Crustal stress, faulting and fluid flow

RICHARD H. SIBSON

Department of Geology, University of Otago, PO Box 56, Dunedin, New Zealand

Abstract: Differential stress exerts both static and dynamic effects on rock-mass permeability, modulating fluid flow in the Earth's crust. Static stress fields impose a permeability anisotropy from stress-controlled features such as faults, extension fractures, and stylolites which, depending on the tectonic regime, may enhance, or counteract existing anisotropic permeability in layered rock sequences. Textural evidence from hydrothermal veins suggests, however, that fluid flow in fault-related fracture systems generally occurs episodically and that dynamic stress cycling effects are widespread. In the vicinity of active faults that undergo intermittent rupturing, permeability and fluid flux may be tied to the earthquake cycle through a range of mechanisms, leading to complex interactions between stress cycling, the creation and destruction of permeability, and fluid flow. Mechanisms for fluid redistribution include: (1) various forms of *dilatancy* (localized to the fault zone or extending through the surrounding rock mass) related to changes in shear stress and/or mean stress that occur during the *fault loading* cycle; (2) localized *post-seismic redistribution around rupture irregularities*, especially dilational jogs and bends which act as *suction pumps*; and (3) post-seismic discharge of fluids from overpressured portions of the crust through *fault-valve* action when ruptures breach impermeable barriers. All of these processes may be involved in fluid redistribution around active faults, but they operate to varying extents at different crustal levels, and in different tectonic regimes.

This paper explores the dynamic interplay between faulting and fluid flow in the Earth's crust. In recent years there has been a growing appreciation of the extent and varying character of fluid flow at all crustal levels (Oliver 1986; Cathles 1990; Fyfe 1990; Torgerson 1990, 1991). Sources for crustal fluids include meteoric water derived from the atmosphere through infiltration of the crust, connate and formation waters and hydrocarbon fluids in sedimentary basins, water derived from dehydration reactions during prograde metamorphism, and magmatic fluids derived from deep crustal levels or the mantle (Fyfe *et al.* 1978; Irwin & Barnes 1980; Gold & Soter 1984; Etheridge *et al.* 1984). Major driving forces for fluid migration include gradients in hydraulic potential that arise from topography, localized heat sources, sedimentary compaction, metamorphic dewatering, and mantle degassing. Because many of the driving potentials are long-lasting, there has been a tendency to analyse flow systems for constant permeability models (e.g. Norton 1982; Bethke 1986).

Hydrothermal vein systems, however, provide abundant evidence for massive focused fluid flow along faults and fractures and their textures frequently record a history of incremental deposition, suggesting that the flow was intermittent (e.g. Hulin 1925; Newhouse 1942; Ramsay 1980; Sibson 1981). Gold–quartz veins, especially, document the passage of substantial volumes of aqueous fluid through fault-fracture networks (e.g. Kerrich 1986; Cox *et al.* 1991; Sibson 1992*a*). As an example, consider the Mother Lode vein system of early Cretaceous age developed within the Melones fault zone in the western Sierra Nevada foothills of California (Knopf 1929). Individual quartz veins are hosted on reverse faults within the fault zone. In places, continuous veins averaging over 1 m in thickness can be traced for kilometres along strike and have been mined to depths in excess of 1 km. Per kilometre of fault strike-length, the volume of fault-hosted quartz is therefore *c.*10^6 m^3. Established reverse displacements on the hosting faults range up to 100 m or so, with ribbon vein textures suggesting hundreds of episodes of hydrothermal deposition. Even assuming a 100% efficient precipitation mechanism, well in excess of 10^9 m^3 of aqueous fluid (the fluid volume of a supergiant oil-field!) would have to be flushed through the fault per kilometre strike-length to deposit this volume of quartz, given its low solubility (Fyfe *et al.* 1978). And the vein system extends along strike for at least 200 km! A likely inference, therefore, is that fluctuations in stress and fracture permeability associated with episodic fault slip act to modulate primary flow systems. In particular structural settings, these fluctuations in stress and permeability appear to trigger episodes of

From PARNELL, J. (ed.), 1994, *Geofluids: Origin, Migration and Evolution of Fluids in Sedimentary Basins*, Geological Society Special Publication No. 78, 69–84.

hydrothermal precipitation and play a crucial role in the formation of mineral deposits.

No attempt is made to review the mathematical theory of time-dependent fluid redistribution in poroelastic media which is a subject of considerable complexity (e.g. Rice & Cleary 1976; Rudnicki & Hsu 1988). The approach adopted is semi-quantitative at best, and seeks only to provide simple physical models that account for the geological evidence of large-scale fluid redistribution around faults in certain tectonic environments. Because of the uncertainties in many of the critical parameters and constitutive relationships, and the heterogeneous character of material properties (strength, permeability, etc.) in the Earth's crust, the analysis is couched in terms of classical rather than modern fracture mechanics. While this approach is less than satisfactory in quantifying the different mechanisms for fluid redistribution, it does identify areas where future detailed modelling may prove worthwhile.

Seismogenic crust

Studies of tectonically active terrains have shown that fault displacements in the upper continental crust are largely accomplished by increments of seismic slip along pre-existing structures. The depth extent of seismic activity, representing the zone of unstable frictional sliding, is temperature-dependent. For moderate or higher geothermal gradients away from areas of subduction, this *seismogenic regime* is restricted to the upper one-third to one-half of deforming continental crust (Sibson 1983). In the context of plate tectonics, therefore, the Earth's crust may be divided into intraplate regions, where the rate of seismic activity is low and the stress field is effectively *static* over long time periods, and areas of active deformation associated with plate boundaries where stress fields in the vicinity of active faults are coupled to the earthquake stress cycle and are *time-dependent*. Thus, whereas fluid flow in intracratonic regions is unlikely to be affected by comparatively short-term stress and permeability cycling, flow in tectonically active areas may undergo episodic perturbation on time-scales of the order of the repeat times of large earthquakes (10–10^4 years). Subsidiary perturbations may be induced by second order seismicity.

Fluid pressure and fault stability

Hydrostatic fluid pressures prevail where interconnected pore space and fracture systems are freely linked through to the Earth's surface. In such circumstances, the pore-fluid factor, $\lambda_v = P_f/\sigma_v = P_f/(\rho g z) \approx 0.4$, where P_f and σ_v are, respectively, the fluid pressure and vertical stress (overburden pressure) at a depth, z, in the crust, ρ is the average rock density, and g is the gravitational acceleration. However, there is a great deal of evidence that at depths greater than a few kilometres within deforming crust, fluid pressures commonly rise above hydrostatic towards lithostatic values ($\lambda_v \rightarrow 1$) (see below, Figs 7 & 9). The condition $\lambda_v \approx 1$ is also widely assumed to prevail in areas undergoing prograde regional metamorphism (Fyfe *et al.* 1978; Etheridge *et al.* 1984).

Hubbert & Rubey (1959) first called attention to the role of overpressured (suprahydrostatic) fluids in reducing the frictional strength of faults, which can be represented by a criterion of Coulomb form:

$$\tau_f = C + \mu_s \sigma_n' = C + \mu_s(\sigma_n - P_f) \qquad (1)$$

where C is the cohesive or cementation strength of the fault (small for an active structure), μ_s is the static coefficient of rock friction, and σ_n is the normal stress on the fault. From laboratory experiments, $\mu_s = 0.75 \pm 0.15$ for a broad range of rock types (Byerlee 1978), and field observations suggest that natural faulting involves comparable friction coefficients (Sibson 1990*a*). Fault reactivation thus occurs when the shear stress along the fault, τ, equals τ_f, and may be brought about by rising shear stress, decreasing normal stress, or by an increase in fluid pressure.

This simple criterion has been shown to be applicable to several cases of induced seismic activity occurring in the top few kilometres of the crust (Nicholson & Wesson 1990) and may well remain valid throughout the seismogenic zone, though fault failure at depth may also be affected by time-dependent processes such as stress corrosion (Das & Scholz 1981). Note, however, that there is now good evidence for seismic rupturing within overpressured portions of the crust such as the Western Taiwan fold/thrust belt and regions adjacent to the San Andreas fault in California (Sibson 1990*a*). Complex coupling between episodes of fault failure, the creation and destruction of fracture permeability, and fluid redistribution is especially likely in such overpressured regions (see later discussion on fault-valve activity).

Coupled time-dependent permeability

While inactive faults often act as impermeable seals through the presence of clay-rich gouge and hydrothermal cementation (Smith 1980;

Hooper 1991), the intrinsic roughness of natural fault surfaces (Power *et al.* 1987) has the implication that freshly ruptured faults should become highly permeable, if tortuous, channelways post-failure. Rupture zone permeability is, however, likely to be short-lived. Evidence from geothermal fields suggests that hydrothermal flow along fractures rapidly leads to hydrothermal precipitation and self-sealing (Grindley & Browne 1976; Batzle & Simmons 1977). Experiments show that flow of hot water ($T > 200°$ C) along pressure gradients in granite leads to dramatic reduction in permeability through hydrothermal dissolution and reprecipitation (Morrow *et al.* 1981). In addition, laboratory experiments suggest that diffusional crack healing in 'wet' quartz is fast at $T > 200°$ C (Smith & Evans 1984; Brantley 1992). Solution-precipitation creep is also particularly effective in clogging porosity and lowering permeability in fine-grained quartz-bearing rocks over the temperature range 200–400° C (McClay 1977; Sprunt & Nur 1977; Cox & Etheridge 1989). Observational and experimental periods over which these various processes are effective in reducing permeability range from hours to months.

