Dynamics of fluid flow and fluid chemistry during crustal shortening

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“It is not the mountain we conquer but ourselves.” - Sir Edmund Hillary

To Mum and Dad - thank you for everything
Declaration

The work in this thesis is my own except where otherwise stated.

Shaun L.L. Barker
Abstract

In this thesis, an integrated structural and chemical approach has been used to investigate the spatial and temporal evolution of fluid chemistry, and fluid flow pathways, during crustal shortening. The Taemas Vein Swarm is hosted in a limestone-shale sequence, the Murrumbidgee Group, in the Eastern Belt of the Lachlan Orogen, in New South Wales, Australia. The Taemas Vein Swarm (TVS) is composed of calcite ± quartz veins, hosted in a series of faults and fractures, which extends over an area of approximately 20 km$^2$. The Murrumbidgee Group is composed of several formations, comprising massive grey micritic limestones, redbed sandstones and shales, and thinly interbedded (10–20 cm scale) limestones and shales.

The sedimentary sequence has been folded into a series of upright, open to close folds, and was probably deformed during either mid-late Devonian, or early Carboniferous, crustal shortening. To the east, the Murrumbidgee Group is overthrust by a Silurian volcanic sedimentary sequence along the Deakin-Warroo Fault System. Crosscutting and overprinting relationships demonstrate that vein growth was synchronous with folding, with different vein types related to different fold mechanisms at various stages of fold growth.

Flexural slip folding led to the development of bedding-concordant veins (hereafter called bedding-parallel veins). Flexural flow in semicompetent to incompetent beds caused en echelon extension vein arrays to grow. Decoupling between beds, and dilatancy at fold hinges led to significant vein growth. In addition, fold lock-up led to limb-parallel stretching, and the growth of bedding-orthogonal extension fractures.

Vein growth is inferred to have occurred in a compressional tectonic regime (i.e. $\sigma_3=$vertical). Oxygen isotope quartz-calcite thermometry suggests that veins formed at temperatures of 100–200 °C. The depth of vein formation may have been between about 5 and 8 km. Vein textures indicate that growth of veins occurred during multiple cycles of permeability enhancement and destruction. Subhorizontal extension fractures, and faults at unfavourable angles for reactivation, imply that fluid pressures exceeded lithostatic levels during the growth of some veins. Coexisting extension and shear fractures imply that differential stress levels varied over time.

Flexural slip continued throughout folding at Taemas, despite some fold limbs being at angles extremely unfavourable for reactivation ($\theta > 60^\circ$). As folds approached frictional lock-up, flexural slip continued to occur when supralithostatic fluid pressures were developed. Therefore, large, bedding-discordant faults were not developed to accommodate strain during folding, explaining a deficiency of larger faults in the TVS.

Infiltration of overpressured fluids occurred into the base of the Murrumbidgee Group, and was channelled into a distributed mesh of small faults and fractures.
At the point that a connected ‘backbone’ flow network developed in the TVS, high-pressure fluids would no longer be available to allow continuing flexural slip on fold limbs approaching lockup. Thereafter, larger faults would develop, which would adjust the fault population in the TVS to a more ‘typical’ displacement-frequency distribution. This had not occurred in the Taemas area by the time crustal shortening ceased. An abundance of small faults, and fracturing driven by invasion of overpressured fluid, implies that the TVS formed via an ‘earthquake swarm’ process.

Modern analytical techniques, utilising laser ablation sampling technology, allow high-spatial resolution chemical data to be collected from syntectonic veins. Insights into the role that fluid-mineral interface processes may have on the chemistry of minerals grown in syntectonic veins were provided by an experimental study. Moderate sized (< 1 – 5 mm) synthetic calcite crystals were successfully grown to investigate the uptake of rare earth elements (REE) into calcite. Changes in crystal morphology are linked to variable solution chemistry, which has important implications for the interpretation of hydrothermal vein textures. High-spatial resolution chemical analyses of synthetic calcite crystals demonstrate significant fluctuations in REE concentrations over distances of < 200 µm within calcite crystals. Time-equivalent regions on different crystal faces have significantly different REE concentrations, indicating that fluctuations in calcite trace element composition cannot be interpreted exclusively in terms of changing ‘bulk fluid’ composition. Rare earth element anomalies (Eu/Eu* and Ce/Ce*) are not significantly influenced by compositional zoning, and may be robust indicators of changes in solution bulk chemistry and fluid oxidation state.

Changes in isotopic ratios (δ¹³C, δ¹⁸O and ⁸⁷Sr/⁸⁶Sr), and trace element concentrations in veins from the TVS are related to variations in fluid source, flow pathways and chemical conditions (e.g. trace element complexation, precipitation rate, fluid oxidation) during hydrothermal fluid flow. This integrated structural, textural and chemical approach has direct application to the examination of hydrothermal veins in fracture-hosted ore deposits, and may allow the fluid source and/or chemical conditions conducive to the formation of high-grade ore to be discerned.

Vein δ¹⁸O compositions systematically increase upwards through the Murrumbidgee Group, caused by progressive reaction of an externally derived, low-δ¹⁸O fluid (of probable meteoric origin) with host limestones. Vein δ¹⁸O and ⁸⁷Sr/⁸⁶Sr compositions vary spatially and temporally within the same outcrop, and within individual veins, which is inferred to be caused by the ascent of packages of fluid along constantly changing flow pathways. Fluid-buffered oxygen isotope ratios at the earliest stages of deformation imply that the TVS formed via an ‘invasion percolation’ process. Fluid pathways are inferred to have changed constantly, with fractures ‘toggle-switching’ between high-permeability and low-permeability states, due to repeated fracture opening and sealing events.
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Crustal fluid flow and hydrothermal mineralisation

Crustal fluid flow is increasingly recognised as being critical to numerous processes in the Earth, and is significant for many economic, scientific and environmental reasons. The formation of hydrothermal ore deposits and petroleum resources is dependent on fluid migration and focussed flow into localised areas (see reviews of Lonergan *et al.* 1999; Sibson 2001; Wilkinson 2001). Other examples of processes where crustal fluid migration is important include the extraction of potable water from subterranean sources, and the disposal of contaminated material (*e.g.* radioactive nuclear waste), both of which require a thorough understanding of the mechanism and rates of fluid migration through the shallow crust (Talbot 1999; Berkowitz 2002; Stuckless and Dudley 2002).

Seismological and field-based research over the last 35 years has demonstrated that fluid migration is commonly associated with active fault zones and earthquake processes (*e.g.* Kanamori 1972; Scholz *et al.* 1973; Sibson *et al.* 1975 1988; Cox 1995; Hickman *et al.* 1995; Husen and Kissling 2001). Recent work has raised the intriguing possibility that fluid chemistry changes in and around active fault zones could provide a useful tool for earthquake-prediction studies (Claesson *et al.* 2004).

The examples above demonstrate the importance of fluid migration and associated material transport through the crust. Ancient hydrothermal systems (now exposed at the Earth’s surface) are the integrated result of hydrothermal activity occurring at depth, many aspects of which cannot be directly observed or measured in active hydrothermal systems. Thus, exhumed paleohydrothermal systems provide an opportunity to examine the underlying controls on fluid migration in the Earth, and the effects that these fluids have on surrounding rocks. One major drawback with the study of paleohydrothermal systems is discerning their temporal and spatial evolution.

