers addressing the behavior of geological materials, including olivine-rich mantle (Weertman, 1975), quartz (White, 1976; Ashby, 1977), quartz and calcite (Rutter, 1976), and olivine fracture and flow (Ashby and Verrall, 1978). The use of transmission electron microscope (TEM) images, in particular, provided an essential check to determine the validity of the interpreted deformation mechanism, such as the presence of dislocations (White, 1977) and the nature of grain boundaries (e.g., White and White, 1981). Taken together, this work resulted in one of the major advances in structural geology: Naturally deformed rocks could now be interpreted in

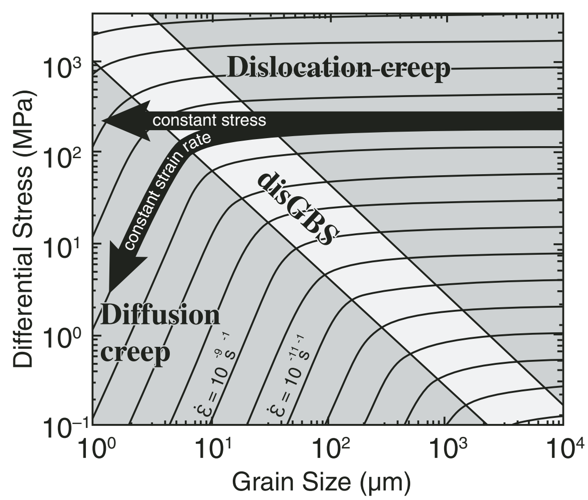


Figure 15. From Hansen et al. (2012). A deformation-mechanism map for dry olivine at 1645 °C. Dominant deformation mechanisms are labeled, with disGBS meaning "dislocation-accommodated grain boundary sliding." The lines are contours of constant strain rate. For localization to occur, one of two paths is inferred (following Schmid, 1983): constant stress (resulting in increasing strain rate) or constant strain rate (resulting in decreased stress). Hansen et al. (2012) addressed this issue with a torsion rig, and suggested that constant stress can produce localization (with a perturbation) and that constant strain rate does not.

terms of extrinsic conditions (P, T, strain rate, fluids), by using microstructural observation (including optical microscopy, scanning and transmission electron microscopy, and CPO) to link to results of experimental deformation, especially when microstructural observations were combined with mineral compositions and geothermobarometry.

An important step in using microstructures to interpret deformation histories in naturally deformed rocks was the characterization of microstructures indicative of different recrystallization mechanisms that reflect the relative rates of dynamic (dislocation producing) versus recovery (dislocation annihilating) processes. Microstructures indicative of different mechanisms of recrystallization dominate under different conditions of temperature and/or flow stress. With increasing temperature and decreasing flow stress, these mechanisms are strain-induced grain-boundary migration recrystallization or bulging recrystallization, subgrain rotation recrystallization, and (high temperature) grain boundary migration recrystallization with both grain boundary migration and subgrain rotation processes operative (e.g., Nicolas and Poirier, 1976; Drury et al., 1985; Urai et al., 1986; Hirth and Tullis, 1992; Stipp et al., 2002). These observations allow relative calibration of deformation conditions in naturally deformed rocks deformed within the dislocation creep regime (e.g., Dunlap et al., 1997).

Theoretical advances also allowed estimation of stress magnitudes in monomineralic shear zones. In one of the earliest and yet still widely used piezometers, Twiss (1977) calculated the steady-state grain size of dynamically recrystallized quartz at different stress conditions. His theoretically derived relation showed good agreement with the earliest experimentally derived relation for olivine (Post, 1973). Empirical calibrations for recrystallized grain size, subgrain size, and dislocation densities in other minerals followed, including calcite (Schmid et al., 1977), olivine (e.g., Ross et al., 1980), and quartz (e.g., Luan and Paterson, 1992; Gleason and Tullis, 1995; Stipp and Tullis, 2003). Later work suggested that piezometers may differ for different mechanisms of recrystallization (e.g., halite: Guillope and Poirier, 1979; calcite: Schmid et al., 1980; Rutter, 1995). De Bresser et al. (1998) further suggested that grain size–stress relations need to account for temperature effects. Austin and Evans (2007) addressed transient recrystallized grain size, as it depends on the balance of nucleation and growth rates.

