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Alpine Fault Pseudotachylytes

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Abstract

Pseudotachylytes produced during coseismic slip are ubiquitous within the rocks of the Alpine Fault zone, New Zealand. Pseudotachylytes from three principal host rocks are described and characterised from field and petrographic observations. The pseudotachylytes range from ~1 mm thick in Alpine Schist-derived mylonites to ~1 cm thick in Western Province-derived augen mylonites and cataclasites. Depth estimates based on field and petrographic relationships divide the differently hosted pseudotachylytes into two broad groups that formed during seismic rupture in the upper crust of the Alpine Fault zone: 1) relatively low volume pseudotachylytes formed in Alpine Schist-derived mylonites at ~8 km depth, where shear stress was relatively low, during low magnitude events just below the brittle-ductile transition; and 2) voluminous pseudotachylytes formed at shallower levels of ~2-5 km, where shear stresses were high, in Western Province granitoid-derived augen mylonites and cataclasites during higher magnitude events.

Bulk-rock powder X-ray Fluorescence (XRF) of host rocks and Electron Microprobe Energy Dispersive Spectroscopy (EMP-EDS) spot analysis on pseudotachylyte matrices reveal that relative to their host rocks, the Alpine Fault pseudotachylytes are depleted in SiO$_2$ and enriched in Al$_2$O$_3$, alkalis and metallic oxides. This is interpreted to be the result of preferential inclusion of hydrous, ferromagnesian mineral phases and to a lesser extent feldspar into the melt, and the exclusion of quartz from the melt. The preferential selection relationship in part controls the voluminous nature of pseudotachylytes within retrogressed host rocks containing high proportions of hydrous, ferromagnesian minerals.

Size analysis of the clasts within the pseudotachylytes reveals power law size-frequency distributions that are modified in the finer fraction of the clasts, suggesting pervasion of friction melt depleted the fine fraction of precursor ultracataclasites by incorporation into the melt phase by uniform rim melting.
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Chapter 1

Pseudotachylytes as Seismological Indicators

1.1 Introduction

Before embarking on a study of fault-related pseudotachylyte, it is necessary to discuss pseudotachylyte field relationships and terminology by which to describe them. The seismological background to understand pseudotachylyte generation is of equal importance. This chapter is designed to address these key aspects by way of reviewing the literature on pseudotachylytes, as well as considering prominent issues surrounding pseudotachylyte origin that have arisen from previous research. The chapter concludes with an outline of the scope and aims of my research into pseudotachylytes within the Alpine Fault Zone, New Zealand.

1.2 Occurrence of Pseudotachylyte in the Field and in the Crust

Pseudotachylyte is a dark grey-black aphanitic material, typically occurring as planar fault-veins, with irregular, sporadic structures injecting off the planar surface into the surrounding host rock. The veins can be individual in nature, or form vein networks, and should preserve evidence of having been through a melt phase if they are true pseudotachylyte. A number of authors have reported this pseudotachylytes within ancient or active crustal fault zones that have experienced frictional generation of high temperatures during seismic slip rates (~ 1 ms⁻¹), subsequently leading to the formation and preservation of solidified frictional melts: Sibson (1975), from the Outer Hebrides Thrust in NW Scotland; Swanson (1988, 1992), from the Fort Foster Brittle Zone in Maine; Lin (1991, 1994), from the Fuyun Fault Zone in NW China; Di Toro & Pennacchioni (2004, 2005), from the Gole Larghe Fault Zone in the Italian Alps; Toy (2007), from the currently active Alpine Fault in New Zealand.

Pseudotachylyte occurrence is not, however, restricted to fault zones, and has been described in meteorite impact zones in vast volumes (Shand, 1916; Thompson & Spray, 1996; Spray, 1997), and also from basal shear surfaces in large landslides (Masch et al., 1985; Grunewald et al., 2000; Legros et al., 2000). Fault generated pseudotachylyte is the focus of this study, and the following paragraphs address pseudotachylyte of a seismic origin.

Pseudotachylyte veins can be hosted in, or associated with a wide variety of fault rocks; and as Magloughlin and Spray (1992) highlight, pseudotachylytes generally display features and structural associations suggest formation at shallow to mid-crustal levels (~ 2-20 km). There are pseudotachylytes that are clearly closely with brittle deformation, where pseudotachylytes and cataclasites occur in intimately associated zones, that either formed at the same time (co-seismically) or at different periods in the seismic cycle (Sibson, 1975; Macaudière & Brown, 1982, 1998; Macaudière et al., 1985; Swanson, 1988, 1989, 1992; Magloughlin, 1989, 1992, 2005; Toyoshima, 1990; O’Hara, 1992; Cooper & Norris, 1994; Ray, 1999, 2004; Ikesawa et al., 2003; Cosca et al., 2005; Di Toro & Pennacchioni, 2004, 2005; Rowe et al., 2005). These fault surfaces that generate pseudotachylyte are preserved as brittle shear fractures overprinting the existing rock fabric, and the pseudotachylytes within them can be over-printed by subsequent brittle deformation. Most
of the recognised pseudotachylyte worldwide appears to have been generated in the upper crustal seismogenic zone (e.g. Sibson & Toy, 2006). However, pseudotachylytes are also known be associated with ductile deformation and often themselves sheared into a mylonitic foliation (Sibson, 1980; Passchier, 1982; Swanson, 1988, 2005; Koch & Masch, 1992; White, 1996, 1998; Pennacchioni & Cesare, 1997; O’Hara, 2001; O’Hara & Sharp, 2001; Toyoshima et al., 2004; Lin et al., 2005). Numerous examples of pseudotachylytes that have formed in the continental crust under HP/HT granulite facies conditions (Clarke & Norman, 1993), and under HP/LT blueschist and eclogite facies conditions (Austrheim & Boundy, 1994; Lund & Austrheim, 2003; Austrheim & Anderson, 2004; Anderson & Austrheim, 2006) are reported in the literature. These assemblages coincide with great depths in the crust ranging from ~25 km to ~80 km (Sibson & Toy, 2006).