This study focuses on a regional scale (> 20 km²), exhumed hydrothermal system, which is preserved in a folded limestone-shale sequence (the Murrumbidgee Group) in the Taemas area, southeastern New South Wales, Australia. The structural setting (*i.e.* the relationships to other veins, and hosting folds and faults) and textures of veins reflect how stress and fluid pressure states evolved through time and space during crustal shortening. In addition, the veins contain a record of the temporal
and spatial evolution of fluid chemistry. This exhumed fault-fracture system provides an opportunity to examine the evolution of fluid pressures, fluid pathways and fluid chemistry during crustal shortening.

1.1 Fracture-controlled fluid flow in rocks

The migration of fluid through rocks requires a rock mass with pore space (containing fluid) and pore connectivity (allowing fluid to move). Pore spaces may be intergranular pores, or grain-scale to macroscopic-scale fractures. The size, number and connectivity of pores will affect the permeability of a rock (Cox, 2005). Permeability (units of $m^2$) is an intrinsic rock property, quantifying the ability of fluid to pass through a rock. Rocks have a vast range of permeabilities. In an interlayered sedimentary sequence, which may consist of shale, sandstone, limestone and dolomite, permeabilities may range between $10^{-20}$ m$^2$ and $10^{-13}$ m$^2$ (Brace, 1980).

Intact, compacted and well cemented sedimentary rocks will generally have relatively low permeabilities. However, the formation of fractures can establish high permeability pathways. Linked fractures will generate permeability orders of magnitude higher than that of their surrounding rocks (Pedersen et al., 1997). Fluid fluxes and flow geometries will be highly dependent on the permeability structure of the rocks through which the fluids are migrating.

Linked fracture sets (which provide high permeability pathways) manifest as faults in the upper to middle continental crust (e.g. 2 – 15 km). Fault slip in the upper continental crust is dominated by earthquake rupturing (Sibson, 2001). Exhumed fault zones provide ample evidence for past fluid flow, with many hydrothermal mineral deposits linked to faults and associated fracture systems (see Sibson, 1987; Parry, 1998; Sibson, 2001; Cox, 2005 for reviews).

Time dependent variations in permeability will change the pattern of fluid flow in and around fault zones (Cox et al., 2001). For example, high permeability pathways will be created as seismic slip causes fracturing in and around fault zones. The development and longevity of high-permeability pathways will be counteracted by the development of low-permeability fault gouge and hydrothermal mineral deposition (Sibson, 2001). These competing processes lead to intrinsically variable permeability in the crust. Syntectonic vein textures provide evidence for cyclic permeability enhancement and destruction, with repetitive banding in crack-seal veins commonly interpreted as a result of episodic fracturing, fluid flow and fracture-sealing events (Sibson et al., 1988; Boullier and Robert, 1992; Cox et al., 1995; Nguyen et al., 1998).

Seismic imaging after earthquakes suggests that significant fluid flow occurs in and around active fault zones (Husen and Kissling, 2001; Braeuer et al., 2003; Miller et al., 2004; Kao et al., 2005). Further evidence for a link between seismic activity and fluid migration is provided by post-seismic discharge of fluids in the vicinity of shallow faults (Muir-Wood and King, 1993). Both ancient, exhumed fault zones, and active fault zones provide evidence that fracture formation, seismic activity and fluid migration are intrinsically linked. Coupling between deformation processes, fluid flow, and fluid pressure states plays a key role in driving the growth of fluid
1.2. The ‘Taemas Vein Swarm’

The common occurrence of fault-fracture networks comprised of low-displacement shear and extension fractures places key constraints on the shear stress levels driving these systems, and suggests that fluid pressures (at least locally and transiently) exceed the least principal stress. This observation is extremely important, as it suggests that fluid pressures may reach, and possibly exceed, lithostatic pressures at depth in the crust, particularly in reverse fault systems (Sibson, 2001). Migration of high-pressure fluids through fault-fracture networks has a significant influence on fault strength, via the principle of effective stress (Hubbert and Rubey, 1959). Syntectonic veins record information on spatial and temporal variations in fluid compositions and fluid sources during progressive vein growth (e.g., Dietrich et al., 1983; Rye and Bradbury, 1988). Exhumed vein systems contain a record of where fluid flow was localised, how flow has influenced the mechanical behaviour of the crust, and the nature of chemical reactions between rocks and fluids along fluid flow pathways.

**1.2 The ‘Taemas Vein Swarm’**

The Taemas area is located southwest of Yass, in the Eastern Belt of the Lachlan Orogen, in eastern New South Wales, Australia (Glen, 1992). Here, a major fold triplet (Wee Jasper Syncline, Narrangullen Anticline, Taemas Synclinorium) form the larger Black Range Synclinorium (approximately 180 km long, 25 km wide). The Taemas Synclinorium is a doubly-plunging synclinal structure approximately 5 km wide, lying on the eastern limb of the Narrangullen Anticline (Cramsie et al., 1975). Within the Taemas Synclinorium, sedimentary rocks of the Murrumbidgee Group have been folded into upright, open to close folds (wavelengths varying between tens-of-metres to kilometre scale), which have a predominant north-northwest - south-southeast trend, and steep axial surfaces. The synclinorium is in the footwall of the major, steeply-dipping, Warroo and Deakin - Devil’s Pass Fault system (Fig. 1.1). Here, Silurian volcanics overthrust the Murrumbidgee Group sediments (Cramsie et al., 1975).

Within the Taemas Synclinorium, the Murrumbidgee Group is approximately 1 km thick, and is composed of several formations, comprising massive grey micritic limestones, redbed sandstones and shales, and thinly interbedded (10–20 cm scale) limestones and shales, which were deposited during the middle Devonian (Browne, 1958).

An extensive fault-fracture network (hereafter called the ‘Taemas Vein Swarm’, or TVS) composed of calcite ± quartz, and minor fluorite and barite is developed throughout most of the Murrumbidgee Group in the Taemas area. The vein swarm is developed mainly in the northern part of the synclinorium, over an area of approximately 20 km². Veining and folding were synchronous, with a variety of fault veins, bedding-parallel slip veins and extension veins developed. Laminated veins related to bedding-parallel slip are common, particularly in thinly interbedded limestone-mudstone units. Bedding-discordant faults, fault-related extension veins, and ex-
tension veins related to flexural flow and bedding-parallel limb stretching are also found.

The age of fold, fault and vein formation at Taemas is poorly constrained. Stratigraphic relationships between the Murrumbidgee Group and overlying Middle Devonian Hatchery Creek Formation (30 km west of Taemas) are uncertain (see [Hood and Durney, 2002] for a review). Possible conformable relationships between these units imply that regional deformation may have occurred during the Kanimbil Orogeny (early Carboniferous age, approximately 325 Ma; [Hood and Durney, 2002]), younger than the mid-late Devonian age proposed by [Browne, 1958].