Empirically derived stress–grain-size relations have been used in conjunction with deformation mechanism maps to interpret deformation pathways through stress-strain rate, and grain-size or temperature space (e.g., Schmid, 1983; Mitra, 1984; Handy, 1989). Elliott (1973) suggested that dynamic recrystallization in calcite resulted in reduced grain size, causing a transition in the dominant deformation mechanism from grain-size insensitive dislocation creep to grain-size sensitive diffusion creep. Schmid (1983) noted that the transition from one deformation mechanism to another, resulting in localization, could occur at constant strain rate or constant stress conditions (Fig. 15). Many others have since suggested similar pathways in other crustal and upper mantle rocks (e.g., Rutter and Brodie, 1988). However, de Bresser et al. (1998, 2001) questioned whether grain size reduction by dynamic recrystallization can lead to a transition to grain-size sensitive creep in monophase aggregates. Rather, these authors suggested that dynamic recrystallization results in a balance between grain size reduction and grain growth processes, so that sufficiently fine grain sizes will not be maintained. The model by Austin and Evans (2007) elaborates on this.

Hansen et al. (2012), in recent high strain torsion experiments of olivine (i.e., monophase) aggregates, explored the effects of constant stress versus constant strain rate boundary conditions on localization behavior in rocks undergoing grain size reduction by dislocation creep accommodated by dynamic recrystallization. Earlier experiments that had been unable to produce localization during dislocation creep were carried out primarily under constant strain-rate conditions. In Hansen et al.'s (2012) experiments, grain size reduction during constant-stress experiments resulted in localization, while constant strain-rate experiments did not. The authors suggest that the localization mechanism in these constant stress experiments was a switch in dominant deformation mechanism to dislocation-accommodated grain boundary sliding (disGBS; Hirth and Kohlstedt, 1995; Drury and Fitz Gerald, 1998) (Fig. 15).

Strain localization in monophase aggregates also occurs by mechanisms other than a transition from grain-size insensitive dislocation creep to grain-size sensitive (dislocation or diffusion) creep. Dislocation-accommodated dynamic recrystallization can lead to new dislocation free and, therefore, weaker grains (Ion et al., 1982; Drury et al., 1985). The development of CPO during deformation can also result in weakening (e.g., Poirier, 1980). Grain-size reduction through fracturing may also result in a change in dominant deformation mechanism to grain-size sensitive diffusion creep (e.g., Elliott, 1973; Mitra, 1984; Wojtal and Mitra, 1986; Rutter and Brodie, 1988).

Strain localization processes in polyphase rocks (polymineralic and/or including a fluid phase) have been far less controversial than in monophase aggregates, and many mechanisms have been identified in both crustal and upper mantle rocks. Dislocation-accommodated dynamic recrystallization in the presence of a second phase can result in a transition to grain size sensitive diffusion creep, if mixing occurs, as a fine grain size can be maintained by pinning of grain boundaries (Olgaard and Evans, 1988; Olgaard, 1990). Metamorphic reactions produce new phases that may pin boundaries, aiding in the maintenance of a fine grain size, but new phases may also have different strengths than original phases. Reaction softening (e.g., White and Knipe, 1978) has been recognized in a wide range of crustal and mantle rock types. While hydration reactions are common, especially at lower temperatures where fluids are important in enhancing reaction kinetics (e.g., Mitra, 1978; Wintsch, 1985; Janecke and Evans, 1988), reactions that are not primarily hydration reactions result in strain localization as well (e.g., Rubie, 1983; Fitz Gerald and Stunitz, 1993).

While the above contributions focused on the influence of metamorphic reactions on deformation, Brodie (1980) and Brodie and Rutter (1985) explored the role of deformation in increasing reaction kinetics, with a focus on basic rocks. The same authors also addressed the role of dehydration reactions in increasing fluid pressure and promoting a transition to grain-size-sensitive creep through cataclasis (Rutter and Brodie, 1988).

Reactions to fine-grained reaction products during deformation in polyphase mantle rocks provide another mechanism for grain size reduction, with reactants serving to pin grain boundaries and maintain very fine grain sizes that likely deform by grain-size-sensitive diffusion creep. Theoretical modeling (Wheeler, 1992) and experimental research (Sundberg and Cooper, 2008) on olivine-orthopyroxene aggregates have suggested that the chemistry of polyphase fine-grained materials may aid weakening when phases share chemical components or influence each other's diffusion. Many of these examples of reaction-enhanced weakening in peridotites are hydration reactions (Brodie, 1980; Handy, 1989; Drury et al., 1991), but high temperatures prevalent at upper mantle conditions facilitate reactions in the absence of fluids, as well (Newman et al., 1999; Furusho and Kanagawa, 1999; Kruse and Stunitz, 1999). Finally, Dijkstra et al. (2002) report a transition from grain-size-insensitive to grain-size-sensitive diffusion creep, and strain localization, as a result of a melt-present reaction. The melt-rock reaction produced small grains and a polyphase material, which inhibited grain growth, preserving the fine grain sizes.

