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Group Report: Fluids, Geochemical Cycles, and Mass Transport in Fault Zones

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OVERVIEW

This report considers the role of fluids and fluid–rock geochemical reactions that potentially influence earthquake behavior and fault zone strength throughout the lithosphere. Our discussions highlighted the distinctions between processes that occur within brittle, upper crustal fault zones (0 to ~10 km) where fluids can be pressured hydrostatically to lithostatically and deeper parts of fault zones (> 10 km) where fluids can be either near-lithostatically pressured or where a free fluid phase is absent. We recognized that fluid flow and geochemical processes vary with the style of faulting. We conclude that the comprehensive understanding of fluid–rock geochemical interactions in seismogenic faults is in its infancy. The major areas of future research can be grouped according to three thematic questions:

- What are the controls on fluid–rock chemical reactions in and adjacent to fault zones?
- How do fluid-flow processes change before, during, and, after earthquakes?
- What are the magnitudes of fluid fluxes in fault zones from the surface down to the asthenosphere and how do these fluxes vary with tectonic environment?

**THEME 1:
WHAT ARE THE CONTROLS ON FLUID–ROCK CHEMICAL
INTERACTION IN AND ADJACENT TO FAULT ZONES?**

Introduction

Several statements can be made regarding fluid–rock chemical interactions within fault zones: Faults and shear zones can be sites of both upward and downward flow of fluids, but they can also act as barriers for fluid flow (Bense and Person 2006). The effect of fluids on rock strength is one of the main reasons for strain localization. Shallower parts of faults tend to show more extensive hydration than deep parts. In the shallow parts, fluids are usually introduced from the surface. At deeper levels faults tend to separate blocks with different hydraulic heads (i.e., pressure compartments). This heterogeneous fluid–pressure structure may change as faulting progresses and the rocks move through different pressure regimes. Thus, fluid pressure within fault zones can fluctuate with time. At yet deeper levels, faults in sedimentary basins and hydrous crystalline rocks are potential sites of fluid escape, at least while the faults are active. Faults and shear zones undergoing retrograde metamorphism (e.g., involving chloritization of crystalline rocks) in nominally dry rocks require the introduction of a fluid phase. The cores of these faults become water-saturated and are lined with quartz or fully hydrated minerals. Rates of fluid production in metamorphism are much too low to account for such highly mineralized/altered fault zones. Transcurrent faulting, overthrusting, and subduction may result in the release and upward migration of fluids from deep levels of fault zones (Screaton and Saffer 2005; Saffer and McKiernan 2005). Aseismic or silent earthquakes may be driven by metamorphic fluid production. Different patterns of metasomatic alteration are associated with down-temperature and up-temperature flow systems in subduction zones.

Chemical Processes

There are a number of geochemical processes by which fluids affect fault-zone strength. Precipitation of phyllosilicate minerals, iron oxides, quartz/calcite cements, and other minerals along fault zones can either increase or decrease fault rock strength while reducing fault zone permeability. Precipitation reactions are potentially driven by a number of mechanisms, including temperature and pressure changes, fluid mixing, and fluid–rock reaction. However, little is known about how these fluids originate, and how they migrate to and from fault zones during the earthquake cycle. Integrated fluid volumes and fluid-flow rates in fault zones are poorly constrained. We know very little about the chemical saturation state of fluids and their pressure evolution in seismogenic zones. We know even less about the background permeability structure of deep crustal fault zones.

In rock masses exhumed from great depth where little or no fluid is initially present, vein minerals crystallized in fractures usually include chlorite which forms at $<250^{\circ}\text{C}$. This indicates that minerals do not begin to precipitate in cracks until rocks are exhumed to a depth of about 10 km due to slow reaction kinetics. There are some exceptions to this, for example, in shear zones where fracture-filling minerals appear to have formed in equilibrium at higher temperature conditions.

Chemical reactions (e.g., dissolution–precipitation and pressure solution) can cause the fault core and the damaged zone around this core to heal (Chapter 12). Within the fault core, hydrothermal alteration tends to be fast due to the relatively large surface area of fine, comminuted grains in cataclastic fault rock. Time scales of geochemical processes in the damage zone and the core vary widely due to differences in temperature history, grain surface area, and magnitude of fluid flow. Temperature changes affect reaction rate-controlled processes more than diffusion-controlled processes (pressure-solution creep and compaction) due to the lesser temperature dependence of the latter. Frictional heating at temperatures in excess of 500°C can occur locally in the fault core (see Chapter 7) leading to flash boiling close to the fault surface followed by rapid cooling to 100°C shortly thereafter. Fission-track analysis of zircon and apatite grains collected adjacent to fault cores could be used to determine whether frictional heating cycles occur and perhaps to establish the duration of these cycles (Murakami et al. 2002). Faulting can juxtapose mineral phases that are geochemically incompatible and lead to rapid reactions, (e.g., ultramafic rocks like serpentinites emplaced next to quartz-bearing pelites that react to form biotite, chlorite, and amphibole-rich zones).

Faults are usually open geochemical systems, at least on the time scale of a large earthquake cycle. Open system reactions entail the flux of water and dissolved chemical components to the reactive surfaces. On the other hand, pressure solution can occur in a closed system and only requires a small amount of fluid to progress.

For deep parts of faults undergoing viscous creep, the fluid-flow regime depends strongly on whether the rocks within and adjacent to the fault are undergoing prograde or retrograde metamorphism (Chapter 11). Prograde reactions create fluid and raise fluid pressure, potentially even leading to fluid overpressure in the case of rapid fluid production in low-permeability rocks. Evidence for lithostatic pore-fluid pressures is indirect and comes from fluid inclusions, mineral stability relations, and structures (e.g., Yardley 1989; Peacock 1990). In contrast, retrograde reactions generally consume water and therefore reduce fluid volume and mineralization rates. A common example for this is the formation of phyllosilicates at the expense of stronger solid mineral phases (e.g., feldspar in granitic rocks). Reactions like these have a profound effect on crustal rheology because the hydrous products are much weaker than the anhydrous (or less hydrous) reactant phases. In the following,

our group focused its discussion on several mechanisms including dissolution–precipitation, pressure solution, stress corrosion, phase transformations, and melting.