**Figure 1.1:** (a) Location of Taemas Vein Swarm in relation to the southeastern Australian Coast. (b-c) Taemas Station, with Taemas Peninsula (site of the TVS) immediately to the north, with Yass and Canberra. (d) Simplified regional geology map showing the Taemas Synclinorium distribution of the Murrumbidgee Group sediments (blue) and underlying volcanic and terrestrial sediments of the Black Range Group (red). Taemas is at approximately 35°S, 148°50’ E.
1.3 Thesis overview

The aim of this thesis is to explore the dynamics of fluid flow and fluid-rock reaction in a fracture-controlled hydrothermal system during regional crustal shortening and associated folding and faulting. This thesis examines the structural setting and timing relationships of different vein types (e.g., extension veins, fault veins, bedding-parallel laminated veins), and examines changes in vein chemistry over a variety of scales (from kilometres to microns). Chemical variations throughout the stratigraphic sequence (100’s of metres to kilometre scale), within individual outcrops (metre to tens of metre scale) and within individual veins (micron to centimetre scale) have been explored. These chemical variations have been related to the structural setting of veins (i.e., larger folds and faults) to examine the development of fluid flow pathways, and the evolution of fluid composition during crustal shortening.

Links between folding, fracturing, permeability evolution, fluid migration and associated vein formation are discussed in Chapter 2. As a basis for the exploration of vein chemistry, a review of trace element incorporation into calcite is presented in Chapter 3. This literature review is followed by a description of some simple calcite growth experiments, which explore the morphology and chemistry of calcite crystals precipitated from solution.

Chapter 4 provides a summary of the geological backdrop to the TVS, and the structural setting of different vein types within the TVS are illustrated using outcrop photographs and sketches. Macroscopic variations in the chemistry of the Taemas Vein Swarm and associated host rocks is examined in Chapter 5.

A major component of this thesis is the exploration of the record of fluid chemistry variations contained within individual syntectonic veins. In Chapter 6, high-spatial resolution (≈ 100–200 µm) analyses of trace element concentrations, and Sr, C and O isotope ratios in different veins are presented. Chapter 7 describes the application of the samarium-neodymium isotope system to explore the dynamics of fluid flow, and potential of Sm-Nd as a chronological tool in hydrothermal systems.

Finally, Chapter 8 draws together structural and chemical data presented in earlier chapters to explore the dynamics of fluid flow in hydrothermal systems, and implications of changes in fluid pressure and fluid chemistry for dynamics of growth of fracture-controlled flow systems, seismic processes, and ore formation in fault-hosted deposits.
Chapter 2
Folding, fracturing and fluid flow

2.1 Fold mechanisms

Different sedimentary rocks have varying mechanical properties (or ‘competence’). Qualitatively, competence describes the relative rate at which a ductile material flows at a given differential stress (Twiss and Moores, 1992). The difference in competence between two materials is termed ‘competence contrast’ ($\Delta C$). If a layered sedimentary sequence of different rocks (e.g. limestones and shales) is forced to deform at the same rate, the more competent material will have a higher differential stress.

The fold mechanisms described below may form distinct outcrop to microscale structures. The way in which a rock deforms will have a significant control on the rock structures (e.g. veins, cleavage, faults) contained within that rock.

2.1.1 Flexural folding of a layer

For flexural folding (class IB folds, see Ramsay and Huber, 1987) the orthogonal thickness of the layer remains constant during folding. However, the fold mechanisms may differ considerably:

1. During orthogonal flexure, all lines that were perpendicular to the layer before folding remain perpendicular after folding. Orthogonal flexure is usually characteristic of folds with low curvature developed in competent layers (e.g. those resistant to ductile deformation, see Fig. 2.1). As curvature increases, the orthogonal condition cannot be preserved.

2. In well bedded rocks, folding is commonly achieved by simple shear parallel to the planar structure. This may be occur as a:

   - **Flexural Flow Fold** – the simple shear displacement is distributed continuously throughout the structure. Flexural flow folds are generally formed where rocks have fairly penetrative and uniform planar fabric (e.g. shales).

   - **Flexural Slip Fold** – the simple shear displacement is distributed discontinuously through the structure. Flexural slip folds commonly occur
2. Folding, fracturing and fluid flow

![Diagram of orthogonal flexure before and after folding. Note that right angles remain right angles before and after folding. Dashed line represents neutral surface, e.g. the layer which does not change length during folding.](image)

**Figure 2.1:** Geometry of orthogonal flexure before and after folding. Note that right angles remain right angles before and after folding. Dashed line represents neutral surface, e.g. the layer which does not change length during folding. (after Ramsay and Huber, 1987; Twiss and Moores, 1992).

where distinct lithological layers are separated by well delineated bedding surfaces).

3. **Volume-loss folding** occurs when folds are formed and/or amplified by the removal of material in a folded layer. Volume-loss folding does not produce a particular class of fold, and any class I or class 2 fold geometry may be formed. Volume-loss may result in offset of beds, even though no shearing has occurred (see Fig. 2.2).

![Diagram of volume-loss folding.](image)

**Figure 2.2:** Schematic sketch of volume-loss folding. Upper diagrams show initial folds, and shaded regions are removed during fold growth. (a) Wedge-shaped areas of volume loss are symmetric about dashed lines that are normal to layer surface. (b) Wedge-shaped areas of volume loss are symmetric about dashed lines (not normal to layer surface). Outer arc of the resulting fold is smooth. However, inner arc surfaces have offsets, implying shear along layer in which volume-loss occurred. (c) Lath-shaped areas of volume loss are parallel to one another, and oblique to layer. Both outer arc and inner arc surfaces are offset, similar to passive-shear folding (after Twiss and Moores, 1992).

### 2.1.2 Folding of multilayers

Folding of sediments usually involves multilayered rock sequences. Commonly these layers will have different mechanical properties, leading to more complex fold geometries. Different fold mechanisms (e.g. flexural flow vs. flexural slip folding)
2.2 Fracture formation

may occur within the multilayer sequence as it is deformed. The competence contrast between layers will determine what fold mechanisms occur, thus controlling the structures developed in the rocks (Fig. 2.3). Additionally, the relative thickness of competent and incompetent layers will affect the fold mechanisms (Ramsay and Huber, 1987).

Some layered sequences have layers composed of materials with essentially the same high competence. If the friction between layers is relatively low, then slip may occur between layers (i.e. flexural slip). During flexural slip, the layer on the convex side must slip toward the fold hinge relative to the layer on the concave side (see Fig. 2.4a,d). Slip is greatest at the limbs, and decreases to zero at the hinge line, where it reverses shear sense. This slip commonly generates linear striations or mineral slickenfibers perpendicular to the fold axis on the bedding surface. Such folds have been described in many places including southwest England (Ramsay, 1974), South Georgia (Tanner, 1989), and southeastern Australia (Fowler, 1996).

If some degree of flexural flow occurs during folding, then less strain is accommodated via interlayer slip. Many folds consist of interlayered competent and incompetent layers of comparable thickness (medium average competence; high ΔC). Incompetent layers will tend to deform via flexural flow, while competent layers deform via flexural slip (Fig. 2.3). This will lead to a variety of resulting rock structures, with contrasting cleavage formation and fracture development throughout the folded sequence (Fig. 2.4 §2.3; note that § refers to a numbered section in the text).