Thus in the vicinity of active faults, there must be continual competition between the creation and destruction of fracture permeability. While permeability destruction through hydrothermal activity is known to be fast-acting in high-level geothermal systems, it also seems likely to be particularly effective in the bottom half of the continental seismogenic regime at, say, 7–15 km depth. These processes of porosity and permeability destruction (Walder & Nur 1984; Nur & Walder 1990) counter the arguments of Brace (1980, 1984), based on laboratory and *in situ* measurements of rock permeability, that crystalline rocks are generally too permeable to allow the development of fluid overpressures. They play an important role in some of the recent models that attempt to account for the apparent weakness of major fault zones in terms of fluid overpressuring (Byerlee 1990, 1993; Sibson 1990*a*; Rice 1992; Sleep & Blanpied 1992).

Static stress field effects

Stress-controlled features affecting rock permeability include brittle faults, microcracks, extension fractures, and stylolitic solution seams. Their relationship to a stress field with principal compressive stresses $\sigma_1 > \sigma_2 > \sigma_3$ in homogeneous, isotropic rock is as shown in Fig. 1, which illustrates the range of fault/fracture/stylolite mesh systems that may develop in different circumstances. Faults tend to initiate as Coulomb shears containing the σ_2 axis and lying at ± 20–30° to σ_1 (Anderson 1951). In 'classical' fracture mechanics, macroscopic extension fractures (mode I cracks in the parlance of modern fracture mechanics) form perpendicular to σ_3 by natural hydraulic fracturing in accordance with the criterion:

$$P_f = \sigma_3 + T \qquad (2)$$

provided $(\sigma_1 - \sigma_3) < 4T$, where T is the long-term tensile strength of the rock (Secor 1965). Microcracks developing by grain impingement also have a preferred orientation perpendicular to σ_3. Impermeable stylolites develop as anticracks in surfaces subperpendicular to σ_1 (Fletcher & Pollard 1981). The different components may combine in various fault/fracture/stylolite meshes as shown, but the full range of components need not be present in all cases. For instance, the development of stylolites by pressure solution processes depends critically on factors such as grain size and composition, being far more widespread in fine-grained sedimentary rocks, particularly limestones (Groshong 1988).

Faults and fractures have the potential to impart directional permeability to the rock mass (e.g. du Rouchet 1981), but may also become choked with hydrothermal deposits and/or clay-rich gouge and alteration products (e.g. Roberts 1991). Moreover, the permeability characteristics of a fault depend to some extent on the character of the host rock. Faults developing in initially high-porosity sedimentary rock may, through comminution and porosity collapse, form deformation bands that are relatively impermeable in comparison with the wallrock (Aydin & Johnson 1978). In contrast, faults in initially low-porosity rocks may enhance permeability locally through cataclastic brecciation.

Extension fracturing on both microscopic and macroscopic scales can greatly enhance permeability in the σ_1/σ_2 plane, the effect becoming pronounced as $P_f \rightarrow \sigma_3$. Comparatively impermeable stylolitic seams are associated with tabular zones of reduced porosity (Groshong 1988; Carrio-Schaffhauser *et al.* 1990) and, overall, will generally tend to restrict flow perpendicular to the σ_2/σ_3 plane.

Existing anisotropic permeability in layered rock sequences may therefore be either enhanced or reduced, depending on the type of stress field prevailing and the dominant attitude of the layering (Fig. 1). In the case of predominantly horizontal layering, and for the more

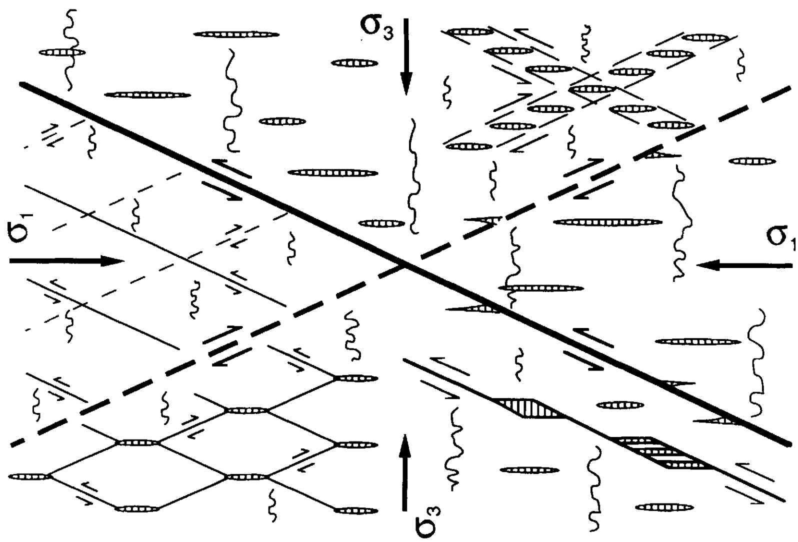

Fig. 1. Faults (Coulomb shears), extension fractures (hachured), and stylolites (squiggly lines) forming in a triaxial stress field, illustrating possible combinations of these stress-controlled features which might affect permeability of the rock mass. A Hill fault/fracture mesh is shown in the lower left. Diagram represents a thrust fault stress regime when upright, a normal fault regime when viewed sideways, and a strike-slip regime in plan view.

common situation where fracture-enhanced permeability predominates over impermeable stylolitic bands, bedding-parallel permeability will be enhanced in a compressional tectonic regime, while trans-bedding permeability will tend to be enhanced in extensional and strike-slip regimes (cf. du Rouchet 1981). In viewing the diagrams, the enhancement of out-of-plane permeability in the σ_2 direction should also be noted; this is likely to be of special significance in strike-slip regimes. Where the dominant anisotropy is sub-vertical, as in belts of upright folds with well-developed slaty cleavage, trans-anisotropy permeability will be enhanced by low-angle faults and extension fractures developed through horizontal compression. The situation may become vastly more complicated in regions of complex structure where rotational strains have developed in heterogeneous stress fields.

Fluid expulsion through Hill fault/fracture meshes

Hill (1977) showed how the seismological characteristics of earthquake swarms could be accounted for by the passage of hydrothermal or magmatic fluids through a 'honeycomb mesh' of interlinked shear and extension fractures (Figs 2 & 3). Such meshes may develop in extensional and strike-slip settings because individual hydraulic extension fractures can only extend over limited depth intervals in the Earth's crust (Secor & Pollard 1975). The ability of such fault/fracture meshes to transmit fluids is highly sensitive to the value of the effective least compressive stress, $\sigma_3' = (\sigma_3 - P_f)$, but becomes large when $\sigma_3' \rightarrow 0$. This can be achieved close to the Earth's surface in areas of extensional and strike-slip tectonics, but the meshes may also develop at depth in the crust when fluid pressure approaches the lithostatic load ($\lambda_v \rightarrow 1$).

As a consequence of the contrasting mechanical properties of alternating layers, Hill fault/fracture-meshes are particularly likely to develop in horizontally-layered rock sequences undergoing extension with σ_3 horizontal and σ_1 vertical (Fig. 3). Under conditions of increasing fluid pressure, the brittle failure mode of intact rock is governed by the level of differential stress, $(\sigma_1 - \sigma_3)$, in relation to its tensile strength, T (Secor 1965). Provided $(\sigma_1 - \sigma_3) < 4T$, hydraulic extension fractures form perpendicular to σ_3 in accordance with equation (2), whereas if $(\sigma_1 - \sigma_3) > 4T$, the rock fails in shear in accordance with the Coulomb criterion (eqn 1). Thus, in a typical sequence of alternating sandstones and shales with $T_{sst} > T_{sh}$, the shales may fail by development of brittle Coulomb shears (perhaps in conjugate sets), while the sandstone fails by hydraulic extension fracturing to form a fault/fracture mesh. Natural fracture meshes of this kind may develop over a range of scales, but generally tend to be less ordered than

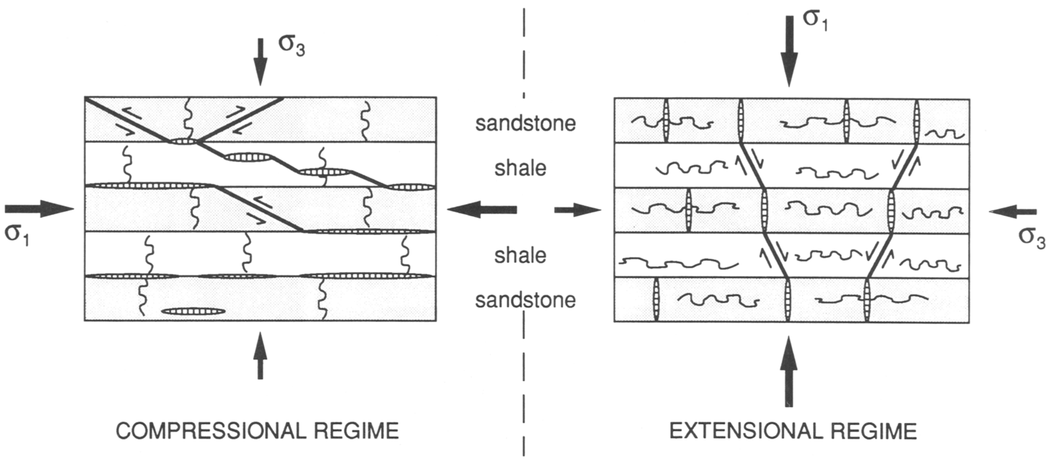

Fig. 2. Schematic representation of stress-control of permeability by 'brittle' structures (faults, extension fractures (hachured), and stylolites (wavy lines)) affecting an anisotropic, sandstone-shale sequence in compressional and extensional stress regimes.