Dissolution–Precipitation

Dissolution–precipitation reactions are driven by chemical potential gradients. These can be induced by cooling or heating of fluids, or by fluids in equilibrium with one mineral assemblage in the crust coming into contact with another mineral assemblage in the fault zone. Both dissolution and precipitation occur on surfaces of mineral grains that are in contact with a fluid. Surface kinetics control the rate of the reaction. Dissolution–precipitation has been investigated in the laboratory (e.g., Rimstidt and Barnes 1980), but the influence of deformation on reaction rates is not well known. Likewise, the effect of the force of crystallization (Fletcher and Merino 2001; Hilgers and Urai 2005) on fault zone strength has not been considered to date. The rate of advective transport of reaction products away from reactive surfaces places an important limit on reaction rate, which in turn often governs the precipitation rate. Mineral growth requires new chemical components to be transferred in a fluid. Surface kinetics are easier to investigate than fluid transport processes. The solubility of aluminum- and silica-bearing minerals increases with pressure and depth. However, increasing chloride concentrations slightly decreases the solubility of aluminum and silicate and therefore hinders their transportability in a fluid phase. Hofstra and Cline (2000) have shown that calcite dissolution can enhance the permeability of faults in carbonate-hosted ore deposits. There, carbonate rocks contain calcite vugs replaced by silicates. Yet, as carbonate rocks dissolve, CO₂ enters the fluid phase and reduces the solubility of all minerals. Many in our group felt that dissolution and precipitation reactions are by far the most important mechanisms for mineral transformation, an opinion that is largely based on isotopic evidence for large scale isotopic exchange (Mulch et al. 2004). However, our understanding of how these geochemical reactions are coupled with deformation is incomplete. The following hypothesis can be tested in well-exposed, exhumed (fossil) fault zones, and may lead to better constraints on the role of fluids in fluid–rock chemical reactions in fault zones.

Hypothesis 1: Downward-flowing fluids in fault zones enhance dissolution reactions, whereas upward-flowing fluids favor precipitation reactions.

Test: Collect geochemical, isotopic and thermochronometric data from exhumed fault zones to determine whether there are monotonic cooling and heating trends. Establish the degree of fluid–rock isotopic exchange within and away from the fault zone. Develop numerical models of fluid–rock isotopic exchange to constrain the depth of fluid circulation, the temperature history, and the duration of the flow systems.

Pressure Solution

Pressure solution is driven by stress-induced gradients in chemical potential and by diffusion-controlled reaction rates that make new chemical components available along the fault zone. There is a fundamental difference between pressure solution at monomineralic grain boundaries and at polymineralic boundaries. In rocks containing both mono- and polymineralic layers, the monomineralic layers that underwent pressure solution always appear to be less deformed (boudinaged) and therefore stronger than the adjacent polymineralic layers (Chapter 12). An explanation for this is that stressed boundaries of like mineral phases (e.g., quartz-quartz, calcite-calcite, feldspar-feldspar) behave as single, large crystals, especially if their crystallographic orientation is similar. In contrast, stressed boundaries of different mineral phases (and other less-soluble species like mica and illite) are likely to remain open to diffusive mass transfer, allowing pressure solution to progress (Hickman and Evans 1995).

There is clear petrographic evidence of pressure-solution reactions in active fault zones. In the post-seismic interval immediately after an earthquake, pressure solution is important for several reasons: First, coseismic micro-fracturing within the fault zone reduces grain size, enhancing diffusion and pressure-solution creep, and in turn relaxing any residual stress in the fault system. Pressure solution does this in part by producing mineral fabrics whose preferred orientation is parallel to the fault plane, thus reducing the shear strength of rocks. Second, pressure solution can produce chemical components that seal coseismic fractures, leading to hardening. Third, pressure solution is an efficient compaction mechanism, especially in fine-grained fault gouge. Compaction can lead to hardening, but also to a decrease in permeability and therefore to weakening if fluid is produced within the fault system. The weakening due to fracturing and dissolution is faster than the strengthening associated with healing and sealing of the fault, and the associated increase of fluid pressure (Chapter 12).

To be effective, pressure solution requires the presence of a fluid phase. Small amounts of trapped fluid forming thin layers on the scale of nanometers are sufficient for pressure solution to be rate competitive with other deformation mechanisms. Pressure solution is affected by the following factors: (a) kinetics of mineral dissolution at highly stressed grain contacts; (b) precipitation rates at fluid–mineral interfaces; (c) catalytic effects of minerals (e.g., phyllosilicates); (d) mass transfer properties of trapped fluid paths under stress; (e) the nature of fluids and minerals present; (f) presence and interconnectedness of the fluid phase; (g) the size of the system in which the transfer process (diffusion, nm to dm; advection, hm to km) operates; and (h) temperature.

The thermodynamics of pressure solution are reasonably well known (e.g., Gibbs 1877; Paterson 1973; Lehner 1995; Shimizu 1995), however, the

thermodynamics of pressure solution in general 3D stress states needs further development. Laboratory measurements of pressure-solution rates (e.g., Rutter 1976; Gratier and Guiguet 1986; Hickman and Evans 1995; de Meer et al. 2002) allow us to make order-of-magnitude predictions of these rates. A major problem is that the characteristic time of deformation by stress-induced mass transfer is several orders of magnitude less than the average human life-span. Acceleration of pressure-solution rates in experiments renders some geological interpretations of extrapolated experimental results uncertain. Another major factor that can only be inferred from the analysis of natural fault zones is the relative contributions of advection and diffusive modes of fluid flow. Related to that is the mean distance of mass transfer within fluid-flow zones (see also Chapter 11). Evaluating the nature of the fluid and the minerals at depth along the fault is also an important issue.