Strain localization as a result of the transition from grain-size-insensitive dislocation creep to grain-size-sensitive diffusion creep has been discussed for upper mantle rocks as well (e.g., Rutter and Brodie, 1988). The recognition of dislocation-accommodated grain-boundary sliding (disGBS) in olivine (Hirth and Kohlstedt, 1995; Drury and Fitz Gerald, 1998) provides an additional weakening mechanism in these rocks (e.g., Jin et al., 1998; Warren and Hirth, 2006). However, disGBS has been identified through rock deformation experiments by its rheology (different from either dislocation creep or diffusion creep), and there are presently no established microstructural criteria for recognizing disGBS; current criteria are based on the grain size within naturally deformed rocks, and the grain size–stress relation for olivine in conjunction with flow laws (as shown graphically on deformation mechanism maps).

Role of Fluids

Water has a drastic effect on the rheology of rocks. The presence or absence of intragranular and intergranular water affects the strength and type of deformation and recovery mechanisms from the grain scale to the lithospheric scale. Minuscule amounts of free water can drastically change point defect populations in minerals, thereby decreasing the climb-accommodated dislocation creep strength of experimentally deformed quartzites (Griggs, 1967; Kronenberg and Tullis, 1984; Ord and Hobbs, 1986; Paterson, 1989; Kohlstedt et al., 1995; Post et al., 1996; Post and Tullis, 1998). For example, quartzite with a ~0.15 wt% water content, deformed at 700 °C, 1500 MPa, and a strain rate of 10^{-6}/s, has a flow stress of ~500 MPa, while quartzite samples with a 0.02 additional wt% (i.e., ~0.17 wt%) of water and deformed under the same conditions have a flow stress of ~250 MPa (Hirth and Tullis, 1992).

On the lithospheric scale, experimental results for wet and dry quartz are often extrapolated and used in strength profiles for the rheology of the lithosphere (Fig. 7) (Kohlstedt et al., 1995; Mackwell et al., 1998; Jackson, 2002; Burov and Watts, 2006). The strength of the lower crust versus the upper mantle is also under debate regarding the presence or absence of water (Jackson, 2002; Burov and Watts, 2006). The experiments of Hirth and Kohlstedt (1996) reveal that the presence of water at a confining pressure of 300 MPa reduces the viscosity of olivine aggregates by a factor of ~140. Water-induced transformation of granulites to eclogites has been suggested as the cause of weakening beneath southern Tibet, whereas a "dry" granulite facies lower crust may be responsible for the strength of the Indian shield (Jackson et al., 2004).

There is a lack of direct evidence from naturally deformed rocks on the weakening effects of water, mostly because it is very difficult to document. Different compositions of fluids can move through rocks at any stage of their deformation or exhumation, and it can be very difficult to pinpoint a specific source or know the timing of the fluid infiltration (e.g., Mancktelow et al., 1998; Bowman et al., 2003). Kronenberg et al. (1990) were able to correlate an increase in water content of fluid inclusions in quartz grains with increases in strain within a small shear zone in granitic rocks from the Sierra Nevada. Similarly, Nakashima et al. (1995) measured water contents in deformed granitic rocks from the Yanazawa-Kamimura area near the Median Tectonic Line (MTL) and demonstrated an increase of water content in quartz from ~300 ppm to 2500 ppm toward the MTL with increasing deformation. Wawrzyniec et al. (1999), using fluid inclusions along the Simplon fault in the Alps, documented a correlation between water-rich zones and ductile deformation, and CO_2-rich zones and brittle deformation. They suggested that the different wetting characteristics of carbonic versus water-rich fluids on grain boundaries influenced the mechanical behavior of the rocks along the fault. Wetting behavior refers to the ability of a fluid to form a discrete phase along interconnected grain edges. Water-rich fluids exhibit wetting characteristics that allow for interconnectivity along pore spaces, while CO_2-rich fluids do not. Non-wetting behavior can result from the addition of even very low amounts of CO_2 (Watson and Brenan, 1987; Holness and Graham, 1995).