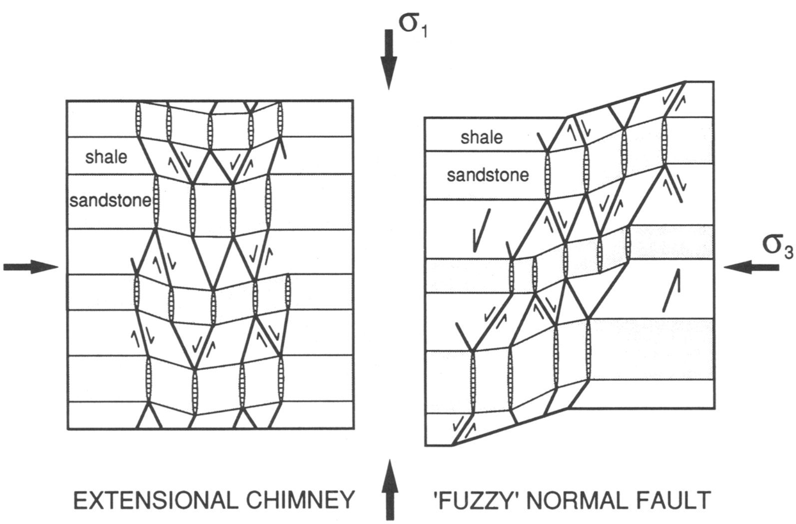

Fig. 3. Hill fault/fracture meshes acting as fluid conduits within an extensional stress regime, either as extensional chimneys or as 'fuzzy' normal faults. Note that the passage of large fluid volumes leads to increased disorganisation and brecciation within the fault/fracture mesh.

those portrayed diagrammatically in Fig. 3. The passage of large fluid volumes through fault/fracture meshes leads to brecciation with rock fragments initially bounded by combinations of shear and extensional fractures. Continued flow causes rotation and progressive comminution of the fragments in a process akin to fluidisation. Such processes may have contributed to the extensive tabular bodies of high-dilation breccia within the Monterey Shales of California (Redwine 1981) and similar formations elsewhere.

Fault/fracture-meshes of this kind, developed in well-layered sedimentary sequences, tend to form either as subvertical extensional 'chimneys', or as 'fuzzy' normal faults with an overall component of shear across the mesh (Fig. 3). In the Monterey Shales they have apparently acted as major fluid expulsion structures, transporting large volumes of aqueous and hydrocarbon fluids across the bedding anisotropy, the relict mesh structure being preserved through cementation by silica, carbonate, and bitumen. Such structures likely developed by the breaching of seals to overpressured fluid compartments.

Stress cycling effects

Earthquake ruptures spread over portions of existing fault surfaces at *c*.3 km s^{-1} (corresponding to the shear wave velocity in the upper crust), so that in the case of moderate (*c*.M5) to large (*c*.M7) ruptures with dimensions of kilometres to tens of kilometres, the period of coseismic rupturing typically lasts for seconds to tens of seconds (Sibson 1989). Accompanying effects relevant to fluid redistribution include: (i) the creation of transient fracture permeability along the primary rupture zone; (ii) the development of subsidiary fracturing at specific structural sites such as fault jogs; (iii) shear stress reduction over the rupture plane and intensification at the rupture tips, with localized enhancement or reduction of mean stress around fault irregularities; and (iv) the abrupt reduction of fluid pressure at dilational sites within the rupture zone.

Several time-scales must therefore be considered in relation to the earthquake stress cycle; long-term accumulation of shear stress through the inter-seismic period lasting tens to perhaps many thousands of years, possible *pre-seismic* stress changes associated with precursory slip, the rapid *co-seismic* drop in shear stress during rupturing which at any one place occurs over a period of a few seconds, and a period of *post-seismic adjustment* (corresponding to the aftershock phase) which may last for days to years depending on the size of the rupture and the physical characteristics (strength, heterogeneity, permeability, etc.) of the rock mass (Fig. 4). Elasticity theory suggests that significant stress cycling around a seismically active fault should be restricted to a *response zone* whose extent compares broadly with characteristic rupture dimensions, extending laterally for perhaps 10–15 km from ruptures that occupy the full depth of the continental seismogenic zone (Sibson 1989). The highest amplitude cycling occurs close to the causative fault but is likely to be locally exaggerated in the vicinity of jogs and other fault irregularities. However, it should be noted that in recent years there has been increasing recognition of triggering of remote seismicity and other effects, likely to be fluid-related, that occur far outside the expected response zone derived from linear elasticity theory (e.g. Hill *et al.* 1993).

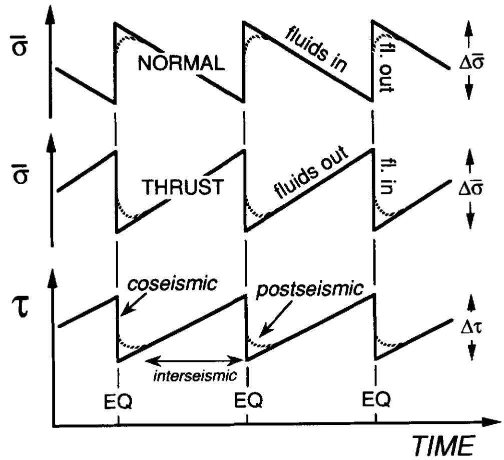

Fig. 4. Variations in mean stress ($\bar{\sigma}$) during the cyclic accumulation and release of shear stress (τ) on thrust and normal faults that are optimally oriented for frictional reactivation with $\mu_s = 0.75$. Different phases of the earthquake stress cycle and fluid movement in and out of the response zone are indicated. Dotted lines indicate possible time-dependence of stresses post-failure.

Dilatancy related to the fault loading cycle

The state of dilatational strain in a rock mass may be affected by variations in the levels of both shear stress and mean stress. In the search for precursory effects to fault failure, attention was focused initially on possible dilatant effects associated with the earthquake cycle of shear stress accumulation and release, the belief being that dilatant strains should develop in the rock mass around a fault as shear stress increases during fault loading (Fig. 4). Fluids would be drawn into the dilatant rock volume pre-failure, only to be expelled post-failure once shear stress was relieved. This dilatancy–diffusion hypothesis for earthquake prediction (Nur 1972; Scholz *et al.* 1973) was predicated on the development of extensive high-stress ($\tau > 100$ MPa) microcrack dilatancy in the surrounding rock mass. It formed the basis of the original *seismic pumping* concept developed by Sibson *et al.* (1975) to account for fluid redistribution in the vicinity of seismically active faults.

Twenty years on, there is little evidence that this particular form of dilatancy or the requisite stress levels are widespread in deforming seismogenic crust (Hickman 1991), so that the original seismic pumping mechanism must be considered invalid. However, as Nur (1975) has pointed out, other forms of dilatancy with different stress-dependences may also operate in the vicinity of fault zones, and need to be

Table 1: *Postulated dilatancy mechanisms*

Mechanism	Primary dependence	Distribution of dilatancy	Depth within seismic zone	References
High-stress microcrack dilatancy at high τ and high $\bar{\sigma}' = (\bar{\sigma} - P_f)$	$\Delta\tau$	Laterally extensive	Deep	Nur 1972; Scholz *et al.* 1973
Low-stress microcrack dilatancy from subcritical crack growth at low τ	$\Delta\tau$	Laterally extensive	Shallow–deep	Crampin *et al.* 1984
Low-stress microcrack dilatancy at low τ, high P_f, and low $\bar{\sigma}'$	$\Delta\tau$	Laterally extensive?	Deep	Fischer & Paterson 1989
'Sand-pile' dilatancy under low τ and low $\bar{\sigma}'$	$\Delta\tau$	Restricted to fault zone	Shallow	Nur 1975; Marone *et al.* 1991
Existing joint/fracture dilatancy	$\Delta\tau$ and/or $\Delta\bar{\sigma}$?	Laterally extensive?	Shallow	Nur 1975
$\Delta\bar{\sigma}$ compactive effects in highly fractured or porous material at low $\bar{\sigma}'$	$\Delta\bar{\sigma}$	Laterally extensive	Shallow?	Sibson 1991
Hydrofracture dilatancy under low τ and high P_f	ΔP_f	Laterally extensive?	Shallow-deep	Sibson 1981
Grain-scale particulate flow under low $\bar{\sigma}'$ and high P_f	ΔP_f	Restricted to fault zone	Deep	Borraidale 1981; Cox & Etheridge 1989

evaluated as possible dilatancy pumping mechanisms. The different dilatancy mechanisms may be grouped into those sensitive primarily to varying shear stress, those sensitive to variations in mean stress, and those driven by high fluid pressure levels (Table 1), though some degree of overlap occurs. More than one of the mechanisms may contribute to fluid redistribution in any particular circumstance. Shear stress sensitive dilatancy is likely to predominate in low porosity rocks, whereas high porosity rocks may develop significant dilatant strains from variations in mean stress.

It has to be emphasized, however, that *no* firm constitutive relationships applicable to the real Earth have yet been established for *any* of the postulated mechanisms. Another key issue, affecting the volume of fluid involved in dilatancy pumping, is the extent of the region experiencing cyclic dilatancy. Are significant dilatant strains restricted to the fault zone itself, to its immediate surrounds, or do they extend over broad areas of the surrounding rock mass?