Mineral Reactions (Hydration and Dehydration)

Mineral reactions affect rheology principally through the variations that they induce in fluid content, fluid pressure, and lithology in a rock. Yardley and Baumgartner (Chapter 11) point out that rocks generate fluid and become water-saturated throughout their heating cycle (prograde metamorphism), but dry out rapidly through retrograde absorption of water once they begin to cool (retrograde metamorphism). Seismic low-velocity zones detected in geophysical surveys have been interpreted as water-rich zones in some regions currently inferred to be undergoing metamorphism, for example off the coast of Vancouver Island, as imaged by the Lithoprobe traverse (Hyndman 1995) and beneath the Southern Alps, New Zealand (Koons et al. 1998; Wannamaker et al. 2002; Upton et al. 2003; see section below on *Key Observations of Fluid Flux in Convergent Settings*).

Prograde mineral reactions commonly involve the release of fluid and are accompanied by changes in volume and grain size. A major rheological consequence of prograde reactions is likely to be local embrittlement, because an increase in fluid pressure through volatile release reduces effective normal stress. Most fluid production occurs stepwise during a series of devolatilization reactions, so it is likely that deformation mechanisms facilitated by the presence of a water-rich fluid (pressure solution or grain-boundary slip) are possible throughout the prograde cycle. Exceptional behavior is possible in carbonate rocks, where the solid volume changes accompanying the breakdown of carbonate minerals may lead to transient porosity, leading to focused fluid flow and the production of skarns.

Retrograde reactions can have two opposing effects. Water consumption can drive water fugacity to just a few bars, potentially inhibiting material transport along grain boundaries even if individual grains are hydrolytically weak

by remaining dissociated water in defect centers. As a result, dry rocks are very strong, retaining old textures and fabrics where wet rocks in the vicinity undergo extensive deformation in response to the same stress field. However, the presence of water can lead to the production of intrinsically weak, fine grained reaction products, especially phyllosilicates, which focus deformation and even favor superplastic behavior.

Some large-scale shear zones contain abundant microstructural evidence for reaction softening assisted by microfracturing in the presence of water. Fracturing and dilatation opens the rock system to episodic fluid infiltration (e.g., Sibson 1988; Bauer et al. 2000) whereas reaction softening obviously localizes strain (Chapter 12, Figure 12.4).

Based on studies of pressure solution and mineral reactions, we advance the following hypotheses to better constrain the role of fluids in fluid–rock chemical reactions in fault zones.

Hypothesis 2: Fluid–rock and mineral interactions affect interseismic permeability and strength, and thus control the recurrence time of earthquakes.

Tests: Determine the reaction rates both in the lab and in naturally deformed, exhumed fault rocks as a basis for formulating macroscopic numerical models that include strain-dependent transport properties and rheologies. Then, test this model with geophysical and geochemical measurements of fluid flux during interseismic periods. Such measurements (involving GPS, inclinometry, seismology, microgravimetry, chemical analysis of springs) may yield insight into how fluid transfer, porosity, permeability, and rock strength evolve at depth. The aim is to incorporate permeability and fluid pressure in empirical rate- and state-dependent constitutive laws. This would yield insight on how fluid–rock interaction “restores” rheological and transport properties during interseismic periods, setting the stage for renewed seismicity.

Stress Corrosion

Research on stress corrosion in rocks was active in the 1970s and 1980s (reviews of Atkinson 1982, 1984, 1987), but has stagnated since. Stress corrosion is controlled by reaction rates, cracking, and rate- and state friction. Subcritical crack growth at grain contacts in fault zones can lead to a time-dependent increase in contact area. Experiments on the rate- and state-friction of olivine reveal a component of crystal plasticity at grain contacts, even in the absence of water. Still missing in many investigations of experimentally and naturally deformed fault rocks are detailed studies of the interplay of fracturing and intracrystalline plasticity at grain boundaries. The search for stress corrosion should involve looking at the leading edges of cracks, for example, with TEM or laser interferometric microscopy.

One of several possible explanations for the origin of rate and state friction (discussed at length in Chapters 5 and 7) is that the real area of contact during frictional sliding increases due to subcritical crack growth. The idea is that if static contacts enlarge with time, or equivalently in the sliding case, enlarge if the slip velocity is lowered, the frictional coefficient will be larger because there is more contact area over which the resistance occurs. Thus, the total amount of bonding across the surface will be larger, thereby increasing resistance. This explanation is plausible because an increase in the area of contact with an increase in normal stress is the accepted explanation for friction, namely the ratio of normal stress to shear stress. One way the contacts might enlarge is by limited dislocation motion at highly stressed grain contacts. Another way is that contacting grains fracture slowly, involving subcritical crack growth. Although fracturing of the contacts might seem to weaken rocks, it could also increase contact area and cause increased friction, leading to strengthening. This explanation is consistent with observations that both subcritical crack growth and changes in the coefficient of friction occur in wet rather than dry environments.

Melting

From a mechanical standpoint, melt is often considered to be a fluid inasmuch as it exerts a pore-fluid pressure that counteracts normal stress and therefore reduces effective pressure. An important distinction between water-rich or CO₂-rich fluids and melts is that melts crystallize completely. When fluids cool, the volume (mass) of the dissolved phase removed by precipitation is small compared to that of the fluid volume (mass). The effect of melt on the deformation mechanisms of solid aggregates is probably similar to compositional effects on fluid viscosity observed in many less-viscous fluids. For example, a transition from dislocation creep to diffusion creep has been observed in experiments by adding a few percent (3 to 5%) melt to fine-grained (<10 μm) granitic samples (e.g., Dell'Angelo et al. 1987) and by adding ~0.8% water to feldspathic aggregates (Tullis et al. 1996). This transition is also inferred to occur in natural fault zones, even for large grain sizes (>300 μm) typical of natural mylonitic fault zones containing a melt (Rosenberg and Berger 2001). Strength reduction due to an increase in pore-fluid (-melt) pressure during melting has also been observed in aplite samples experimentally deformed in the presence of 5 to 10% melt (Dell'Angelo and Tullis 1988).