2.2 Fracture formation

Different fracture types form under different stress and fluid pressure conditions. The occurrence of these different fractures in a body of rock may be used to infer the paleostress and fluid pressure conditions in those rocks. Here, various brittle
fractures are classified, and the stress conditions (magnitude and orientation) under which these fractures form are described. Additionally, the critical importance that fluid pressure plays on the formation of fractures is highlighted. This section will be used as a basis for understanding the location and relative timing of different fractures formed during deformation of the Murrumbidgee Group.

### 2.2.1 Macroscopic brittle failure

Cracks in rocks occur over a vast range of scales. Large structures such as the San Andreas or Alpine Faults extend for hundreds of kilometres, while intra-granular cracks may be only a few microns long. [Jaeger and Cook (1979)](jaeger_and_cook_1979) defined a crack as a confined region within a body of rock across which the displacements (shear or nor-
Fracture formation may be discontinuous under certain stress conditions. A macroscopic fracture will be made up of innumerable, much smaller asperities. This study primarily deals with field-based observations of fractures. Hence, macroscopic failure criteria will be of chief consideration, and provides information on factors influencing fracture formation (e.g. differential stress magnitudes, fluid pressure states).

Several experimental studies have demonstrated that three macroscopic fracture types form when brittle, intact, isotropic rock is loaded to failure in a ‘triaxial’ compression rig. These are known as extension fractures, extensional-shear (or ‘hybrid’ fractures) and shear fractures (Hancock, 1985). These fractures are symmetrically oriented in relation to the three effective principal stresses:

\[
\sigma'_1 = (\sigma_1 - P_f) \geq \sigma'_2 = (\sigma_2 - P_f) \geq \sigma'_3 = (\sigma_3 - P_f)
\]  

Fluid pressure \((P_f)\) reduces normal stress in a fluid-saturated rock under triaxial load (Hubbert and Rubey, 1959):

\[
\sigma'_n = (\sigma_n - \alpha P_f)
\]

where \(\alpha\) is a constant related to pore geometry (typically \(\sim 1\); Lockner, 1995). At depth \((z)\) in the crust, fluid pressure is related to the vertical stress \((\sigma_v)\) by the pore-fluid factor:

\[
\lambda_v = \frac{P_f}{\sigma_v} = \frac{P_f}{\rho g z}
\]

The type of fracture developed is controlled by the value of differential stress \((\sigma'_1 - \sigma'_3)\), and the tensile strength \((T)\) of the rock. Each class of fracture has its own range of dihedral angles \((\theta)\) about \(\sigma'_1\), where \(\theta\) is dependent on the failure envelope considered and the angle of internal friction \((\phi_i)\). The coefficient of internal friction \((\mu_i)\) is related to the angle of internal friction by:

\[
\mu_i = \tan\phi_i
\]

The failure modes for the three macroscopic fracture types may be constructed on a Mohr diagram, which relates effective normal stress \((\sigma'_n)\) and shear stress \((\tau)\) (Fig. 2.5). Here, failure envelopes are constructed with \(\mu_i = 0.75\). This value was chosen because it is around the average experimentally determined range for internal friction (Jaeger and Cook, 1979), and near the middle of the range for sliding friction determined by Byerlee (1978). When \(\sigma'_n > 0\) in intact, isotropic rock (e.g. compressional field), faults form in accordance with the linear coulomb criterion (Secor, 1965; Hancock, 1985):

\[
\tau = C + \mu_i \sigma'_n \approx 2T + \mu_i (\sigma_n - P_f), \quad \text{ when } (\sigma'_1 - \sigma'_3) > 5.66T
\]

where \(C \sim 2T\) and \(\mu_i\) is the coefficient of internal friction (commonly \(0.5 < \mu_i < 1.0\) Jaeger and Cook, 1979). It is important to note that \(P_f\) has no influence on the angle that faults initiate. Faults usually form along planes containing the \(\sigma_2\) direction, at an angle \(\theta\) to \(\sigma_1\) (typically \(25^\circ < \theta_i < 30^\circ\)):

\[
\theta_i = 0.5 \tan^{-1}(1/\mu_i)
\]
In the extensional field \((\sigma'_n < 0)\), the macroscopic Griffith failure criterion of parabolic form describes the stress conditions for extensional shear along planes oriented at \(\theta < \theta_i\) to \(\sigma_1\):

\[
\tau^2 = 4(\sigma_n - P_f)T + 4T^2 \quad \text{when} \quad 4T < (\sigma'_1 - \sigma'_3) < 5.66T \tag{2.7}
\]

For zero shear stress, this expression simplifies to (Sibson, 2001):

\[
P_f = \sigma_3 + T \quad \text{when} \quad (\sigma'_1 - \sigma'_3) < 4T \tag{2.8}
\]

describing stress conditions for the formation of pure extension fractures perpendicular to \(\sigma_3\). The macroscopic failure modes and failure criterion outlined above are illustrated in Figure 2.5.

![Figure 2.5: Generalised Mohr diagram of shear stress (\(\tau\)) plotted against effective normal stress (\(\sigma'_n\)). A composite Griffith-Coulomb failure envelope for intact rock and reshear criteria for a cohesionless fault are illustrated. Stress circles for the three different brittle failure modes are shown, along with their expected orientation to the principal stress axes (after Brace, 1960; Secor, 1965; Sibson, 2001).](image)

The composite failure envelope outlined above implies that different failure modes cannot occur simultaneously. However, varying levels of differential stress, stress axes orientations, fluid pressure and cohesion (via cementation) in the crust mean that different fracture types of various orientations may form at the same time.
2.3. Relationships between folding and associated structures

Reshear of a cohesionless fault

The failure criterion outlined in equation \(2.5\) is for intact, isotropic rock. Once a fault has formed, the rock is no longer intact, and the fault may act as a cohesionless surface. If this occurs, then equation \(2.5\) simplifies to:

\[
\tau = \mu_s \sigma'_{n} = \mu_s (\sigma_n - P_f)
\]

where \(\mu_s\) is the static coefficient of friction (Sibson, 1985). For fault planes containing \(\sigma_2\), equation \(2.9\) may be expressed in terms of the ratio of effective principal stresses:

\[
\frac{\sigma'_1}{\sigma'_3} = \frac{(\sigma_1 - P_f)}{(\sigma_3 - P_f)} = \frac{(1 + \mu_s \cot \theta_r)}{(1 - \mu_s \tan \theta_r)}
\]

This expression gives the differential stress level required to reactive a cohesionless fault at a given angle, \(\theta_r\), between \(\sigma_1\) and the fault plane. The optimum angle for reactivation \((\theta^*_r)\), where \(\sigma'_1/\sigma'_3\) reaches a positive minimum is:

\[
\theta^*_r = 0.5 \tan^{-1}\left(\frac{1}{\mu_s}\right)
\]

Byerlee (1978) found rock static friction coefficients to be in the range \(0.6 < \mu_s < 0.85\), and largely independent of rock type. For this range of friction coefficients, \(\theta^*_r\) occurs between 25° and 30°. Reactivation becomes progressively more difficult as \(\theta_r\) deviates from \(\theta^*_r\). As \(\theta_r\) approaches \(2\theta^*_r\), very large \(\sigma'_1\), or very low \(\sigma'_3\) is required to enable reactivation. Thus, for \(\theta > 2\theta^*_r\), \(\sigma'_3\) must be < 0 for cohesionless reactivation. Very low or negative \(\sigma'_3\) may be achieved by increasing fluid pressure. An important implication of cohesionless faults is the limit they impose on the level of differential stress and failure conditions that may be reached in the surrounding region of crust (Sibson, 1985).