Dilatancy dependent on shear stress variations. Nur (1975) recognized three varieties of dilatancy dependent on shear stress; *microcrack dilatancy* which is sensitive to small variations in shear stress at high ambient levels of shear stress, *sand-pile dilatancy* which is stress-sensitive at low ambient levels of shear stress, and *joint dilatancy* in fractured rock masses which maintains an approximately uniform stress sensitivity. As previously discussed, it seems unlikely that high-stress microcrack dilatancy is widespread in seismogenic crust. Crampin *et al.* (1984) have suggested, however, that microcrack dilatancy may also develop by slow sub-critical crack growth under comparatively low shear stress levels and give rise to extensive-dilatancy anisotropy (EDA) over broad areas of deforming crust. In addition, Fischer & Paterson (1989) have demonstrated that significant microcrack dilatancy may also develop under low stress levels at high fluid pressures and low effective confining pressures.

Sand-pile dilatancy is likely to develop in any granular aggregate material under low levels of shear stress and must undoubtedly operate in the granular cataclastic material contained within brittle fault zones at high crustal levels (e.g. Marone *et al.* 1990). However, the volume of fluid involved is then likely to be comparatively minor. Extensively fractured rock masses adjacent to fault zones at high crustal levels are also likely to experience joint dilatancy as shear stress varies.

Dilatancy dependent on mean stress variations Tectonic shear stress on a fault cannot generally be increased without also changing the normal stress on the fault (altering its frictional strength) and the level of mean stress ($\bar{\sigma} = [\sigma_1 + \sigma_2 + \sigma_3]/3$) (Sibson 1991). This in turn affects the state of dilatational strain, especially within high porosity rock masses. Patterns of fluid flow during fault loading must therefore be influenced by poroelastic effects arising from changes in $\bar{\sigma}$ as the shear stress on a fault increases. This is illustrated in Fig. 4 for the end-member cases of thrust and normal faults that are optimally oriented for frictional reactivation; the coupled changes in mean stress ($\Delta\bar{\sigma}$) are somewhat greater than the shear stress drop ($\Delta\tau$). For the two cases illustrated, a typical earthquake shear stress drop, $\Delta\tau = 1$ MPa, causes a post-failure increase in mean stress, $\Delta\bar{\sigma} = 1.25$ MPa in the vicinity of a normal fault, or a post-failure decrease, $\Delta\bar{\sigma} = 1.25$ MPa, in the case of a thrust fault. This corresponds to a change of $\pm$ 125 m in equivalent hydrostatic head.

Full analysis of these poroelastic effects in crust with a heterogeneous distribution of stress and permeability is likely to be complicated (e.g. Rice & Cleary 1976) but, neglecting for the time-being any shear stress related dilatancy and rupture tip effects, it is clear that as shear stress rises during the loading of a thrust fault, mean stress also increases so that fluids should be driven away from the fault and out of the response zone by poroelastic compaction, only to be drawn back in post-failure. In the case of normal faults, the to-and-fro motion should be reversed, with fluids drawn in towards the fault during loading to be expelled post-failure. For strike-slip faulting, the coupling of mean stress to shear stress can lie anywhere between these end-member cases (Sibson 1993). More complicated patterns of fluid redistribution from mean stress changes are to be expected at rupture tips (Muir Wood & King 1993).

Fluid redistribution from varying mean stress is likely to be important when the effective mean stress, $\bar{\sigma} = (\bar{\sigma} - P_f) \rightarrow 0$, and the ratio ($\Delta\bar{\sigma}/\bar{\sigma}'$) is large. These conditions will be satisfied in the near-surface, where intensely fractured portions of the upper crust respond as a fluid-saturated, blocky aggregate. In such settings, cyclic variations in mean stress should induce to-and fro-motion of fluids, with fluid pressures staying close to hydrostatic. $\Delta\bar{\sigma}$ effects may, however, also be strong at depth in the crust when $\bar{\sigma} = (\bar{\sigma} - P_f) \rightarrow 0$ and the rock mass is deforming by particulate flow (see below).

The 1983 M7.3 earthquake at Borah Peak, Idaho, which involved rupturing of a range-bounding normal fault, provided a good example of post-seismic discharge coupled to an increase in mean stress postfailure. The earthquake was followed by a large fluid discharge (c.0.3 km^3) over a period of several months, including immediate post-seismic fountaining from a series of fissures striking parallel to the rupture trace (Wood *et al.* 1985). The discharge seems largely attributable to the increase in horizontal stress and mean stress in the upper few kilometres of the crust consequent on shear stress relief along the normal fault (Sibson 1991; Muir Wood & King 1993).

Dilatancy driven by fluid pressure. Arrays of extension fractures develop in the crust by hydraulic fracturing when the requisite conditions ($P_f = \sigma_3 + T$, and $(\sigma_1 - \sigma_3) < 4T$) are met (Secor 1965). Because the tensile strength of rocks is generally low ($T < 10$ MPa), such *hydrofracture dilatancy* is a low differential stress phenomenon. The necessary conditions may be satisfied by hydrostatic levels of fluid pressure within perhaps a few hundred metres of the earth's surface in extensional and strike-slip regimes, but fluid overpressures with $\lambda_v \rightarrow 1$ are needed at greater depths (Sibson 1981). Provided the rock mass retains a finite tensile strength, fluid pressures in excess of the lithostatic load ($\lambda_v \geq 1$) are required at all depths for hydraulic extension fractures to develop in compressional stress regimes where $\sigma_v = \sigma_3$. Localization of hydrofracture arrays to the immediate vicinity of fault zones suggests that, in many cases, the fault zones are themselves the principal conduits for the migration of overpressured fluids (Cox *et al.* 1991). The development of hydrofracture dilatancy in compressional regimes that are strongly fluid overpressured is an accompaniment to extreme fault-valve action (see below).

Grain-scale microcrack dilatancy and associated particulate flow (approximately equivalent to sand-pile dilatancy) may also develop in fluid overpressured rock masses where $\lambda_v \rightarrow 1$, and the effective mean stress, $\bar{\sigma}' = (\bar{\sigma} - P_f) \rightarrow 0$ (Borraidale 1981; Cox & Etheridge 1989; Fischer & Paterson 1989; Knipe 1989).

Mixed dilatancy effects. In most natural settings it is likely that more than one of these various dilatancy mechanisms may be operating at a given time. For example, consider the coupled variation of both shear and mean stress around thrust and normal faults as illustrated in Fig. 4. In the case of the thrust fault, any dilatancy related to increasing shear stress during fault loading is opposed by the coupled rise in mean stress, whilst any post-failure tendency for crack closure from reduced shear stress is counteracted by the lowered mean stress. In the case of normal faults, however, development of dilatancy during loading is favoured both by the increasing shear stress and by the coupled reduction in mean stress. Post-failure, reduced shear stress and increased mean stress both

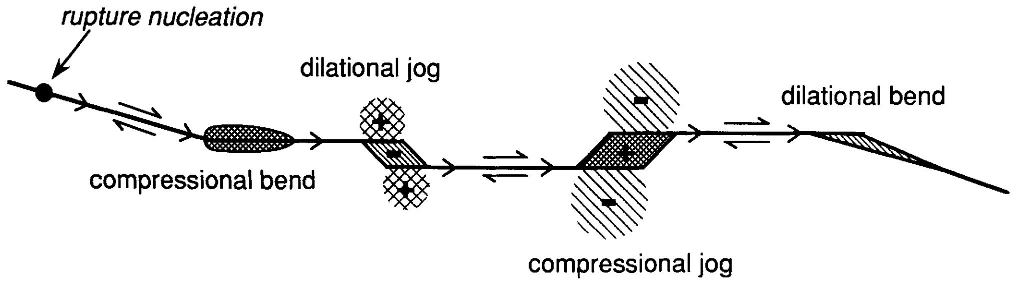

Fig. 5. Seismotectonic carbon of an irregular rupture trace (not to scale), showing areas of enhanced (+ and cross-hatched) and reduced (− and diagonal hachures) mean stress arising from a rupture propagating from left to right. Note that the response of isolated fault bends depends on the direction of rupture propagation. Diagram represents a map view of a strike-slip fault, or a cross-section through a dip-slip fault.

contribute to crack closure. On these arguments, dilatancy effects throughout the fault loading cycle should be more pronounced in extensional rather than compressional stress regimes. However, another factor to be considered is that at the same depth and fluid pressure level (λ_v value), the shear stress required for frictional reactivation of a thrust fault is about four times that needed to reactivate a normal fault (Sibson 1974). Effects from shear stress related dilatancy are therefore likely to be more pronounced in the vicinity of thrust faults.

Until the details of the stress levels driving faulting and the constitutive laws for the different dilatancy mechanisms are fully resolved, it is impossible to evaluate the contributions of the various mechanisms to fluid redistribution in the crust. Of critical concern, because it affects the volume of fluid redistribution by dilatancy pumping, is the question of whether cyclic dilatational strains are generally localized to the material within fault zones, or whether they extend over broad regions of the surrounding crust. While there is considerable geological evidence for localized dilatancy within shear zones at all crustal levels (Hobbs *et al.* 1990), documentation of microstructures demonstrating cyclic dilatancy on a regional scale is generally lacking. Thus, the extent to which dilatancy pumping of one form or another may redistribute fluids, and in particular may draw fluids down from the near-surface to levels of the crust undergoing prograde metamorphism, as proposed by McCaig (1988), remains unresolved.