The conclusions of Wawrzyniec et al. (1999) are consistent with the experimental results of Post et al. (1996) and Post and Tullis (1998) that the creep strength of quartz does not depend solely on the amount of intragranular water but on the fugacity of water, probably through its effects on intrinsic and extrinsic, water-derived point defects. CO_2-rich fluids apparently decrease the permeability by exhibiting non-wetting behavior; this process decreases the available intragranular water that is important for reduction of creep strength.

Rheology of Fault Zones

Sibson's (1977) observations concerning the correlation between styles of localization (faults vs. shear zones) and deformation mechanisms became the basis for later work on the evolution of fault rocks (e.g., Anderson et al., 1983; Aydin and Johnson, 1983; Chester and Logan, 1986). More recently, the study of the evolution of fault rocks has been bolstered by scientific drilling programs such as the Cajon Pass drill hole (Zoback et al., 1988), SAFOD (the San Andreas Fault Observatory at Depth: Zoback et al., 2011), the drill hole into the Chelungpu fault in Taiwan (Ma et al., 2006), drilling into the Nojima Fault in Japan (Ohtani et al., 2001), the Japanese offshore drilling programs (e.g., Kinoshita et al., 2009), and the Alpine Fault (Townend et al., 2009). These ventures have sampled fault rocks directly from active faults and generated invaluable downhole geophysical observations. Key rheological debates concern the strength of fault zones, the frictional properties of fault rocks, and what indications of seismic behavior are preserved in fault rocks. For example, work in the Nankai Trough provides evidence for very large amplitude (~350 °C) shear heating at shallow (~400 m) depths below the seafloor; this zone is also characterized by strain localization and a cataclastic black gouge layer (Sakaguchi et al., 2011; Yamaguchi et al., 2011).

An important theoretical advance in understanding the rheology of fault zones has been the concept of rate and state dependent friction (e.g., Dieterich, 1992). The model allows materials to be simply classified with respect to their seismogenic potential into velocity strengthening and velocity weakening behavior. These characteristics can be determined in the laboratory, and the general factors that determine velocity strengthening versus velocity weakening are now understood (e.g., Marone 1998; Mair and Marone, 1999). Testing of fault gouge from the San Andreas fault zone suggests that velocity strengthening or weakening behavior of gouges, as modeled by the rate and state dependent friction law, correlates with creeping or seismic areas of the fault, respectively (Carpenter et al., 2012). However, rate and state dependent friction laws are essentially phenomenological (e.g., Niemeijer et al., 2012), and it is not clear if they apply to high-slip velocities.

Fault zones are commonly considered to have frictional strengths approximated by Byerlee's Law with typical static coefficients of friction of ~0.6 (e.g., Scholz, 2000; Collettini and Sibson, 2001), but frictional sliding experiments for a range of materials, including direct measurement of natural fault gouge (Carpenter et al., 2012), instead suggest that friction coefficients can be as low as 0.1–0.2, particularly where they contain layer silicates (e.g., serpentine, Reinen et al., 1994; bentonite, Shimamoto and Logan, 1981). Fabrics in fault gouge materials may play a critical role in weakening by creating an interconnected weak layer (e.g., Chester and Logan, 1986; Collettini et al., 2011), observations similar to those made in shear zones (e.g., Mitra, 1978, 1984; Simpson, 1985; Evans, 1988). Fault weakening is also attributed to solution-transfer creep (e.g., Gratier et al., 2011; Holdsworth et al., 2011). In situ stress measurements suggesting that the maximum principal stress is subperpendicular to the San Andreas Fault are taken to imply that the fault has a low strength (Zoback et al., 1987).

Fault weakening in a different sense may be caused by pore fluid pressure (e.g., Fulton and Saffer, 2009). The association between high pore fluid pressures and seismogenic faulting is the key feature of the very influential fault-valve model of Sibson et al. (1988), which has been widely applied to hydrothermal mineralization (e.g., Cox and Knackstedt, 1999). In the fault valve model, the mechanical properties of the fault remain strong (except during failure), and Faulkner et al. (2006) showed how the elastic properties of a damage zone around a fault may be changed by fracturing to allow high pore fluid pressures to exist without hydrofracture. Direct evidence for high pore fluid pressures in fault zones has not yet been found.