An additional similarity in the way fluids and melts may affect the deformation of crustal rocks concerns the localization of deformation within viscously deforming aggregates. The presence of a few volume percentages of fluids and/or melts in a viscously deforming aggregate can induce the transition from homogeneous- to highly localized deformation along interconnected networks

of fluid- or melt-bearing shear bands (Bauer et al. 2000; Rosenberg and Handy 2000). Analogue experiments indicate that localization coincides instantly in space and time with the introduction of fluids into the deforming aggregate. More rheological data is needed to estimate the effect of this transition on the strength of real rock.

The residence time of melt in a rock can be much longer (10^7 yr) than for metamorphic fluids. This is probably due to the higher density and viscosity of melts with respect to aqueous fluids, reducing the mobility of melts compared to other fluids. The long residence time of melts in the lower crust may not only affect fault nucleation, but the geometry of entire faulted domains, even orogens (e.g., Beaumont et al. 2001; Chapter 13, this volume). When present, melt volumes in the crust are generally much higher than fluid volumes. This conclusion is based on the debatable interpretation of seismic low-velocity zones as representative of high melt fractions. From a rheological standpoint, however, the critical question is not just “How much melt?”, but “Is the melt interconnected?” (see Chapters 4 and 7). Inferred melt percentages at a depth >20 km beneath the Tibetan Plateau in the Himalayas and southern Altiplano (Puna) Plateau in the Andes are $\sim 20\%$ (Schilling and Partzsch 2001). In this case, it is not certain what length scale of melt interconnection is represented by this low-velocity zone. A rheologically critical amount of melt (3–7 % by volume) could exist in much smaller amounts, e.g., on the grain scale, without its being observable with existing imaging methods. Within exposed fossil orogenic migmatites, the inferred melt volumes commonly range from 10 to 40% (Teyssier and Whitney 2002). The strength of a shear zone is nonlinearly proportional to the melt volume it contains and the strength versus melt fraction relationship probably follows an exponential law (Chapter 13). However, no experimental data exist yet to reliably constrain the mechanism and amount of melt-induced weakening in the viscous creep regime. The presence of water in the melt phase can also change the rheology of the melt-rock system. Water enters the melt phase above 650°C in granitic rocks. The melting temperature can be as low as 500°C when water is introduced. Formation of silica gels can also occur in fault zones at temperatures in excess of 200°C . An important question that needs to be resolved regarding the role of melts on fault dynamics is addressed with the following hypothesis:

Hypothesis 3: The relationship between melt content and viscous strength of shear zones follows an exponential law.

Tests: Deformation experiments on partially melted crustal rocks in the viscous deformation regime (dislocation and diffusion creep regimes) must be conducted with different melt contents (especially at very low melt fractions, $<3\%$) in order to obtain a relationship relating strength to melt fraction.

**THEME 2:
HOW DOES FLUID FLOW CHANGE
BEFORE, DURING, AND AFTER EARTHQUAKES?**

Introduction

Fluid pressure has significant effects on the mechanical and chemical behavior of the crust and fault zones, because it can change dramatically on different time- and length scales throughout the earthquake cycle. The earthquake cycle can last from hundreds of years along major plate boundary faults to thousands of years in intraplate environments. During interseismic intervals, diffusive processes in surrounding stable crust relax any overpressures, leading to hydrostatic fluid pressures. In low-permeability fault zones, however, pore pressures can rise through a variety of relatively sluggish mechanisms such as pressure solution, shear-enhanced compaction, chemical osmosis, or dehydration of water-bearing clays. Quantification of pore-pressure changes due to these processes is difficult because the conditions at depth that control them are still not sufficiently constrained. Crack healing and mineral precipitation can also increase pore pressure by drastically reducing the fault and damage zone permeability between successive earthquakes.

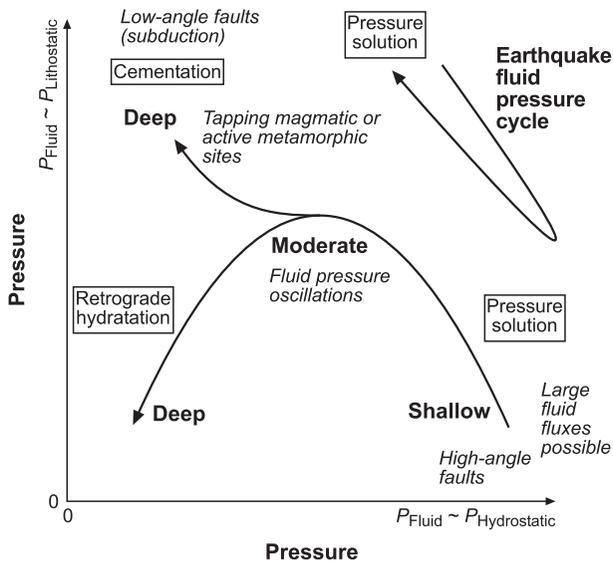


Figure 14.1 Schematic diagram showing the range of fluid pressures and fluid–rock geochemical interactions at different depths (shallow, moderate, and deep). Dry conditions are represented at the origin, water-saturated conditions at the top and right, with hydrostatic and lithostatic fluid-pressure environments represented by the x - and y -axes, respectively.