Post-seismic fluid redistribution

Rupturing of an irregular fault surface leads to abrupt post-failure changes in mean stress localised around the structural irregularities (Segall & Pollard 1980; Sibson 1989) (Fig. 5). As a consequence, there is a tendency for fluids to be redistributed from areas of raised mean stress to areas of lowered mean stress (Nur & Booker 1972). Fluids are driven out of compressional jogs and bends where mean stress increases post-failure, while the most intense fluid influx, coupled with aftershock activity, is concentrated in regions of sharply reduced mean stress such as dilational jogs and bends. At these dilational sites, rapid slip transfer during rupture propagation causes abrupt local reductions in fluid pressure below ambient (hydrostatic?) values. In some instances, the induced suctional forces may lead to rupture arrest (Sibson 1985, 1989). Dilational fault jogs and bends thus act essentially as *suction pumps* and are often characterized by multiply recemented wallrock breccias resulting from repeated hydraulic implosion. Larger dilational structures typically comprise a mesh of extension veins, implosion breccias, and subsidiary shears in the form of a Hill fault/fracture mesh.

Active geothermal systems are commonly localized within dilational sites in extensional and transtensional fault systems. Where transected by faults in the top kilometre of the seismogenic zone (the epithermal environment), pressure reductions from slip transfer may induce episodes of boiling within the hydrothermal system, triggering mineral deposition throughout the phase of aftershock activity (Sibson 1987). The association of epithermal mineralization with the high levels of extensional/transtensional fault systems thus arises from the coincidence of this 'boiling horizon' with the near-surface region where extensive hydrofracturing may occur under hydrostatic fluid pressure conditions (Sibson 1992*a*).

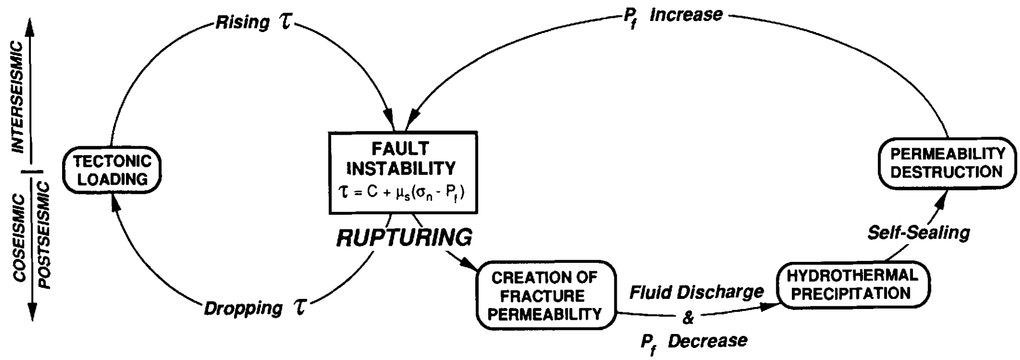

Fig. 6. Schematic representation of *fault-valve* activity, illustrating the dependence of frictional fault failure on the cycling of both tectonic shear stress and fluid pressure (after Sibson 1992*b*).

Fault-valve action in overpressured crust

From the time of Hubbert & Rubey's (1959) seminal paper on the mechanics of low-angle thrusting, it has become increasingly clear that overpressured fluids play a critical role in faulting, and may be especially important in the case of faults that are unfavourably oriented for frictional reactivation (Cello & Nur 1988; Sibson 1990*a*). Recent oil-field studies have demonstrated the widespread occurrence of fluid pressure compartments in sedimentary basins, and analysis of oil migration paths suggests that overpressured compartments become breached from time to time with episodic discharge of overpressured fluids (Hunt 1990; Powley 1990). There is also now good evidence that, in at least some areas, seismic rupturing is occurring within overpressured portions of the crust.

Fault-valve action thus occurs wherever ruptures transect suprahydrostatic gradients in fluid pressure and breach impermeable barriers, leading to upwards fluid discharge along the transient permeability of the fault zone and local reversion towards a hydrostatic fluid pressure gradient before hydrothermal self-sealing of the rupture zone occurs, and fluid overpressures rebuild at depth. Under such circumstances, the timing of successive failure episodes is controlled by the cycling of both tectonic shear stress and fluid pressure through the interseismic period (Fig. 6). A key issue, affecting the volume of fluid available for post-failure discharge and the time for fluid replenishment of the fault zone in the interseismic period, is the extent to which fault zones are locally overpressured in relation to the surrounding crust (Fig. 7).

On mechanical grounds, a case can be made that valve action is most likely to be prevalent in compressional or transpressional fault systems. Steep reverse faults are particularly likely to be effective as fault-valves because they are severely misoriented for reactivation, and frictional shear failure can only initiate when slightly supralithostatic fluid pressures are attained (Sibson *et al.* 1988; Sibson 1990*b*). Under these conditions, arrays of subhorizontal extension fractures develop by hydraulic fracturing adjacent to the fault (hydrofracture dilatancy) prior to failure, to form a lithostatically pressured reservoir that, together with associated grain-scale dilatancy, may contain a substantial fluid volume (Fig. 8). Fault failure and rupturing through the upper crust then allows fluid from the overpressured reservoir to drain upwards, with fluid pressures dropping rapidly towards hydrostatic values. The discharge may be accompanied by phase separation (if CO_2 is present), rapid mineral precipitation and hydrothermal self-sealing of the fault. Fluid pressure then rebuilds towards the critical value needed to trigger the next episode of fault slip, and the cycle repeats. Such faults may give rise to *extreme fault-valve action* involving the discharge of large fluid volumes, with fluid pressure potentially cycling between pre-failure lithostatic and post-failure hydrostatic levels.

Extreme valve action appears responsible for many mesothermal Au-quartz lodes hosted in steep reverse or reverse-oblique faults, such as the Mother Lode vein system described previously (Sibson *et al.* 1988; Cox *et al.* 1991; Boullier & Robert 1992). A typical vein assemblage consists of flat-lying extensional veins, inferred to have developed by hydraulic fracturing prior to episodes of fault slip, in mutual cross-cutting relationships with steep fault-veins formed during episodic upwards discharge along the steeply dipping fault zone (Fig. 8). The hosting shear zones are typically of mixed discontinuous–continuous ('brittle–ductile') character and developed under greenschist

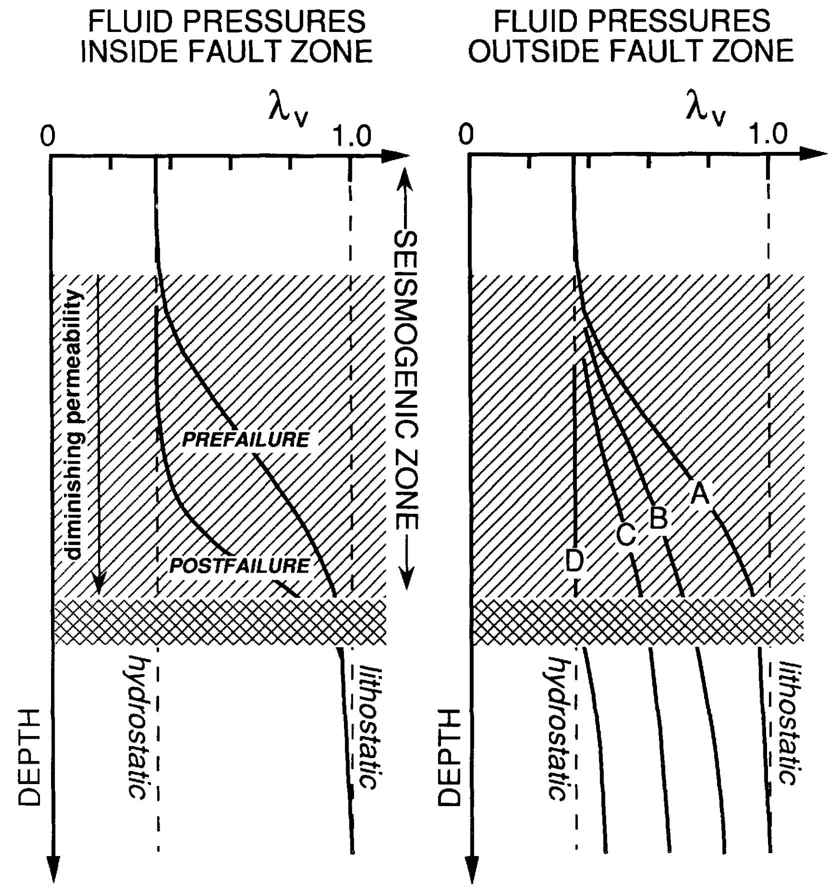

Fig. 7. Hypothetical plots of pore-fluid factor, λ_v, versus depth, illustrating pre-failure and post-failure fluid pressure gradients inside a fault zone undergoing fault-valve action in relation to possible gradients (curves A, B, C, D) in the surrounding rock mass. Curve A represents an end-member case where the pressure distributions inside and outside the fault zone are identical. Curves B, C, and D represent different degrees of localization of fluid overpressure within the fault zone, with D representing a crust that is hydrostatically pressured throughout the seismogenic zone. Conceivably, curves A–D could reflect fluid pressure gradients with increasing distance from a major fault zone.

facies metamorphic conditions at crustal levels corresponding to the base of the seismogenic zone (depths of $c.10 \pm 5$ km) with the vein material precipitated from low salinity, mixed H_2O–CO_2 fluids (Robert & Kelly 1987). Localization of the prefailure arrays of hydraulic extension fractures to the vicinity of the fault zones (generally to within tens to hundreds of metres) suggests that the fault zones are locally overpressured with respect to the surrounding crust, but the degree of comparative overpressuring is unclear (see Fig. 7). Valving action leading to the formation of mesothermal vein systems may also occur at depth within strike-slip fault systems in the vicinity of compressional jogs and bends (Fig. 5). This is in direct contrast to epithermal mineralization which typically develops in extensional/transtensional fault systems (Sibson 1992*a*)