Fault weakening may also occur at dynamic rupture rates by such effects as thermal pore-fluid pressurization, flash heating, and melting (e.g., Kitajima et al., 2011). The presence of pseudotachylyte has been the "gold standard" for evidence of paleoseismic activity (e.g., Cowan, 1999), since the general acceptance that pseudotachylyte is formed by melting during meteorite impacts or by faulting (e.g., Sibson, 1975; Spray, 1997; Reimold, 1998). However, new suggestions are proposed for recognition of seismic slip rates based on (1) the recognition of fluidized textures in fault rocks (Otsuki et al., 2003; Smith et al., 2008); (2) pulverized textures (e.g., Wilson et al., 2005; Mitchell et al., 2011); (3) clay-clast aggregates, which are features produced in the laboratory and have been observed in natural fault zones (e.g., Boullier et al., 2009); and (4) petrologic evidence for localized heating (e.g., O'Hara, 2004).

Perhaps one of the most exciting discoveries about fault rheology in the last decade has been the realization that deformation in fault zones occurs on a wide range of time scales beyond those described by quasi-state and rate and state friction laws. A simple division between seismic and aseismic behavior is no longer adequate, and displacement rates must be described in a range from creep at tectonic rates to rapid, dynamic rupture during great earthquakes. The observational phenomena giving rise to these new insights include episodic tremor and slip, transient slow slip, slow earthquakes, silent slip, and "quasi-static" behavior, and some variety of these behaviors has now been detected in subduction zones (e.g., Rogers and Dragert, 2003; Gomberg et al., 2010), on normal faults in rifts (Calais et al., 2008), and within the San Andreas fault system (Shelly, 2010). It seems likely that fluids are integral to this behavior in all cases. Structural geology is now faced with the challenging task of integrating these observations with evidence from the geological record, and understanding the mechanics and deformation mechanisms underlying these observations (e.g., Fagereng and Sibson, 2010).

Other Rheological Methods

Field-Based Methods

Structural geologists have developed two approaches to extract rheological information directly from observations in naturally deformed rocks other than the use of microstructures to link through experimental deformation to flow laws: (1) relative deformation of adjacent materials through some measure of strain, or (2) dynamic instabilities. Ramsay (1982) provided the first coherent summary of the many ways that field-based structural geologists use mesoscopically and macroscopically observable features of deformed zones—and geometric relations of identifiable rock domains within them—to constrain rheology. Of particular note was the use of deformed polymictic conglomerates, which allow the relative viscosities of the different clast types to be determined. An extensive data set of deformed conglomerates collected by Czeck et al. (2009) suggests that the relative viscosities of different rock types are not always constant, even at relatively constant P and T conditions. Treagus (1983, 1988) demonstrated that cleavage refraction can be used to estimate viscosity contrasts across deformed but non-slipped layers.

Other methods were developed to evaluate the power law exponent of viscous materials. Hudleston and Holst (1984) used a combination of intra-layer strain and fold shape, whereas Hudleston and Lan (1994) focused on the curvature of the fold hinge to evaluate the power-law exponent of viscous layers involved in buckle folds; increasingly angular hinges require higher strain localization and thus higher power-law exponents. Masuda et al. (1995) noted that in particular cases the warping of foliation into porphyroclasts can provide information on power-law exponents.

Finally, it is possible to use buckle folding or boudinage to arrive at quantitative information about relative rheology based on dynamic instability analysis (e.g., Biot, 1961; Smith, 1975). For example, Tikoff et al. (2010) used the presence of fold wavelength–layer thickness relationship to evaluate the relative viscosity of folded orthopyroxenite dikes in an olivine-dominant matrix (and thus, orthopyroxene rheology and olivine rheology at upper mantle conditions). An important result from a combination of these field studies is that the relative rheology of different materials rarely varies by more than one order of magnitude (e.g., Lisle et al., 1983; Treagus and Treagus, 2002). Rheological estimates done over a variety of scales show similar variations (e.g., Horsman et al., 2008).