There is a wide range of possible fluid-pressure regimes within seismogenic zones of the shallow and deep crust (Figure 14.1). We argue here that each of these regimes should produce characteristic seismogenic features such as pseudotachylites, low-angle normal faults, and postseismic, limited-magnitude earthquakes, known as aftershocks. If deep crustal rocks are exhumed and undergo retrograde mineral reactions, rocks dry out and become strong. This is conducive to the formation of pseudotachylite (region labeled 'Deep' in Figure 14.1). Supracrustal rocks undergoing burial attain near-lithostatic pressures if permeabilities are sufficiently low (10^{-20} m²; upper left portion of Figure 14.1). Examples of this include both the deep parts of oilfields, as well as the cases from New Zealand and Vancouver Island noted above. However, if rocks are more permeable ($>10^{-16}$ m²), they are hydrostatically pressured (region labeled 'Shallow' in Figure 14.1) and may be flushed by surface fluids, depending on the permeability structure of crustal rocks with depth and within the fault zone. We would expect that these environments would host high-angle faults. Examples of this include the Rio Grande Rift, U.S.A. (Mailloux et al. 1999), the Rhine Graben, central Europe (Smith et al. 1998), and the Basin and Range, U.S.A. (McKenna and Blackwell 2004). These different environments give rise to different styles of fluid flow (escape, infiltration, or circulation) and hence of mineral reaction. It is clear that fluid-rock geochemical interactions need to be integrated into hydrothermal and mechanical descriptions of earthquake cycles.

There is evidence for hydrostatically pressured saline fluids in some deep boreholes. For example, at the German KTB borehole hydrostatically pressured fluids have been observed at 9 km depth (Moller et al. 1997). This is quite old water and raises questions about its chemical evolution and salinization mechanisms. One possible mechanism involves hydration reactions, whereby water is removed from the system and incorporated within retrograde minerals (e.g., phyllosilicates). This is augmented if Cl-bearing minerals, such as high-temperature biotite, are replaced by chloride-free phases such as chlorite. However, a number of other mechanisms, including the ingress of brines from former overlying sedimentary basins and the release of brine from fluid inclusions, have been proposed (Fritz and Frape 1987).

Various mechanisms for generating anomalous pore-fluid pressures have different characteristic time- and length scales of operation. For example, Miller et al. (2004) recently proposed that the coseismic release of high-pressure fluids at depth can initiate a pressure pulse that may propagate at a rate of up to a kilometer per day. This pressure pulse can trigger aftershocks by reducing the effective normal stress of incipient slip planes. Normally pressured fluids flowing within high permeability environments may flow at rates of up to about 10 m yr⁻¹.

Fluid flow is controlled predominantly by permeability, and permeability may change drastically at the onset of seismic rupture (e.g., Rojstaczer and Wolfe 1994). Just prior to an earthquake, the fault zone is most likely sealed

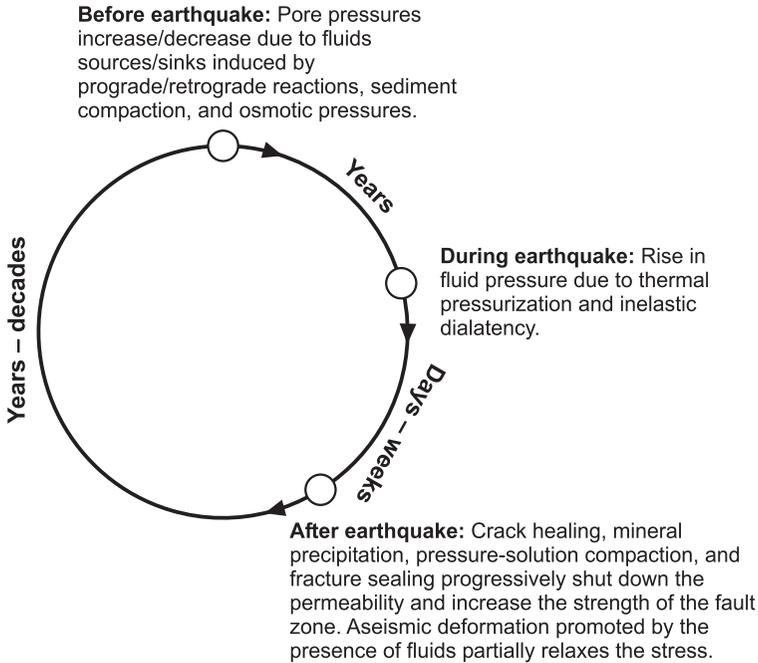


Figure 14.2 Schematic diagram indicating processes affecting pore-fluid pressure during the earthquake cycle.

from the damaged surroundings by crack healing and other long-term, time-dependent processes that operated during the previous earthquake cycle. The temporal evolution of pore-fluid pressure in fault zones can be thought of in terms of the time cycle of earthquakes (Figure 14.2). Prior to an earthquake, fluid pressure can rise due to inflow of fluids from depth, prograde metamorphic reactions, sediment compaction (in sedimentary basins), and possibly also osmotic pressurization provided there is a build up of salinity within the clay-rich core zone (salinity could increase due to hydration reactions in the fault gouge). Increases in osmotic pressure (Neuzil 2000) in shallow fault zones (<5 km) have not received much attention and we therefore propose the following hypothesis to study this mechanism.

Hypothesis 4: Relatively shallow earthquakes are triggered by the buildup of high pore-fluid pressures in a clay-rich fault core due to the generation of osmotic pressure.

Tests: Conduct ring-shear experiments in which salts are introduced to the clay-rich core of an experimental fault zone. Determine if this procedure reduces the frictional coefficient during shearing. In the field, drill a seismogenic fault zone

and sample pore fluids within the clay-rich core of the fault zone. If the core has elevated salinities and low porosity, this is consistent with osmotic pressure generation. Conduct numerical experiments using realistic hydration kinetic rates and osmotic pressure terms included into the groundwater flow equation (e.g., Neuzil 2000).