While extreme valve action (involving the episodic discharge of large fluid volumes and the formation of major hydrothermal vein systems) may be comparatively rare and restricted to specific tectonic settings, the frequent presence of minor syn-tectonic veins in fault zones (e.g. Chester *et al.* 1993), and the growing evidence from fluid inclusion studies for fluid pressure cycling (e.g. Parry & Bruhn 1990; Mullis 1990), suggest that basic fault-valve activity is widespread in the earth's crust. Even minor valving action involving small fluid volumes is likely to have a considerable effect on the nucleation and recurrence of earthquakes (Sibson 1992*b*). The depositional environment of mesothermal vein

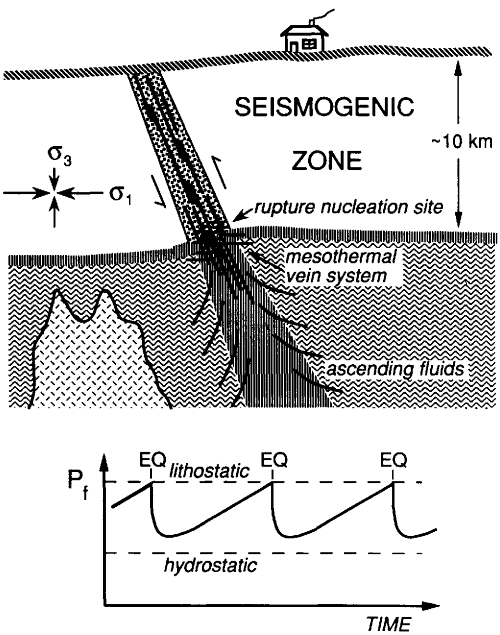

Fig. 8. Schematic diagram of extreme fault-valve action and associated fluid-pressure cycling on a high-angle reverse fault. Mesothermal gold–quartz lodes form in the region of large rupture nucleation and intense fluid pressure cycling near the base of the seismogenic zone.

systems also suggests that the region towards the base of the seismogenic zone may in general represent a time-dependent interface between hydrostatically and lithostatically pressured portions of the crust, as represented schematically in Fig. 9.

General fault zone model

The absence of a localized heat flow anomaly centred on the trace of the San Andreas strike-slip fault in California limits the time-averaged frictional shear resistance in the seismogenic zone to <20 MPa (Lachenbruch & McGarr 1990). Additional evidence for a very weak fault zone comes from the accumulation of data suggesting that the San Andreas fault is extremely unfavourably oriented for frictional reactivation within the regional stress field, lying at 65–85° to the greatest compressive stress (Mount & Suppe 1987). Present knowledge of the frictional characteristics of rock materials suggests that, despite the problem of containment, fluid overpressures are the most probable weakening mechanism that could account for this low frictional strength. As a consequence, a range of models accounting for the generation and maintenance of fluid overpressures in transcrustal fault zones have recently been proposed. A key issue in the different models is whether the high fluid pressures are derived from the fault zone acting as a migratory conduit for overpressured fluids (Sibson 1990*a*; Stark & Stark 1991; Rice 1992), or whether the fluid overpressures are continually regenerated from essentially the same fluid volume during cyclical loading (Byerlee 1990, 1993; Sleep & Blanpied 1992). The development of extensive hydrothermal veining in fault zones, especially the gold–quartz mineralization precipitated at structural levels corresponding to the lower half of the seismogenic zone, provides geological evidence that fault zones, in at least some circumstances, are acting as migratory conduits for the passage of rather substantial fluid volumes (Kerrich 1986).

If fluid overpressures and associated fault-valve action leading to fluid pressure cycling are as widespread as evidence is beginning to suggest, then fluid pressure gradients within transcrustal fault zones must be regarded as time-dependent, affecting the shear resistance profiles derived from rheological modelling. Figure 9 is an attempt to illustrate the effects on rheology and fault-strength for a transcrustal fault zone that are likely to arise from fluid pressure cycling associated with fault-valve action. Both the depth and amplitude of the peak shear resistance become time-dependent. In addition, the integrated strength of the fault zone is at a minimum pre-failure and reaches its maximum value at the end of the post-failure discharge phase, before self-sealing occurs and fluid pressure starts to reaccumulate. Some of the implications of such a model for rupture nucleation and recurrence have been explored by Sibson (1992*b*)

Discussion

While static stress fields may induce directional permeability in the rock mass, textural evidence from hydrothermal veins hosted in fault/fracture systems suggests that fluid flux through such features is commonly episodic. Moreover, it is difficult to account for the great fluid mobility in the vicinity of fault zones in terms of static stress fields and constant permeability. It appears that in many circumstances fault and fracture permeability must continually be renewed for them to remain as effective channelways. Stress cycling thus appears as the great *modulator* of fluid flow in deforming upper crust, affecting flow systems driven by long-term hydraulic

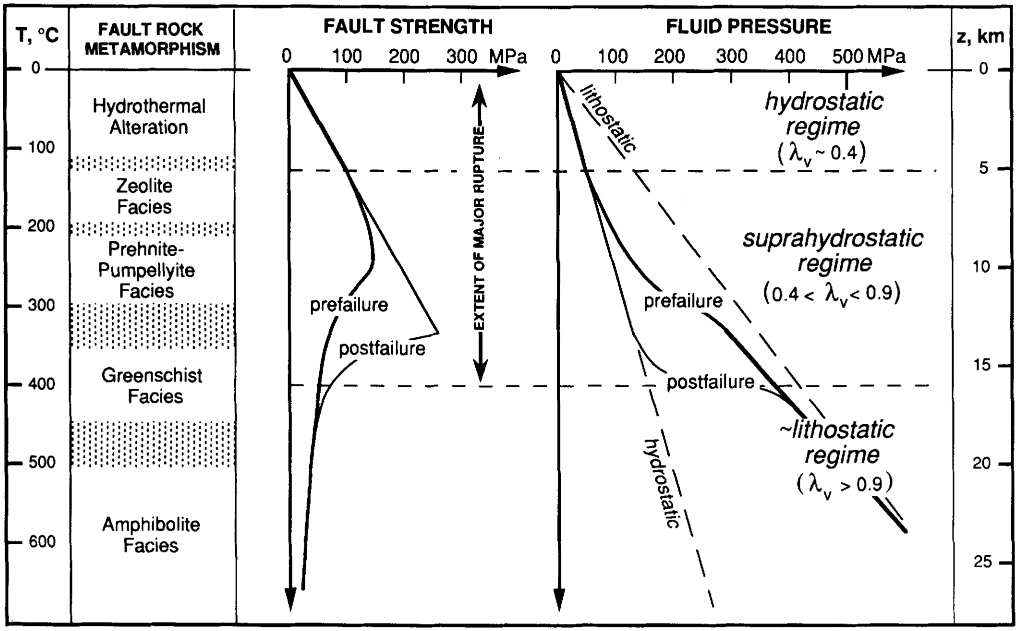

Fig. 9. General fault zone model incorporating fluid pressure cycling, assuming a linear geotherm of 25° C km^{-1}. The extent of the different fluid pressure regimes and their relationship to metamorphic environment is conjectural.

potentials such as topography, intrusive heat sources, metamorphic dewatering, etc.

A considerable range of mechanisms tied to the earthquake stress cycle may contribute to fluid redistribution around fault zones. 'Suction pump' effects at dilational jogs and bends have characteristic structural associations, as have the vein assemblages that result from 'extreme fault-valve action'. However, with the exception of hydrofracture dilatancy linked to extreme valve action, the relative importance for large-scale fluid redistribution of the various pumping mechanisms involving pre-failure dilatancy and post-failure fluid expulsion remains uncertain. From consideration of hydrothermal veining at the scale of individual outcrops, it may be difficult or impossible to discriminate between the different mechanisms for fluid redistribution. In evaluating the possibilities, critical account must be taken of the structural site, the mode of faulting at the time of hydrothermal deposition, the level of exposure, and the evidence for a predominantly hydrostatic or suprahydrostatic fluid pressure regime. Similar difficulties arise when attempting to attribute post-seismic discharge at the ground surface to a particular fluid redistribution mechanism (e.g. Sibson 1981; Rojstaczer & Wolf 1992; Muir Wood & King 1993). Systematic monitoring of discharge chemistry over extended time periods may help to distinguish varying depths of origin for the discharging fluids, and allow discrimination between the different mechanisms.

It is, however, becoming increasingly evident that fluid redistribution tied to the earthquake stress cycle has application to the development of fault-hosted mineralization and perhaps also to the migration of hydrocarbons in certain tectonic settings (e.g. Burley *et al.* 1989). Gold–quartz veins, especially, record the episodic passage of substantial volumes of aqueous fluid through fault zones and allied fracture networks (Cox *et al.* 1991; Boullier & Robert 1992). Regions near the top and the bottom of the seismogenic zone are favoured sites for the development of fault-hosted mineralization, with fluctuations in stress and permeability acting as triggers for episodes of hydrothermal precipitation at different stages of the fault loading cycle (Sibson 1992*a*). Rapid slip transfer across upwelling hydrothermal systems hosted within dilational jogs and bends in extensional and transtensional fault systems leads to abrupt pressure reductions, triggering episodes of boiling within the epithermal environment and mineral deposition throughout the phase of aftershock activity. Disseminated epithermal mineralization may also develop from the to-and-fro passage of fluids driven by cyclic loading within intensely fractured portions of the uppermost crust. At deeper levels in compressional/transpressional fault systems, intense valving action towards the base of the seismogenic zone gives rise to high-amplitude fluid pressure cycling and the formation of mesothermal gold–quartz lodes.