Stylolites

While sensu stricto not a brittle fracture, stylolitic and planar pressure-solution seams are closely related both spatially and temporally to other brittle structures. Rispoli (1981) documented stylolites and extension fractures around strike-slip faults, and found mutually overprinting relationships between stylolites and veins, suggesting that vein and stylolite formation is a synchronous process. Some workers use the orientation of stylolites to directly infer a paleo \(\sigma_1\) orientation (Hancock, 1985, and references therein).

2.3 Relationships between folding and associated structures

A variety of small-scale structures can form in folds, including veins and fractures, cleavage, boudins, and lineations (including bedding-parallel slickenfibres). The type, orientation, and timing relationships of fractures provide a record of differential stress levels, and fluid pressure states during deformation. In this section, the various
fracture mechanisms summarized in §2.2.1 are used in conjunction with the fold mechanisms discussed in §2.1 to outline different fold-fracture relations.

The orientation and intensity of cleavage is related to the orientation and magnitude of the maximum shortening strain. Cleavage forms perpendicular to the maximum finite shortening direction. Its intensity is proportional to the aspect ratio of the finite strain ellipse. Consequently, the orientation and intensity of cleavage is related to the fold mechanism (and the material properties of the host rock).

Many cleavages occur subparallel to the axial surfaces of folds, and are called ‘axial planar cleavage’. Commonly, the cleavage orientation progressively changes from one side of the fold to the other. Foliations are generally strongly fanning across folds which form in rocks such as sandstones, which have a small proportion of platy minerals. The presence, style and orientation of cleavage in rocks may be used to infer fold histories and the relative competence of different rock types (Ramsay and Huber, 1987).

Folding and faulting are intimately related in many situations, with interdependent processes leading to at least three modes of fault-fold interaction: fault-bend folding, fault-propagation folding and detachment folding (Chester et al., 1991). A fault is defined as any macroscopic fracture along which shear offset has occurred. This includes bedding-parallel slip veins (discussed below). It has been suggested that fault zone drag and the folding resistance of layers influence fold geometry and the distribution of internal strain. The folding resistance of a layer is related to the rheological properties of each layer, and the overall stratigraphic unit competence. Analog modelling by Chester et al. (1991) demonstrated that interlayered, high-competence contrast sequences (analogous to an interlayered limestone-shale sequence) will shorten by fault-propagation folding, and favours the formation of rounded fold hinges, promoting buckling and extensive faulting on the fold limbs, rather than in the fold hinge. This prediction contrasts with the observations of Ramsay (1974), who found that competence contrast between beds leads to dilation and extensive fracturing in hinge zones as folds tighten. This difference may be due to the different degrees of strain between the models of Chester et al. (1991) and the field observations of Ramsay (1974).

Bedding-parallel fault veins (BPVs) are thin (usually millimetres to tens of centimetres wide), laminated veins, which may be continuous for hundreds of metres both down dip and along strike (Tanner, 1989; Fowler, 1996). Various pre-folding, early-folding and syn-folding mechanisms have been proposed for the origin of BPVs. Mineral slickenfibres (usually quartz or calcite) are commonly found within the plane of the BPV, and are usually subperpendicular to the hinges of related folds. This has led various workers to suggest that BPVs form during flexural-slip folding (Ramsay, 1974; Tanner, 1989; Fowler, 1996). If slip continues between cohesionless beds to dip-angles that exceed $\theta^*$, and approach $2\theta^*$ (i.e. severely misoriented fold limbs), then frictional lockup of fold limbs may occur (Ramsay, 1974). To continue flexural-slip folding at these angles require fluid pressures that exceed $\sigma_3$ (see §2.2.1 Sibson, 1985).

Ramsay (1974) presented the chevron folding model, invoking slip between layers, and dilation at hinge zones during folding. Ramsay demonstrated that hinge dilation and accompanying saddle reef formation was most prevalent during the latest
stages of folding, Ramsay (1974) suggested that chevron fold development, and accompanying bedding-parallel slip is halted once interlimb angles reach $\sim 60^\circ$, and frictional lockup occurs.

Extension veins (or hydraulic fractures) provide constraints on the stress and fluid pressure conditions present during deformation. Failure is governed by equation 2.8, with high fluid pressure levels and relatively low levels of differential stress required. For the formation of subhorizontal extension fractures, such as those formed in a compressional tectonic regime (i.e. $\sigma_3 =$ vertical), then $\lambda_v$ must exceed 1.0 (i.e. greater than lithostatic fluid pressures). During the folding of a rock sequence by pure flexural slip, all the strain is accommodated by sliding between bedding planes, and no extension fracturing occurs. For a flexural flow mechanism, where shear strain is more evenly distributed, extension fractures may develop. The orientation and distribution of these fractures will be dependent on the strain distribution, and timing of fracturing during fold growth (Fig. 2.6).

During simple flexural flow folding, fractures will initiate at about $45^\circ$ to $135^\circ$ to layering surfaces. As fold growth continues, veins will be rotated to higher angles, and new veins may crosscut earlier veins. Veins may widen via internal deformation. Veins may continue propagating during rotation, and acquire a sigmoidal form. Extension veins are commonly found in en echelon arrays, and are commonly interpreted as forming in a simple shear array, with shearing parallel to the vein array (Fig. 2.6). New veins may cut existing veins, and coalesce to form a through going vein.

Later stages of chevron fold growth are characterised by a slowing in shortening rate and fold growth. This stage may be characterised either by fold lockup (see above), and/or modification of fold geometry by hinge thickening and limb thinning. Stretching of fold limbs may result in boudinage, and/or extension fracture formation orthogonal to bedding (Fig. 2.6d).

2.4 Fluid flow and hydrothermal veins

Fluid flow in rocks is driven by variations in hydraulic head, which may be affected by topography, poroelastic effects, or mineral dewatering reactions. In exhumed rocks, hydrothermal veins provide a record of mass transfer which occurred at depth. Mass transfer may occur via diffusion or advection. Understanding the underlying mechanical controls on fracture formation (§2.2), combined with observations of the geometric and temporal relationships between other rock structures, allows inferences to be made about changes in stress and fluid pressure conditions in the Taemas Vein Swarm.