During an earthquake, fluid pressure can change dramatically due to inelastic dilatancy and thermal pressurization. Once a rupture begins, frictional heating of trapped pore fluids can raise the pore-fluid pressure to near-lithostatic values, thus weakening the fault. However, this process is self-limiting because lithostatic fluid pressures imply near-frictionless faulting, which in turn reduces frictional heating (Mase and Smith 1987a, b). Dynamic rupture across a region of fault overpressure is not yet understood in the context of the rate- and state formulation of friction, so this needs much additional study. Dilatant rupture causes pore-pressure dissipation both parallel and perpendicular to the fault zone. Based on known kinetic rates of precipitation/dissolution reactions, the rates of chemical reaction during earthquakes are probably small. However, frictional heating can generate silica gels rapidly, as discussed in Chapters 5 and 7.

When considering the dissipation of pore-fluid pressure, it is important to consider how pore-pressure changes both within the low-permeability, clay-rich core of a fault, and the higher permeability damage zone within adjacent basement rocks. Seismic rupture can reset the hydraulic properties of the system and any postseismic flow is controlled by the new permeability structure. At the crack tip during dynamic rupture and at fault bends, significant tensile stresses are developed just adjacent to the fault, providing a mechanism for the creation of a highly fractured damage zone surrounding the slip surface. Calculations and geophysical imaging indicate that the width of the damage zone diminishes with depth, from about 1–100 m near the surface to much less at midcrustal levels (~10 km, Chapter 2). The damage zone plays an important role in the postseismic redistribution of fluid pressure because it represents a highly permeable crack network that can transport fluid and mitigate fluid pressure over short time periods. Once earthquake rupture ceases, the heavily damaged and fractured region near the slip plane controls both the rate at which rupture-induced overpressures subside and the rate at which trapped pockets of overpressure dissipate. If there exist locally overpressured compartments with high permeability and earthquake-related fractures access these pockets, then the magnitude of the fluid-pressure pulse can be significant. The ensuing pressure pulse will exceed any local changes in shear and normal stress, thus providing a viable mechanism for the generation of aftershocks.

The build-up of osmotic pressures could be an important mechanism for generating earthquakes at shallow depths (<5 km) if porosity is low (<10%) and there is an order-of-magnitude increase in salinity within the fault zone and the clay-rich core (Neuzil 2000). Elevated fluid pressures can clearly occur within the fault core zone. An important issue centers on how fluids feed into

faults at great depth during the interseismic period. Depending on the fluid source, it may be necessary to break some sort of seal at midcrustal levels (10 km). This seismic zone is expected to widen toward the surface (Chapter 2).

Earthquakes can be triggered by changes in the Coulomb Failure Stress (ΔCFS):

$$\Delta\text{CFS} = \Delta\tau + \mu(\Delta\sigma_n + \Delta P_f) \quad (14.1)$$

where $\Delta\tau$ is the shear stress change on incipient fault planes, μ depicts the friction coefficient, $\Delta\sigma_n$ is the normal stress change on incipient fault planes, and ΔP_f represents pore pressure change.

If dynamic rupture of an earthquake taps into a source of high-pressure fluids at depth, then the propagating pressure pulse can have a significant effect on ΔP_f , potentially much more than on $\Delta\tau$ or $\Delta\sigma_n$. This hypothesis can be tested by accurate measurements of the space-time evolution of aftershocks. If this mechanism is important for aftershock generation, then a correlation should be found between aftershock evolution and predictions from a model of this process (Miller *et al.* 2004). Some data is already available for analysis (e.g., San Andreas fault, North Anatolian fault), but much more data is needed from a variety of tectonic environments.

In the postseismic interval immediately after a large earthquake, chemical reactions driven by residual stresses or by other earthquake-related processes (fractures, fluid transfers, temperature and pressure changes, etc.) progressively restore the strength and mass-transfer properties (i.e., permeability) of the gouge and damage zone to values that prevailed in the interseismic period up to the earthquake (Chapter 12).

Time and Length Scales of Fluid Flow and Geochemical Reactions in Fault Zones

Processes like pressure solution operate at the grain scale, whereas precipitation/dissolution reactions can occur at the km scale, linking areas of mineral dissolution and precipitation. Dehydration reactions occur over a very specific temperature range. At temperatures below about 300°C, reaction kinetics can be very important. At higher temperatures, thermodynamic equilibrium can be maintained provided there is ample fluid to facilitate material transport between crystals. Grinding up mineral grains within the fault core drives chemical reactions faster, primarily due to the increase of surface area with respect to volume. Kinetics may play an important role in the history of the earthquake cycle. Retrograde reactions can consume water and increase salinity, but the kinetics of these reactions suggest that they continue long after the earthquake. Reaction rates are strongly temperature-dependent and reaction times on the order of days to many tens of years seem probable (Wood and Walther 1983).

Hydroseismicity and Surface Expressions of Fluid Flow

It is well known that natural and anthropogenic changes in pore-fluid pressure can induce seismic events (e.g., Saar and Manga 2003). Following an earthquake, pore space collapses (Muir-Wood and King 1993) and the resulting increase in matrix permeability (Rojstaczer and Wolf 1994) has been proposed as a mechanism to explain changes in surface water discharge as the rocks adjust to changes in the stress field. Changes in surface water chemistry have also been observed following earthquakes (Curry et al. 1994). While fluid flow is known to occur during the earthquake cycle, the mechanisms driving fluid flow during seismic events remain poorly understood. Below we propose the following hypothesis which may shed light on this issue.

Hypothesis 5: Seismicity is triggered by fluid movement within and around fault zones. Earthquakes can tap mid- to lower-crustal reservoirs of high-pressure fluid, generating a fluid-pressure pulse and triggering aftershocks by suddenly reducing the effective normal stress on incipient slip planes.

Test: Analyze the space-time evolution of aftershocks to determine precisely the hypocenter locations. The spatio-temporal patterns of hypocenters will have predictable patterns that depend on tectonic setting. These should be correlated to surface flow features, including changes in aquifer geochemistry, subtle land deformation features associated with pressurization of the upper crust and depressurization of the lower crust as revealed through geodetic (InSAR and GPS) surveys, changes in seismic velocity associated with high pore-pressure zones, and possible changes in electrical conductivity.