There are several important points to all of these field-based studies of rheology. A significant disadvantage is that these methods only provide relative rheology. However, by linking the observed features to microstructural analysis (and thus experimental deformation), quantitative constraints can be evaluated. There are, however, significant advantages to this approach. The largest is that field-based methods evaluate the rheology in polyphase rocks deformed at natural conditions (P, T, natural tectonic strain rates). Although experimental deformation has mostly been carried out on monophase materials, polyphase rocks are the rule rather than the exception in nature. In addition, experimental deformation requires a trade-off between temperature and strain rates. Increasing temperatures allows for experiments to be carried out on laboratory time scales (hours to weeks), so that data must be extrapolated to natural strain rates (Paterson, 2001). As described by Talbot (1999), it is relatively straightforward to provide qualitative estimates of relative rheology, based on field observations. Thus, coordination of field studies with experimental deformation is likely to be an opportunity for significant insights in the future.

Numerical Methods

Because crustal rocks are rarely monomineralic, there have been attempts to understand the rheology of polyphase rocks. To date, most of the work has been done on simple, two phase rocks, evaluating the rheology and the development of microstructures (e.g., Burg and Wilson, 1987; Handy, 1994). Much of this work grew out of S.M. Schmid's group, starting with a set of experiments published by Jordan (1987), which investigated foliation development in an initially nonfoliated, polymineralic system. Based partly on these experimental results, Handy (1990, 1994) developed a numerical model for the relation between strength and composition of bimineralic materials. In general, the relation is highly nonlinear and depends on three end-member mechanical and microstructural types: (1) the competent phase, forming a load-bearing framework around a physically isolated incompetent phase; (2) two or more relatively incompetent minerals controlling bulk rheology; and (3) competent minerals forming clasts in an incompetent matrix, with the incompetent material controlling the bulk rheology. The strength of this approach was its basis in the observed evolution of microstuctures during the deformation. Other experimental studies have been conducted on multiphase materials (e.g., Dell'Angelo and Tullis, 1996; Dresen et al., 1998; Bruhn et al., 1999; Ji et al., 2001), and they generally corroborate numerical models of two-phase or multiphase systems (e.g., Tullis et al., 1991).

One other approach has been to model microstructural deformation using advanced numerical models. Elle is a well-known example of this type of model (e.g., Jessell et al., 2001; Jessell and Bons, 2002). This 2D model simulates both microstructural development and metamorphic processes, including grain nucleation, grain growth, diffusion, and lattice rotations. Because experimental deformation at natural strain rates is not available, Elle and similar modeling programs provide an important avenue for future studies.

The Effect of Melt on Rheology

In the past 50 years, a significant amount of research in migmatitic terranes has drastically reshaped our understanding of these systems, owing in part to the critical effects of partial melting on rock rheology, and therefore geodynamic processes (Fig. 16). Arzi (1978) was the first to quantify the effect of partial melting on rock strength in confined experimental rock-deformation studies by varying the melt fraction and applying the empirical law of Roscoe (1952). His results showed that there exists a drastic decrease in rock viscosity at melt fractions between 10% and 30% melt, which was coined the *rheological critical melt percentage* (or RCMP). This fundamental contribution was confirmed and revised by subsequent experimental studies (e.g., van der Molen and Paterson, 1979; Paquet et al., 1981; Rutter and Neumann, 1995). Based on these experimental results, it is now assumed that rock strength is decreased by two to three orders of magnitude for high-grade metamorphic units undergoing anatexis, even for as little as 7% melt fraction (Rosenberg and Handy, 2005). A second decrease in rock strength describes the transition from dominantly solid behavior to dominantly fluid behavior, owing to the loss of continuity in the solid framework (Arzi, 1978; van der Molen and Paterson, 1979; Rushmer, 1996). Further work by Vigneresse et al. (1996) established the contrasting character in rock viscosity during melting versus crystallization. In the case of crystallization, partially molten rocks are stronger than those on the melting path for the same melt fraction, owing in part to the availability of melt and crystal interactions (Fig. 16).

This experimental work was immediately relevant to migmatites. In metatexites, which display plastic to viscoplastic behaviors, the microstructures and fabrics that develop are largely a function of strain rate (cf. Kohlstedt, 1995), resulting in the production of folds, magmatic shear zones, and mafic enclaves or boudins. In contrast, diatexites lose the continuity of their solid framework and deform in a manner similar to magma bodies at high melt fraction (>30%) (e.g., Shaw, 1972; Dingwell et al., 1993). Expansive regions of migmatitic and granitic complexes in exhumed orogens indicate that the deep continental crust typically undergoes partial melting during orogeny (e.g., D'Lemos et al., 1992; Malavieille, 1993; Brown and Solar, 1998a, 1998b; Ledru et al., 2001; and many others). Moreover, modern geophysical surveys in active orogenic settings (Tibetan Plateau–Himalaya, Nelson et al., 1996; Andean Altiplano, Schilling and Partzsch, 2001; Pyrenees, Pous et al., 1995) indicate that much of the orogenic crust contains a significant fraction of partial melt (~20 vol%) at mid- to lower crustal depths (>10–20 km). Given the rheological consequences of partial melting, migmatites likely represent weak zones that may facilitate the mobilization of low-viscosity, partially molten crust during orogeny (e.g., channel flow, gneiss dome formation) and/or aid in late orogenic collapse, thereby having profound effects on lithospheric strength and strain partitioning in the crust.