There is, therefore, mounting evidence for mechanical involvement of fluids with all stages of the earthquake cycle. In the lower half of the seismogenic zone, the competition between creation and destruction of fracture permeability plays a critical role in the accumulation of fluid overpressures, affecting earthquake nucleation and recurrence, while transient fluid pressure reductions at dilational jogs and bends contribute to rupture arrest and aftershock activity (Sibson 1992*b*). The cyclical character of fluid pressure gradients arising from fault-valve action on seismically active faults has the implication that rheological models of fault zones and crustal shear strength profiles must also be considered time-dependent, with integrated shear strength at a minimum pre-failure, but attaining a maximum value postfailure at the end of the discharge phase.

Many of the ideas expressed here have arisen through interaction and conversations over many years with N. Brown, J. Moore, H. Poulsen, N. Price, and F. Robert, while S. Cox, I. Main, J.-R. Grasso and an anonymous reviewer provided constructive criticism of the manuscript. I thank the organizers of Geofluids '93 for the financial support which made my participation in the conference possible.

References

ANDERSON, E.M. 1951. *The Dynamics of Faulting and Dyke Formation with Application to Britain*. 2nd edn, Oliver & Boyd, Edinburgh.

AYDIN, A. & JOHNSON, A.M. 1978. Development of faults as zones of deformation bands and as slip surfaces in sandstones. *Pure and Applied Geophysics*, **116**, 932–942.

BATZLE, M.L. & SIMMONS, G. 1977. Geothermal systems: rocks, fluids, fractures. *American Geophysical Union Monograph*, **20**, 233–242.

BETHKE, C.M. 1986. Hydrologic constraints on the genesis of the Upper Mississippi Valley mineral district from Illinois basin brines. *Economic Geology*, **81**, 233–249.

BORRAIDALE, G.J. 1981. Particulate flow and the generation of cleavage. *Tectonophysics*, **72**, 306–321.

BOULLIER, A-M. & ROBERT, F. 1992. Paleoseismic events recorded in Archean gold-quartz vein networks, Val d'Or, Abitibi, Quebec. *Journal of Structural Geology*, **14**, 161–179.

BRACE, W.F. 1980. Permeability of crystalline and argillaceous rocks. *International Journal of Rock Mechanics and Mining Science*, **17**, 241–251.

—— 1984. Permeability of crystalline rocks: new in situ measurements. *Journal of Geophysical Research*, **89**, 4327–4330.

BRANTLEY, S. 1992. The effect of fluid chemistry on quartz microcrack lifetimes. *Earth and Planetary Science Letters*, **113**, 145–156.

BURLEY, S.D., MULLIS, J. & MATTER, A. 1989. Timing diagenesis in the Tartan Reservoir (UK North Sea): constraints from combined cathodoluminescence microscopy and fluid inclusion studies. *Marine and Petroleum Geology*, **6**, 98–120.

BYERLEE, J.D. 1978. Friction of rocks. *Pure and Applied Geophysics*, **116**, 615–626.

—— 1990. Friction, overpressure and fault-normal compression. *Geophysical Research Letters*, **17**, 2109–2112.

—— 1993. Model for episodic flow of high-pressure water in fault zones before earthquakes. *Geology*, **21**, 303–306.

CARRIO-SCHAFFHAUSER, E., RAYNAUD, S., LATIERE, H.J. & MAZEROLLE, F. 1990. Propagation and localisation of stylolites in limestones. *In*: KNIPE, R.J. & RUTTER, E.H. (eds) *Deformation Mechanisms, Rheology and Tectonics*. Geological Society, London, Special Publications, **54**, 193–199.

CATHLES, L.M. 1990. Scale and effects of fluid flow in the upper crust. *Science*, **248**, 323–329.

CELLO, G. & NUR, A. 1988. Emplacement of foreland thrust systems. *Tectonics*, **7**, 261–271.

CHESTER, F.M., EVANS, J.P. & BIEGEL, R.L. 1993. Internal structure and weakening mechanisms of the San Andreas Fault. *Journal of Geophysical Research*, **98**, 771–786.

COX, S.F. & *Etheridge, M.A.* 1989. Coupled grain-scale dilatancy and mass transfer during deformation at high fluid pressures: examples from Mount Lyell, Tasmania. *Journal of Structural Geology*, **11**, 147–162.

——, WALL, V.J., ETHERIDGE, M.A. & POTTER, T.F. 1991. Deformational and metamorphic processes in the formation of mesothermal vein-hosted gold deposits – examples from the Lachlan Fold Belt in central Victoria, Australia. *Ore Geology Reviews*, **6**, 391–423.

CRAMPIN, S., EVANS, R. & ATKINSON, B.K. 1984. Earthquake prediction: a new physical basis. *Geophysical Journal of the Royal Astronomical Society*, **76**, 147–156.

DAS, S. & SCHOLZ, C.H. 1981. Theory of time-dependent rupture in the earth. *Journal of Geophysical Research*, **86**, 6038–6051.

DU ROUCHET, J. 1981. Stress fields, a key to oil migration. *American Association of Petroleum Geologists Bulletin*, **65**, 74–85.

ETHERIDGE, M.A., WALL, V.J., COX, S.F. & VERNON, R.H. 1984. High fluid pressures during regional metamorphism and deformation: implications for mass transport and deformation mechanisms. *Journal of Geophysical Research*, **89**, 4344–4358.

FISCHER, G.J. & PATERSON, M.S. 1989. Dilatancy during rock deformation at high temperatures and pressures. *Journal of Geophysical Research*, **94**, 17,607–17,617.

FLETCHER, R.C. & POLLARD, D.D. 1981. Anticrack model for pressure solution surfaces. *Geology*, **9**, 419–424.

FYFE, W.S. 1990. Plate tectonics and the hydrosphere. *In*: NESBITT, B.E. (ed.) *Fluids in Tectonically Active Regimes of the Continental Crust*. Mineralogical Association of Canada Short Course Handbooks, **18**, 299–312.

——, PRICE, N.J. & THOMPSON, A.B. 1978. *Fluids in the Earth's Crust*, Elsevier, Amsterdam.

GOLD, T. & SOTER, S. 1984. Fluid ascent through the solid lithosphere and its relation to earthquakes. *Pure and Applied Geophysics*, **122**, 492–529.

GRINDLEY, G.W. & BROWNE, P.R.L. 1976. Structural and hydrological factors controlling the permeability of some hot-water geothermal fields. *In*: *Proceedings of the 2nd United Nations Symposium on Development and Use of Geothermal Resources*. New York, 377–386.

GROSHONG, R.H. 1988. Low-temperature deformation mechanisms and their interpretation. *Geological Society of America Bulletin*, **100**, 1329–1360.

HICKMAN, S.H. 1991. Stress in the lithosphere and the strength of active faults. *Review of Geophysics Supplement*, **29**, 759–775.

HILL, D.P. 1977. A model for earthquake swarms. *Journal of Geophysical Research*, **82**, 1347–1352.

——, REASENBERG, P.A. & 29 OTHERS 1993. Seismicity remotely triggered by the magnitude 7.3 Landers earthquake. *Science*, **260**, 1617–1623.

HOBBS, B.E., MULHAUS, H-B. & ORD, A. 1990. Instability, softening, and localization of deformation. *In*: KNIPE, R.J. & RUTTER, E.H. (eds) *Deformation Mechanisms, Rheology and Tectonics*, Geological Society, London, Special Publications, **54**, 143–166.

HOOPER, E.C.D. 1991. Fluid migration along growth faults in compacting sediments. *Journal of Petroleum Geology*, **14**, 161–180.

HUBBERT, M.K. & RUBEY, W.W. 1959. Role of fluid pressure in the mechanics of overthrust faulting. *Geological Society of America Bulletin*, **70**, 115–205.

HULIN, C.D. 1925. Structural control of ore deposits. *Economic Geology*, **24**, 15–49.

HUNT, J.M. 1990. Generation and migration of petroleum from abnormally pressured fluid compartments. *American Association of Petroleum Geologists Bulletin*, **74**, 1–12.

IRWIN, W.P. & BARNES, I. 1980. Tectonic relations of carbon dioxide discharges and earthquakes. *Journal of Geophysical Research*, **85**, 3115–3121.

KERRICH, R. 1986. Fluid infiltration into fault zones: chemical, isotopic, and mechanical effects. *Pure and Applied Geophysics*, **124**, 226–268.

KNIPE, R.J. 1989. Deformation mechanisms – recognition from natural tectonites. *Journal of Structural Geology*, **11**, 127–146.

KNOPF, A. 1929. *The Mother Lode System of California*. US Geological Survey Professional Paper, **157**.

LACHENBRUCH, A.H. & MCGARR, A. 1990. *In:* WALLACE, R.E. (ed.) The San Andreas Fault System, California. *Stress and Heat Flow*. US Geological Survey Professional Paper, **1515**, 261–277.

MCCAIG, A.M. 1988. Deep fluid circulation in fault zones. *Geology*, **16**, 867–870.

MCCLAY, K. 1977. Pressure solution and Coble creep in rocks. *Journal of the Geological Society, London*, **134**, 57–70.