In this section, controls on fluid flow and crustal permeability are outlined. Links between permeability, seismicity and fluid flow are discussed, with particular attention paid to the redistribution of fluids and variation in fluid pressure during the seismic cycle.
2.4.1 Fracturing, permeability and fluid pathways

Fluid flow in permeable media may be described by Darcy’s Law. For one-dimensional, horizontal flow, the fluid flux \( Q \) (fluid volume) across a cross-sectional area \( A \), per unit time \( t \) is:

\[
\frac{Q}{At} = \frac{k}{u} \frac{dP}{dx}
\]  

(2.12)

where \( k \) is the rock permeability and \( u \) is the viscosity of the fluid. The driving force for the flow is dependent on the hydraulic pressure gradient (change in pressure, \( dP \)), over the change in distance, \( dx \). Permeability (units m\(^2\)) is an intrinsic rock property, and is a measure of the ability of a rock to ‘pass fluid’. Many fluid flow regimes have a component of vertical fluid flow. The change in fluid pressure \( dP_f \) for a change in depth \( dz \) for a column of stationary fluid is given by:

\[
dP_f = \rho g dz
\]  

(2.13)

where \( \rho \) is the local fluid density. Thus, for vertical flow, Darcy’s law has the form:

\[
\frac{Q}{At} = \frac{k}{u} \left( \frac{dP}{dz} - \rho g \right)
\]  

(2.14)

For fluids with nonuniform density, the \( \rho g \) term in equation 2.14 describes the force for buoyancy-driven fluid flow. Fluid density is strongly dependent on salinity,
2.4. Fluid flow and hydrothermal veins

pressure and temperature. Typically, the increase in temperature with depth causes fluids to become less dense with increasing depth. This generates buoyancy-driven flow in permeable, fluid-saturated rocks (Cox, 2005).

Darcy’s Law illustrates that fluid flow in the crust is controlled by variations in rock permeability, fluid pressure gradients, fluid viscosity and fluid density. As natural rock permeabilities vary by over ten orders of magnitude, the permeability structure of a rock mass will control the flow of fluids in most crustal-scale hydrothermal systems (Ingebritsen and Manning, 1999; Cox, 2005).

2.4.2 Fluid sources

Overpressured fluids may be derived from a variety of sources. These include the mantle (e.g. devolatilisation of subducting slabs), dewatering mineral reactions during prograde metamorphism, magmatic fluid, surface-derived fluid (oceanic or meteoric) and pore fluids in sedimentary basins. In collisional orogens, devolatilisation reaction during prograde metamorphism may be a major fluid source, typically producing low-salinity H$_2$O-CO$_2$ fluids (Powell et al., 1991). Various estimates have been made of fluid production rates and fluid fluxes during prograde metamorphism (Walther and Orville, 1982; Yardley, 1986).

Several studies have demonstrated that surface fluids (meteoric fluids or fluids trapped in sedimentary basins) may penetrate to significant crustal depths (Holm et al., 1989; Forster and Smith, 1990; Taylor, 1990; Upton et al., 1995). Several mechanisms for deep fluid penetration have been suggested, including topographic flow (Holm et al., 1989; Forster and Smith, 1990; Upton et al., 1995) and earthquake rupturing (McCaig, 1988). The time-integrated volume of fluids transported by earthquake rupturing could be substantial. A simple calculation, assuming a fault rupture 50 km long, 20 km deep, with a fault damage zone 50 m wide containing 2% porosity could store 1 cubic kilometre of fluid. One hundred rupture cycles represents a total fluid volume of 100 km$^3$, which, assuming a modest recurrence interval of 100 to 1000 years, could be expelled from a fault zone over 10,000 to 100,000 years (Cox, 2005).

Hydraulic gradients

At shallow crustal depths, topography causes marked lateral and vertical variations in hydraulic gradients. In the Southern Alps of New Zealand, fluids are driven to depths of more than 6 km by topographically-driven flow (Holm et al., 1989; Upton et al., 1995). Progressive changes in the pore volume of deforming rocks has been suggested as a driving force for crustal-scale fluid flow events, especially in compacting sedimentary basins (Oliver, 1986). Poroelastic effects, that is changes in pore volumes associated with the elastic deformation of intergranular pores and cracks, will generate temporary hydraulic gradients. Such changes have been suggested to cause significant fluid flow around active fault zones, where substantial stress changes are believed to occur during the seismic cycle (Muir-Wood and King, 1993; Sibson, 1993). However, Cox (2005) suggests that poroelastic-driven fluid flow will be prevalent only in areas which have a predominance of high-aspect ratio fractures.
2. Folding, fracturing and fluid flow

(e.g. the shallow crust), when large normal stress changes occur.

Deeper in the crust, hydraulic gradients are more strongly affected by confining pressure, inelastic deformation processes, and chemical reactions which close pores and fractures. Porosity reduction drives pore fluid pressures towards lithostatic values, and sometimes supralithostatic values (Etheridge et al. 1984, Sibson et al. 1988). Fluid overpressures occur where the fluid pressure is greater than hydrostatic, and are significant driving forces for fluid migration. Overpressuring is dependent on the rates of reduction in porosity and permeability, and the generation of fluid via dewatering reactions. Fluid overpressuring is commonly observed in deep sedimentary basins (Hunt, 1990). Here, fluid is generated by low-temperature dewatering reactions and compaction, while vertical fluid escape is inhibited by low permeability ‘seals’ (e.g. low-permeability rock units, hydrothermal mineral horizons). Additionally, low permeability sidewalls may form (e.g. sealed fault zones, Knipe et al. 1998), creating highly overpressured fluid ‘compartments’. Changes in rock permeability (e.g. due to fault rupture and associated fracturing) provide ‘short circuit’ pathways between different fluid pressure compartments, and generate significant fluid flow (Cox 2005).

2.4.3 Fracture permeability

Grain-scale fracture permeability

If intergranular porosity is very low and poorly interconnected, then permeability will be controlled by competition between fracture growth (porosity creation) and porosity destruction (compaction, crack-healing and crack sealing). In the presence of non-reactive pore fluids, permeability during straining of initially low permeability rocks evolves as follows (Zhang et al. 1994; Cox 2005):

1. During the initial stage of deformation microcracks start to nucleate and grow. However, as cracks are isolated, the permeability is invariant.

2. With increasing strain the number of microcracks increases, and their lengths progressively grow until the percolation threshold is reached (the point at which enough elements connect to allow fluid flow across the entire width of the fracture system).

3. With progressive strain, rapid growth of grain-scale crack networks leads to rapid growth in connectivity, and a subsequent increase in permeability. Near total interconnectivity is achieved in the grain-scale network by 5 % strain, although this value is highly dependent on the confining pressure.

4. Further gains in permeability beyond 5 % strain have been attributed to increases in fracture aperture (rather than greater connectivity or fracture density; Zhang et al. 1994).

Constant strain, pore fluid pressure stepping experiments, indicate that increases in pore fluid pressure during deformation can rapidly enhance permeability (Fischer and Patterson 1992). Experimental work demonstrates that microfracture
networks develop high crack connectivity and high permeability at very low strains, provided pore fluid factors are high (see summary in Cox, 2005). Hence, low strain deformation can have a major impact on crustal permeability. Field evidence for grain-scale dilatancy and related permeability enhancement has been reported, with grain-scale and intergranular microfractures filled by mineral precipitates (e.g. Cox and Etheridge, 1989).