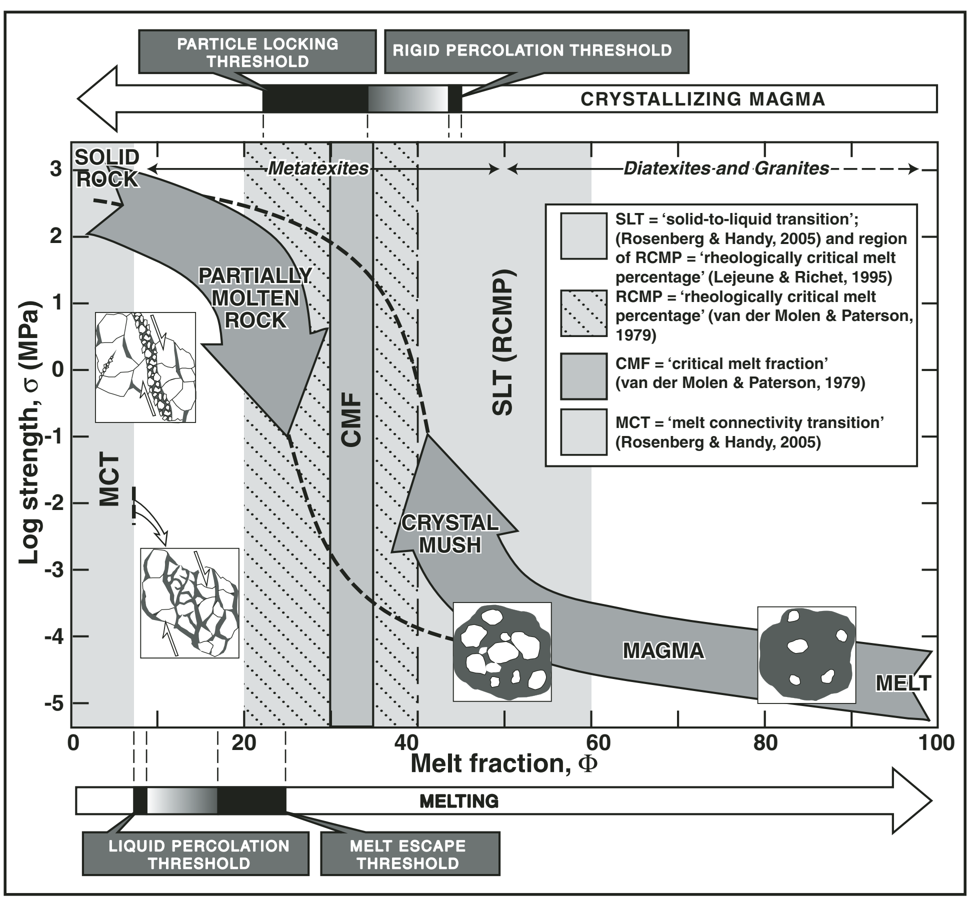


Figure 16. Synoptic diagram, illustrating the role of melt on rheology and critical thresholds associated with changes in melt fraction. Note that the material properties (i.e., rock strength–viscosity) vary by orders of magnitude and depend on whether a system is melting or crystallizing. Corresponding rock types and schematic rock microstructures are shown for clarity. Modified after Vigneresse and Tikoff (1999); Burg and Vigneresse (2002); Rosenberg and Handy (2005); Vanderhaeghe (2009); and references therein.

CONCLUDING REMARKS

We would like to reiterate that the topics chosen reflect the biases of the authors. Our choice not to address tectonic questions dictated that we forgo summarizing the major advances in understanding mountain belts, including those in the North American Cordillera and the Himalayas. Additionally, we have underemphasized the role that numerical models play in our understanding of deformation at all spatial and temporal scales.