Porosity and Permeability within Fault Zones

Studies within petroleum basins have generated a large body of knowledge regarding pore-fluid pressure distribution and the hydraulic properties of faults in the shallow crust. In soft shales or clays, fault zones act as barriers to fluid flow (e.g., Knipe 1997; Caine et al. 1996; Bense and Van Balen 2004). This is evidenced by pressure changes or compartments that have been observed to develop in fault-bounded oil reservoirs in the North Sea and Gulf of Mexico. On the other hand, petroleum basins are replete with examples of vertical migration of petroleum along fault zones (Rinkbeiner et al. 2001). Evidently, the relative permeability of fault rocks can evolve through time, especially during and after movement. In some situations, the main effect of faulting on fluid flow is to juxtapose high- and low-permeability units (e.g., Mailloux et al. 1999).

Despite several attempts to quantify the permeability of the upper crust (e.g., Manning and Ingebritsen 1999; Saar and Manga 2003), little is known about the hydraulic properties of fault zones beneath the sedimentary pile. There, the only suggestion of fluid flow comes from indirect measures such as the

isotopic alteration of fault rocks (Mulch et al. 2004), the genesis of fault-hosted ore deposits (Sibson 1989), and thermochronological and numerical modeling (Mailloux et al. 1999). To date, studies of fault mineralization have rarely been undertaken with seismogenesis in mind. Due to the relatively low permeability of wall rocks, faults in basement rocks are much more likely to act as fluid conduits rather than barriers to flow, as evidenced by the common occurrence of ore deposits along and adjacent to faults. However, Sibson (1989) pointed out that large ore deposits are commonly associated with relatively low-displacement faults, perhaps because the roughness of the fault plane permits it to act as a valve. Fluid flow may take place through tubes opened at jogs from the time of displacement until the permeability has been resealed by mineral precipitates. This should be tested by careful monitoring of fluid flow in the vicinity of jogs along active faults, for example, in the Walker Lane area of the Great Basin, Nevada, U.S.A.

THEME 3: WHAT ARE THE MAGNITUDES OF FLUID FLUX THROUGHOUT THE LITHOSPHERE IN DIFFERENT TECTONIC ENVIRONMENTS?

Introduction

Some information regarding fluid fluxes in fault zones is already available from drilling campaigns in European and North America. For example, the Corinth drilling project sponsored by the European Union (www.corinth-rift-lab.org/index_en.html) has drilled an active fault system to a depth of about 2 km in an extensional setting in Greece. Pressure solution was found to be an important mechanism controlling deformation (compaction and sealing) in this carbonate-hosted fault zone. Other sites offer additional data: Kobe (Japan), Chichi (Taiwan), and Parkfield (California, U.S.A.). At Parkfield, boreholes have been emplaced to monitor active processes on the San Andreas fault.

Key Observations of Fluid Flow in Rift Zones

Fluid inclusions from fault-bound ore deposits in extensional tectonic settings (e.g., Irish Pb-Zn) indicate that fluids were at near-hydrostatic pressures. The ore deposits formed in the damage zone where there is a source of sulfide-rich waters, suggesting that the damage zone is very permeable. Interestingly, big ore deposits do not normally occur directly along the slip planes of large faults (Sibson et al. 1988). Biological processes may have enhanced the formation of ore precipitation in the Irish deposits, providing a source of reduced S to precipitate metals from hydrothermal solutions (Blakeman et al. 2002). Given the likely importance of damage zones for fluid mixing in the

upper continental crust, it is probable that detailed fluid inclusion studies could further constrain flow models by providing information about fluid pressure fluctuations.

Orogenic gold deposits at midcrustal levels are more controversial (Chapter 11). Perhaps the fluids associated with these gold deposits come from prograde metamorphic reactions. Alternatively magmatic fluids or even high level waters may be responsible.

Key Observations of Fluid Flux in Convergent Settings

Fluid flow associated with active, prograde metamorphism has been imaged as a zone of anomalously high conductivity beneath both the Southern Alps of New Zealand (Figure 14.3) and Vancouver Island, Canada. High-conductivity zones in the vicinity of Nanga Parbat (Himalayas) can also be interpreted as a zone of melts (see discussion above). The Southern Alps is a remarkable system in that rapid convergence and erosion has brought rocks to the surface that are inferred to have passed through the shallowest of the metamorphic zones within only the past 1–2 Ma.

The U-shaped region of high conductivity within the lower crust in Figure 14.3 is interpreted to be a region of fluid production and strain-induced permeability (Upton et al. 2003). Several fluid-flow paths connect this U-shaped region both with the surface and with sources at depth. Each path may have a different cause: topographically driven meteoric flow in the upper crust around the Main Divide, fluid expulsion from the lower crust, interconnected fluid including meteoric and basinal components corresponding to the eastern vertical conductor which may be undergoing buoyancy-driven flow through the upper crust and episodic upward flow in the Main Divide region.

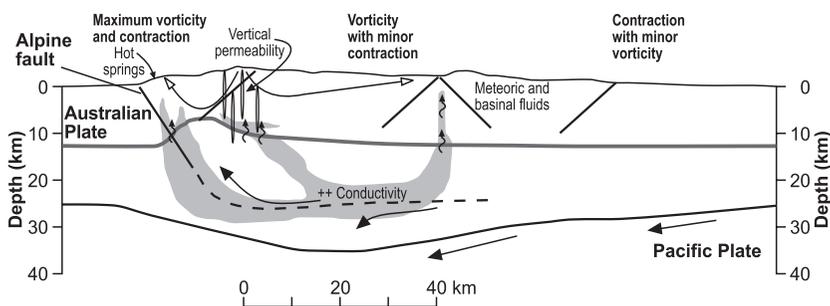


Figure 14.3 Schematic cross section of the Southern Alps and South Island, New Zealand, showing inferred paths of fluid flow based on conductivity studies (Upton et al. 2003). West is to the left, and the Alpine Fault is the prominent steep fault to the west of the topographic culmination (Main Divide). Strain regions (bold arrows) are taken from Koons et al. (1998).