MARONE, C., RALEIGH, C.B. & SCHOLZ, C.H. 1990. Frictional behavior and constitutive modeling of simulated fault gouge. *Journal of Geophysical Research*, **95**, 7007–7025.

MORROW, C., LOCKNER, D., MOORE, D. & BYERLEE, J. 1981. Permeability of granite in a temperature gradient. *Journal of Geophysical Research*, **86**, 3002–3008.

MUIR WOOD, R. & KING, G.C.P. 1993. Hydrological signatures of earthquake strain. *Journal of Geophysical Research*, **98**, 22035–22068.

MOUNT, V.S. & SUPPE, J. 1987. State of stress near the San Andreas fault: implications for wrench tectonics. *Geology*, **15**, 1143–1146.

MULLIS, J. 1988. Rapid subsidence and upthrusting in the Northern Appenines, deduced by fluid inclusion studies in quartz crystals from Porretta Terme. *Schweizerische Mineralogische und Petrographische Mitteilungen*, **68**, 157–170.

NEWHOUSE, W.H. 1942. *Ore Deposits as Related to Structural Features*. Princeton University Press, Princeton, New Jersey.

NICHOLSON, C. & WESSON, R.L. 1990. *Earthquake hazard associated with deep well injection – a report to the U.S. Environmental Protection Agency*. US Geological Survey Bulletin, **1951**.

NORTON, D.L. 1982. Fluid and heat transport phenomena typical of copper-bearing pluton environments, south-eastern Arizona. *In*: TITLEY, S.R. (ed.) *Advances in Geology of the Porphyry Copper Deposits, Southwestern North America*. University of Arizona Press, Tucson Arizona, 59–72.

NUR, A. 1972. Dilatancy, pore fluids, and premonitory variations of t_s/t_p travel times. *Bulletin of the Seismological Society of America*, **62**, 1217–1222.

—— 1975. A note on the constitutive law for dilatancy. *Pure and Applied Geophysics*, **133**, 197–206.

—— & BOOKER, J.R. 1972. Aftershocks caused by pore-fluid flow? *Science*, **175**, 885–887.

—— & WALDER, J. 1990. Time-dependent hydraulics of the Earth's crust. *In*: *The Role of Fluids in Crustal Processes*. National Academy of Sciences, Washington, DC, 113–127.

OLIVER, J. 1986. Fluids expelled tectonically from orogenic belts: their role in hydrocarbon migration and other geologic phenomena. *Geology*, **14**, 99–102.

PARRY, W.T. & BRUHN, R.L. 1990. Fluid pressure transients on seismogenic normal faults. *Tectonophysics*, **179**, 335–344.

POWER, W.L., TULLIS, T.E., BROWN, S., BOITNOTT, G.N. & SCHOLZ, C.H. 1987. Roughness of natural fault surfaces. *Geophysical Research Letters*, **14**, 29–32.

POWLEY, D.E. 1990. Pressures and hydrogeology in petroleum basins. *Earth Science Reviews*, **29**, 215–226.

RAMSAY, J.G. 1980. The crack-seal mechanism of rock deformation. *Nature*, **284**, 135–139.

REDWINE, L. 1981. Hypothesis combining dilation, natural hydraulic fracturing, and dolomitization to explain petroleum reservoirs in Monterey Shale, Santa Maria area, California. *In*: GARRISON, R.E., DOUGLAS, R.G., PISCIATTO, K.E., ISAACS, C.M. & INGLE, J.C. (eds) *The Monterey Formation and Related Siliceous Rocks of California*. Special Publications of the Society of Economic Paleontologists and Mineralogists, 221–248.

RICE, J.R. 1992. Fault stress states, pore pressure distributions, and the weakness of the San Andreas fault. *In*: EVANS, B. & WONG, T-F. (eds) *Fault Mechanics and Transport Properties of Rocks*. Academic Press, New York, 475–503.

—— & CLEARY, M.P. 1976. Some basic stress diffusion solutions for fluid-saturated elastic porous media with compressible constituents. *Reviews of Geophysics and Space Physics*, **14**, 227–241.

ROBERT, F. & KELLY, W.C. 1987. Ore-forming fluids in Archean gold-bearing quartz veins at the Sigma Mine, Abitibi greenstone belt, Quebec, Canada. *Economic Geology*, **82**, 56–74.

ROBERTS, G. 1991. Structural controls on fluid migration through the Rencurel thrust zone, Vercors, French Sub-Alpine Chains. *In*: ENGLAND, W.T. & FLEET, A.J. (eds) *Petroleum Migration*. Geological Society, London, Special Publications, **59**, 245–262.

ROJSTACZER, S. & WOLF, S. 1992. Permeability changes associated with large earthquakes: an example from Loma Prieta, California. *Geology*, **20**, 211–214.

RUDNICKI, J.W. & HSU, T-C. 1988. Pore pressure changes induced by slip on permeable and impermeable faults. *Journal of Geophysical Research*, **93**, 3275–3285.

SCHOLZ, C.H., SYKES, L.R. & AGGARWAL, Y.P. 1973. Earthquake prediction: a physical basis. *Science*, **181**, 803–810.

SECOR, D.T. 1965. Role of fluid pressure in jointing. *American Journal of Science*, **263**, 633–646.

—— & POLLARD, D.D. 1975. On the stability of open hydraulic fractures in the earth's crust. *Geophysical Research Letters*, **2**, 510–513.

SEGALL, P. & POLLARD, D.D. 1980. Mechanics of discontinuous faults. *Journal of Geophysical Research*, **85**, 4337–4350.

SIBSON, R.H. 1974. Frictional constraints on thrust, wrench and normal faults. *Nature*, **249**, 542–544.

— 1981. Fluid flow accompanying faulting: field evidence and models. *In*: SIMPSON, D.W. & RICHARDS, P.G. (eds) *Earthquake Prediction: an International Review*. American Geophysical Union, Maurice Ewing Series, **4**, 593–603.

—— 1983. Continental fault structure and the shallow earthquake source. *Journal of the Geological Society, London*, **140**, 741–767.

—— 1985. Stopping of earthquake ruptures at dilational fault jogs. *Nature*, **316**, 248–251.

—— 1987. Earthquake rupturing as a hydrothermal mineralising agent. *Geology*, **15**, 701–704.

—— 1989. Earthquake faulting as a structural process. *Journal of Structural Geology*, **11**, 1–14.

—— 1990*a*. Rupture nucleation on unfavorably oriented faults. *Bulletin of the Seismological Society of America*, **80**, 1580–1604.

—— 1990*b*. Conditions for fault-valve behaviour. *In*: KNIPE, R.J. & RUTTER, E.H. (eds) *Deformation Mechanisms, Rheology and Tectonics*. Geological Society, London, Special Publications, **54**, 15–28.

—— 1991. Loading of faults to failure. *Bulletin of the Seismological Society of America*, **81**, 2493–2497.

—— 1992*a*. Earthquake faulting, induced fluid flow, and fault-hosted gold-quartz mineralization. *In*: BARTHOLOMEW, M.J., HYNDMAN, D.W., MOGK, D.W. & MASON, R. (eds) *Basement Tectonics 8: Characterization and Comparison of Ancient and Mesozoic Continental Margins*. Proceedings of the 8th International Conference on Basement Tectonics, Butte, Montana, Kluwer, Dordrecht, 603–614.

—— 1992*b*. Implications of fault-valve behaviour for rupture nucleation and recurrence. *Tectonophysics*, **211**, 283–293.

—— 1993. Load-strengthening versus load-weakening faulting. *Journal of Structural Geology*, **15**, 123–128.

——, MOORE, J.MCM. & RANKIN, A.H. 1975. Seismic pumping – a hydrothermal fluid transport mechanism. *Journal of the Geological Society, London*, **131**, 653–659.

——, ROBERT, F. & POULSEN, K.H. 1988. High-angle reverse faults, fluid pressure cycling and mesothermal gold-quartz deposits. *Geology*, **16**, 551–555.

SLEEP, N.H. & BLANPIED, M.L. 1992. Creep, compaction, and the weak rheology of major faults. *Nature*, **359**, 687–692.

SMITH, D.A. 1980. Sealing and non-sealing faults in Louisiana Gulf Coast Salt Basin. *American Association of Petroleum Geologists Bulletin*, **64**, 145–172.

SMITH, D.L.& EVANS, B. 1984. Diffusional crack healing in quartz. *Journal of Geophysical Research*, **89**, 4125–4135.

SPRUNT, E.S. & NUR, A. 1977. Destruction of porosity through pressure solution. *Geophysics*, **42**, 726–741.

STARK, C.P. & STARK, J.A. 1991. Seismic fluids and percolation theory. *Journal of Geophysical Research*, **96**, 8417–8426.

TORGERSON, T. 1990. Crustal-scale fluid transport: magnitude and mechanisms. *EOS, Transactions of the American Geophysical Union*, **71**, 1, 4, 13.

—— 1991. Crustal fluid flow: continuous or episodic. *EOS, Transactions of the American Geophysical Union*, **72**, 18–19.

WALDER, J. & NUR, A. 1984. Porosity reduction and crustal pore pressure development. *Journal of Geophysical Research*, **89**, 11,539–11,548.

WOOD, S.H., WURTS, C., BALLENGER, N., SHALEEN, M. & TOTORICA, D. 1985. The Borah Peak, Idaho, earthquake of October 28, 1983: hydrologic effects. *Earthquake Spectra*, **2**, 125–150.