In writing this review article, we were struck by several patterns. The first was the truly multidisciplinary nature of structural geology. It is increasingly common that many structural geologists are also major practitioners in at least one other geoscience discipline. Further, it is clear that practitioners are influenced by, and contribute to, fields outside of the geological sciences, including material science, engineering, and computer science. Another line of evidence for this broad nature is the cited publications. It was surprising to realize how many of the articles cited in this review are from general journals (e.g., *Geological Society of America Bulletin*), rather than specialty (e.g., *Journal of Structural Geology* or equivalent) journals.

Another insight in writing this article is to note that generalists tend to take the more historical approaches, while specialists tend to be more focused on processes. With increasingly sophisticated equipment and numerical approaches, specialization will continue to occur. There is, however, still a role for generalists, and the critical role that they play in science (including integration and synthesis). The structural geology community's ability to foster generalists is somewhat unique in the sciences, even among subdisciplines within the geological sciences. This factor also seems to have the additional, and possibly deleterious to the field, side effect of luring structural geologists into university administration, a role to which they seem disproportionally drawn. Further, it is also interesting to note that two prominent scientists who started their careers as structural geologists are best known for their work outside of structural geology: Walter Alvarez for his major contributions to understanding the cause of the K-T extinctions, and Paul Hoffman for his support and popularization of the Snowball Earth hypothesis.

Structural geology is at a crossroads, as are many sciences that have an applied component. There has clearly been a diversification away from studying mesoscale geological structures, ancient mountain ranges, and finite strain. Increasingly, research in structural geology has turned toward society-relevant outcomes, including hazards, natural resources, archaeology, and engineering or building materials. Moving forward, it seems that structural geology will grow in a variety of interesting directions, mostly requiring interactions with other fields. A non-exclusionary list follows:

1. Structural geology in active tectonic regions will explore the interaction of landscape evolution, climate, and tectonics.
2. The interactions between surface processes and deep-Earth processes, where the connection and long-term record is crustal deformation. This subject includes the role of mantle processes (hotspots, mantle drips) on lithospheric processes and structures, but also the possible role of the lithosphere in initiating mantle processes (e.g., hotspot initiation caused by lithospheric rifting).
3. The discrepancy between geodetic and geological rates, and how long-term geodetic deformation links to distributed deformation and mountain building.
4. The link between material science and structural geology. The increasing use of electron backscatter diffraction (EBSD), the scanning electron microscope (SEM), and the transmission electron microscope (TEM) among the structural geology community allows significant insight into the understanding of microscale deformation.
5. The interaction between structural geology and geochronology, with the role of mineral deformation and its effect on radiometric ages. For example, Reddy et al. (2006, 2007) have shown that deformation structures can affect dates in zircons (see also Moser et al., 2009). The use of geochronology for tectonic analysis will certainly continue, particularly with new methods of in situ dating, such as monazite "chemical dating" (Williams et al., 2007).
6. Structural geology and planetary geology. This connection has been active for the past 30 years, but will likely continue with increasing data from other planetary bodies and with advances in understanding impacts.
7. Structural geology and geological engineering. Geological engineering will increasingly become a societal priority, with applications in groundwater availability, waste disposal, carbon sequestration, etc.
8. Structural geology and sedimentology. This connection is as old as either of these subdisciplines, but the increased exploitation of nonconventional reservoirs or smaller (conventional) resources will necessarily cause closer interactions of these fields.
9. Thermodynamic approaches to structural geology. The exciting possibility of integrating deformation, stress, heat, fluid flow, and chemical reaction is offered by generalized thermodynamic approaches, which include non-equilibrium concepts (e.g., Hobbs et al., 2008, 2011).

There seems to be no decrease in interest or relevance in core structural geology, which is critical to advances in any of the above topics. Further, as long as structural geologists continue to concentrate on the rock record, there will inherently be a historical component to their analyses. The geometry and the inferred deformational history is, in some sense, a necessary prerequisite to addressing processes. Or, to put it colloquially: "You need to know what happened, to figure out how it happened, if you want to figure out why it happened." Which brings us back, finally, to the documentation of geometry, inferences of deformation history, and interpretations of dynamics (Fig. 1). This approach is the strength of our science, and this component of natural history is also why geology is not just a combination of biology, chemistry, and physics. So structural geology goes, with its balance between historical-based and process-based approaches, into the next century.

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