SUMMARY

What Have We Learned?

To date, very little is understood about the role of fluid–rock chemical interactions, the mechanisms that drive hydrochemical fluid flow, and the magnitude of fluid flux in seismogenic zones. Even less is known about how these factors and processes interact during the earthquake cycle. Fluid–rock geochemical reactions have received considerable attention by metamorphic petrologists and geoscientists studying sedimentary basins, but not from a seismogenic perspective. Given the complexity of fluid–rock geochemical interactions (e.g., pressure solution, dissolution–precipitation reactions), the mineralogy in different fault zones, and the thermobarometric history during faulting (burial versus exhumation), few generalizations can be made regarding this topic. It is critical to develop well-posed, testable hypotheses that can better constrain the role of fluid–rock chemical reactions within the framework of the seismogenic cycle.

Our understanding of the role of pore fluids and associated geochemical interactions in earthquake dynamics would be greatly improved with the realization of advances in isotope geochemistry, hydrogeology, reaction kinetics, and numerical modeling. In addition, integrated field campaigns are needed in different tectonic settings. These would include working out the kinetics of deuterium isotope exchange for different minerals at elevated temperature and pressure, constraining the effective surface areas of fault zones, and developing a testable model of fault permeability and permeability anisotropy for faults in crystalline and sedimentary rocks.

In rock-dominated systems, geochemical markers are overwhelmed by the local wall–rock signature. Whereas oxygen isotopes are often useless as fluid tracers (because fluid compositions are dominated by wall–rock interactions in a fluid-poor system), hydrogen isotopes may work reasonably well to determine whether fluids in fault zone originate from fluid-producing metamorphic reactions in the crust or at the surface. Nevertheless, the effects of hydrogen (in adsorbed water in defects as well as in free aqueous fluid) on mineral–fluid fractionation of hydrogen isotopes are not well understood. The presence of other fluid species, notably methane, may have a major influence on hydrogen isotopic composition. Cl-stable isotopes, and Cl/Br ratios are very conservative (i.e., are not modified by mineral interactions), but they may not be very distinctive (Banks *et al.* 2000). He^3/He^4 isotope ratios give good indications of inputs of mantle volatiles but must be carefully linked to fluid flows, because He is very insoluble in saline water. Improvements to analytical techniques for conservative element and isotope ratios are needed to assess the origin of fluids in orogenic belts.

With the development of hot dry rock projects, the seismicity observed during fluid injection demonstrates the potential for using seismicity to track fluid migration. The Camborne School of Mines HDR project, the German KTB deep drilling project, and the Soultz-sous-Forets HDR project are promising examples of this. Microseismicity is a marker of where fluid-induced crack propagation leads to higher permeability. Thus, more high-resolution instrumentation of seismogenic zones is needed.

New analytical techniques are needed to date mineralization events and associated mineral fabrics, as well as fault zone fluids. Where it can be established that fluid inclusions are conserved chemically and physically during exhumation of the fault zone, detailed fluid inclusion studies can reveal changes from hydrostatic to lithostatic pressures. Fault permeability and permeability anisotropy within fault zones is poorly understood. Additional measurements of these parameters are needed, both with depth and in different tectonic settings.

Integrative community models such as those developed by atmospheric scientists to study climate change (e.g., COHMAP, Hadley Center) are required to quantify coupled fluid flow and geochemical-thermomechanical processes within fault zones. Processes to be considered include changes in porosity and permeability due to variations in effective stress, thermal expansion, and mineral precipitation, fluid flow induced by fluid source/sink terms, multicomponent fluid transport (H_2O , CO_2 , NaCl), and elastic-plastic-viscous deformation. The model should also include advective-diffusive heat and solute mass transport, fluid-rock chemical interactions by pressure solution, and precipitation-dissolution reactions. In addition, we need a suite of models at the pore- or grain-scale to compliment those at the continuum scale (e.g., Discrete Elements Models). Grain-scale models may help to constrain field observations at the hand-specimen scale by determining effective parameters for continuum-scale models. Such models may help us understand the rheological and geochemical transitions which occur along fault zones connecting Earth's surface with the asthenosphere. The problem remains of testing of the numerical codes, an exercise that is too often neglected by the geoscience community.

Our group recognized the need for monitoring both modern flow systems and for reconstructing fossil flow systems in exhumed fault zones. The best field sites for modern flow systems are characterized by extensive internal brittle deformation and are covered with an extensive seismic network to obtain continuous, 3D images of the microseismicity. Further drilling of active fault zones are required to measure fluid characteristics (pressure, chemistry, isotopic) in and around the damage zone and fault core. Possible candidate field areas are the Apennines in Italy and the Southern Alps of New Zealand. These areas would compliment existing or planned sites in the U.S.A., Japan, Greece, and Taiwan where long-term observation of active faults is scheduled.

Unresolved Questions

Where do fluids in fault zones originate and at what rate are they available to fault zones?

How do different fluids from different sources mix in fault zones, if at all?

What is the nature of fluid–chemical–mechanical–thermal interactions within fault systems?

How can fluids be pumped down into dry basement rocks? The transient fluid pressure gradient is there, but what is the mechanism?

More generally, can we reconcile the petrological prediction of a dry, strong lower crust with seismic data?

How much time is required for the fluid to achieve local geochemical equilibrium with a rock?

What criteria can be used to distinguish pressure solution from precipitation–dissolution reactions?

Which fluid–rock geochemical reactions are most relevant to fault strength?

Why do aftershocks on most fault zones have the same frequency?

Why do oceanic transform faults not have many aftershocks?

These questions can only be addressed with an integrated effort among different specialists in the Earth Science community. They need the support of governmental funding agencies that are willing to make long-term investments in basic research.

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