High-permeability microcrack networks are likely sealed on geologically short timescales. Crack-seal textured veins are inferred to be formed by hundreds to thousands of individual fracturing events. If a vein formed over a period of c. 1,000,000 years, then fracturing and sealing probably occurs on timescales of 1,000 years or less (Ramsay, 1980; Cox and Etheridge, 1983). Numerical modelling by Lowell et al. (1992) suggested that a 1 mm wide fracture could be sealed by quartz precipitation over a period of 20–40 years. This rapid permeability destruction means that continuing deformation and fracturing is required to sustain high permeability pathways.

**Macroscopic fracture-controlled permeability**

For laminar flow in a parallel-sided fracture, the effective permeability \( k \) is:

\[
k = \frac{a^2}{12}
\]

where \( a \) is the fracture aperture (Cox, 2005). Most fractures have variable apertures, and flow rates become complicated as flow becomes tortuous (Taylor et al., 1999). The importance of aperture is highlighted by equation 2.15. For example, if the...
aperture doubles, the flow rate increases four fold. Hence, dilatant sites on fracture surfaces (e.g. dilational jogs or other irregularities) with greater aperture will have substantially higher permeability, and flow will become localised in these sites.

For networks of randomly distributed fractures, which are not fully connected, permeability may be described by:

\[
k = \frac{\pi}{120} \frac{f a^3 r^2}{l^3}
\]  

(2.16)

where \( a \) is the mean fracture aperture, \( r \) is the mean fracture length, \( l \) is the mean fracture spacing and connectivity, \( f \), is \( 0 \leq f \leq 1 \) (Gueguen and Dienes, 1989). This simple relationship demonstrates the influence that fracture development will have on flow in hydrothermal systems. Highest fluid fluxes (and localised flow) occur where fracture apertures, fracture density and fracture connectivity are at their highest. An example of this would be a dilational jog on a fault (Sibson, 1987).

In cracked materials, permeability is extremely sensitive to changes in stress state and fluid pressure. For a fluid-filled fracture, permeability varies according to:

\[
k = k_0 \exp \left( \frac{-\sigma'_n}{\sigma_0} \right)
\]  

(2.17)

where \( k_0 \) is permeability at zero effective stress, and \( \sigma_0 \) is a constant (Rice, 1992). Hence, either decreasing normal stress or increasing fluid pressure may enhance permeability.

The overall permeability of the fracture network will be controlled by the competing processes of fracture opening (permeability creation) and crack sealing (permeability destruction). At elevated temperatures with varying fluid pressures, mineral deposition may occur rapidly, blocking dilatant fractures. Thus, in hydrothermal environments, with competing fracture formation and destruction, permeability is a dynamically evolving processes (Cox, 1999; Sibson, 2001; Cox, 2005).

### 2.4.4 Percolation networks

Fluid migration in the crust involves flow through pervasive, grain-scale fractures, and flow through macroscopic fracture systems (see §2.4.3). Both of these processes are critical to the formation of a hydrothermal vein system. In the fluid source region, pervasive flow at grain scales leads to intensive fluid-rock interaction, creating a fluid rich in dissolved constituents. This fluid reservoir must then be connected to a macroscopic fracture network (probably by coalescence and interconnection of microscopic fractures), allowing transport of fluid and chemical constituents over large distances (Cox, 1999, 2005). Varying temperature, pressure and chemical conditions cause supersaturation of some chemical species, resulting in mineral precipitation and the formation of hydrothermal veins (Sibson, 1987, Parry, 1998, Cox, 1999, 2005). Fluid focussing occurs around the higher pressure (upstream) levels of fault zones, while fluid discharge occurs at the lower pressure (downstream) levels of fault zones (Cox et al., 2001).

Macroscopic percolation networks (comprising faults and fractures) consist of (Cox, 1999, Cox et al., 2001):
1. **Backbone elements** provide a direct connection from one side of the flow system to the other, and carry the majority of the fluid flux.

2. **Dangling elements** branch from the flow backbone and act as fluid feeders to the backbone in the upstream part of the system, and as discharge structures in the downstream part of the system.

3. **Isolated elements** are disconnected from both the backbone and dangling elements in the network. These are low flux structures, not connected to fluid reservoirs.

Sahimi (1994) recognises two end members for the growth of percolation networks; ordinary percolation and invasion percolation.

**Ordinary percolation**

‘Ordinary’ percolation occurs when fracture growth is controlled largely by shear stress (Cox 2005). In this case, fractures nucleate and grow in almost random locations throughout a deforming, uniform rock mass. During progressive deformation, active faults and shears grow in length, and new structures nucleate and grow. Thus, fracture connectivity progressively increases as strain intensifies. At the percolation threshold enough elements connect to allow fluid flow across the entire width of the fracture system, and flow begins across the system (Fig. 2.8; Cox, 2005). The percolation threshold can be reached at bulk strains of a few percent (Cox et al., 2001).

Flow is partitioned depending on the proportions of backbone, dangling and isolated elements. Just above the percolation threshold, the backbone is a very small fraction of the total fracture network. Most of the flow is localised along this part of the system. At higher strains, the percentage of the fracture network which is part of the backbone progressively increases. Thus, fluid flow becomes more evenly distributed across the system (Cox et al. 2001; Cox, 2005).

**Invasion percolation**

Ordinary percolation assumes that all elements of the percolation network grow at the same rate. For fracture networks connected to an overpressured fluid reservoir, the invasion of high pressure fluids along fluid-accessing faults and fractures may greatly increase their growth rates (relative to elements isolated from fluid reservoirs; Cox, 2005). Here, the development of the percolation network will occur in those parts of the network which connect to the high pore fluid factor reservoir(s) (Fig. 2.8). These percolation networks will be dominated by backbone and dangling elements (Cox et al. 2001; Cox, 2005).

This ‘self-generation’ of percolation networks in response to invasion of overpressured fluid provides a positive feedback between fluid access and fracture growth rate. This will tend to localise slip on the flow backbone which formed at the percolation threshold. Hence, that part of the fracture system will repeatedly access the same fluid reservoir (Cox et al. 2001).
2.5 Faults, fluids and earthquakes

In the middle to upper crust, fluid migration through rocks with low permeability likely occurs via networks of macroscopic fractures. Fractures and microscopic grain-scale porosity will only remain permeable as long as they are not sealed by mineral precipitation. In the upper continental crust, fault slip is accommodated mainly by earthquake rupturing (Sibson, 1983). Studies of exhumed fault zones provide plentiful evidence for their role as fluid conduits (see reviews of Parry, 1998; Cox et al., 2001; Sibson, 2001; Cox, 2005).

Various workers have shown that many fault-fracture vein networks are formed by incremental processes, with each episode of fracture growth followed by fracture-filling via mineral precipitation (‘fracture healing’). Seismic rupture has been cited

![Figure 2.8: Progressive evolution of connectivity in a region of uniform deforming crust. ‘Ordinary’ percolation is a model where fault growth is stress driven. Faults (dotted lines) randomly nucleate throughout the rock mass, increasing in surface area and connectivity in time. For ‘invasion’ percolation, fault nucleation and growth is fluid-driven. Here, faults which connect to the fluid reservoir (hatched area) nucleate and grow until a backbone network (solid line) forms. The backbone pathway is represented by the bold lines. The percolation threshold is reached at time 3. After Cox (2005).]
as a key generator of permeability (notable contributions include Sibson et al. 1975, 1988; Boullier and Robert 1992; Cox 1995; Sibson 2001). Fluid flow will be episodic in faults where seismic slip generates fracture permeability, and interseismic sealing progressively reduces permeability. Fluid flow will be dominated by immediate post-seismic transient flow, rather than by steady-state flow (Braun et al. 2003).

Direct evidence of fluid involvement in shallow crustal earthquakes comes from observations of post-seismic fluid discharge in the vicinity of active faults (Muir-Wood and King 1993). Remote seismic observations of some deeper earthquakes show that \( V_p/V_s \) ratios change around fault zones following an earthquake (cited as evidence for the presence of fluid in or around a fault zone; Husen and Kissling 2001). Migrating aftershock behaviour has also been attributed to fluid pressure migration (Miller et al. 2004).

Earthquake rupture occurs to relieve accumulated shear stress \( (\tau) \) on faults. Stick-slip models suggest that shear stress on a seismogenic fault follows an approximately sawtooth pattern (Scholz 1990). Periods of shear stress accumulation are followed by sudden release of this stress during earthquake rupture. One key point of this observation is that the stress state around a fault after an earthquake may be markedly different from the prefailure state (Fig. 2.9). Sibson (2001) suggests that the aftershock phase of an earthquake, which is accompanied by both abrupt localised stress changes, and the generation of fracture permeability, is the time of maximum fluid redistribution around a seismically active structure.

### 2.5.1 Stress regimes and fluid flow

Three basic stress regimes were recognised within the crust by Anderson (1951). Anderson assumed that the Earth’s surface is approximately horizontal, and is unable to support shear stress. Hence, the Earth’s surface is a principal stress plane, and the principal stress axes must be parallel and perpendicular to this surface.

If brittle faults form through Coulomb failure in homogeneous intact rock (see §2.2.1), then three fundamental modes of faulting exist in the Earth; steeply dipping normal faults (inclined at c. 60° to the Earth’s surface when \( \sigma_v = \sigma_1 \)), gently dipping thrust faults (inclined at c. 30° to the Earth’s surface when \( \sigma_v = \sigma_3 \)) and subvertical strike-slip faults (when \( \sigma_v = \sigma_2 \)). It must be noted that Anderson’s fault classification was based on fault orientation with respect to the stress field at the time of fault inception. Some regional studies have demonstrated that large, rather uniform stress provinces extend over large areas of the Earth’s crust (Zoback et al. 2003).

As discussed above, fluid pressures in the crust may reach supralithostatic levels \( (e.g. \lambda_v > 1.0) \). It has been postulated that fluid overpressures are easier to maintain in compressive tectonic regimes, due to the predominance of gently dipping to flatlying fault and fractures (Sibson and Scott 1998; Sibson 2001). In low-permeability rocks, the degree of fluid overpressuring is limited by the formation or reactivation of fractures and faults (which act as fluid conduits, draining high pressure fluids).
2.5.2 Fluid redistribution and the seismic cycle

The complexity and coupling between between tectonic stress, fault-fracture permeability and fluid pressure lead to a range of potential links between the earthquake stress cycle and fluid redistribution. More than one mechanism may operate at once, and the relative importance and occurrence of these mechanisms is poorly constrained.

Cyclical dilatancy pumping

Various types of grain-scale microcracking and fracture dilatancy during stress changes in the earthquake cycle have been proposed (see review of Sibson, 1994). Suggestions of regional microcrack dilatancy operating at high levels of shear stress (\( > 100 \) MPa) in the crust adjacent to large active fault zones (Scholz \textit{et al.}, 1973; Sibson \textit{et al.}, 1975) have not been substantiated (Sibson, 2001). This suggests that dilatancy pumping is not an important mechanism for fluid redistribution during the seismic cycle.

The ‘suction pump’

Cycling tectonic stress not only increases shear stress on faults, but also changes the normal stress acting on the fault. This alters both the fault’s frictional strength and the level of mean stress. For dip-slip faults, changes in \( \Delta \sigma \) are roughly comparable to the shear stress drop (typically \( 1 < \Delta \tau < 10 \) MPa, equivalent to changes in hydraulic head of 0.1 to 1 km; Kanamori and Anderson, 1975; Sibson, 2001). However, this coupling is opposed for reverse and normal faults. Assuming constant fluid pressure, normal faults are load-weakening with average mean stress decreasing during loading, and abruptly increasing post-failure. Reverse faults are load-strengthening, with average mean stress increasing during loading. These opposite phenomena will cause significant differences in the way fluid behaves around reverse and normal fault zones (see Sibson, 2001, for a review). However, Cox (2005) suggests that these effects will be only significant around normal faults with high fracture density immediately adjacent to the fault.

Post-seismic fluid redistribution around fault irregularities

Rupturing and slip transfer across fault irregularities generate significant postfailure \( \Delta \sigma \) around the irregularities (Segall and Pollard, 1980). Fluids will be redistributed from areas of high to low mean stress. Sibson (1986) demonstrated that aftershocks tended to cluster in areas of inferred mean stress reduction, and suggested that this was due to fluids being redistributed into dilational jogs. More recently, Coulomb stress transfer modeling (King \textit{et al.}, 1994) has been used to predict regions of mineralisation in exhumed fault-fracture hosted gold deposits (Cox and Ruming, 2004; Micklethwaite and Cox, 2004, 2006).
Fault-valve behaviour

Fault-valve action involves the postfailure discharge of overpressured fluids through fault-fracture systems. This may occur when fault rupture breaches low-permeability barriers which bound overpressured portions of the crust. Extreme valving behaviour is likely when fluid pressure compartments with very high fluid pressures (e.g. $\lambda_v > 1.0$) are separated from shallower, near-hydrostatically fluid pressured compartments. Situations for fault valving include overpressured sedimentary basins, areas of active magmatic intrusion and the brittle carapace of prograding metamorphic terranes (Sibson et al. 1988; Boulier and Robert 1992; Cox 1995; Sibson and Scott 1998; Fournier 1999).
The conditions favourable for ‘extreme fault-valve action’ were outlined by Sibson (2001). Sibson suggests that for significant mineralisation to develop, each fault-valve cycle should involve the rapid discharge of large fluid volumes. Rapid fluid discharges would be favoured by the creation of high-permeability flow pathways, high \( dP/dz \) (i.e. connection of a lithostatic to a hydrostatic pressured compartment) and a large fluid reservoir. This may be achieved in regions of crust that lack low-cohesion faults favourably oriented for reactivation, which will enable higher fluid pressures to develop before fault rupture occurs (Sibson, 2001). A major feature of fault-valve behaviour is that fault rupture generates permeability, and connects high and low pore fluid factor regions of crust. Following rapid fluid discharge, gradual hydrothermal mineral sealing of the fault occurs. This allows fluid pressure to build once again beneath the low-permeability domain. Thus, the fault-valve cycle is repeated, generating episodic fluid flow through the fault-controlled hydrothermal system (see Sibson, 2001; Cox, 2005 for reviews).