One school of thought infers that the deep crustal pseudotachylytes formed in-situ, during transient ductile seismic slip (Hobbs et al., 1986; Koch & Masch, 1992; White, 1996), while the other hypothesis is that they represent the downward propagation of ruptures into ductile regime of the mid-lower crust (Sibson, 1977; Strehlau, 1986, Lin et al., 2005). The origin of the pseudotachylytes from the HP/LT assemblages is inferred to be due to local hydration, embrittlement and subsequent seismic failure (Anderson & Austrheim, 2006). Pseudotachylytes from the Alpine Fault, of particular interest to this study, are mostly associated with brittle structures and are therefore thought to have mostly formed in the upper crust at depths of ~2-10 km (Sibson & Toy, 2006).

As mentioned above, pseudotachylytes occur as planar veins with irregular injection structures emanating from the plane into the host rock. Vein sets can occur as simple individual veins and injection structures, or as complicated networks such as the en echelon duplexes and complex brittle zones described by Swanson (1992). Pseudotachylyte vein sets occur at a range of scales from several meters to several kilometers along strike and/or through thickness (Lin, 2008). Sibson (1975) provided a detailed description of pseudotachylyte vein geometry in foliated rocks of the Outer Hebrides Thrust, where he used the term ‘fault vein’ to describe the planar veins, and ‘injection vein’ to describe the irregular, erratic intrusions away from the fault vein into dilatant fractures and structures in the host rock. The fault veins or ‘generation zones’ can be further subdivided into discordant or concordant veins that cut across metamorphic anisotropies or lie subparallel to the anisotropies, respectively (Sibson, 1975; Grocott, 1981).

It is important to distinguish between the different veins, as this enables understanding of the kinematic evolution of the fault zone (Magloughlin & Spray, 1992). Fault veins or generation zones are generally planar, contain relatively abundant lithic clasts, vary in thickness, and although most of them can rarely be traced for distances greater than ~10 m (Sibson & Toy, 2006), they typically have high length-thickness ratios. The networks the fault veins formed can sometimes be systematic, and can be interpreted to be Riedel shear arrangements (e.g. Swanson, 1988, 1989, 1992; Bossière, 1991). Injection veins or reservoir zones often have irregular boundaries, contain relatively few lithic clasts, are shorter in length, and tend to terminate by tapering out into the host rock. Reservoir zones can concentrate large volumes of pseudotachylyte, for example in dilational jogs, blebs, pinch and swell lenses (Sibson, 1975), and splay fault junctions at the trailing ramp of sidewall ripouts (Swanson, 1992). These zones can contain large rotated blocks of host rock and breccias, and with continued injection of pseudotachylyte from the generation surfaces and subsequent melting and rounding of the lithic clasts, quasi-conglomerates can form (Sibson, 1975).
1.3 Terminology and the Physical Origin of Pseudotachylyte

Shand (1916) first described dark, aphanitic, glassy material that occurs as networks and veins in the Vredefort meteorite impact structure in the Parys region of South Africa. Shand used the term ‘pseudotachylyte’ to describe the material due to its close resemblance to, but with a physical and chemical origin distinct from, the glassy basaltic rock known as tachylyte (Magloughlin & Spray, 1992; Lin, 1998). Various authors have described this same material, and many of them have adopted different names for the material: ‘flinty crush-rocks’ (Clough, 1888), ‘trap-shotten gneiss’ (Holland, 1900), ‘injection mylonite’ or ‘pseudotachylite’ (Philpotts, 1964), and ‘hyalomylonite’ (Wallace, 1976; Masch et al., 1985). These terms are related to the respective authors’ interpretations on the physical origin of pseudotachylyte, which has led to considerable confusion regarding nomenclature, therefore all of these names, except ‘pseudotachylite’, have consequently been abandoned. In this thesis the term ‘pseudotachylyte’ will be used to describe the fault-related veins.

Some authors (e.g. Magloughlin & Spray, 1992) reserve the term ‘pseudotachylyte’ for veins that have originated from frictional melting processes. It is also known that a continuum exists between melt-origin pseudotachylyte and ultracataclasite veins and therefore they are often intimately associated in fault zones and sometimes the same vein. Finely comminuted ultracataclasite veins superficially resemble melt-origin pseudotachylyte and have been termed crushing-origin pseudotachylyte by some authors (Wenk, 1978; Lin, 1989, 1996, 1997, 2001, 2008; Wenk et al., 2000). This crushing-originated pseudotachylyte or ultra-cataclasite is composed of very fine-grained fragments and forms veins and injections structure that closely resemble melt-origin pseudotachylyte. This is due to this fine-grained crush material becoming dynamically fluidised (fluidised fault gouge), and obtaining extreme mobility as a gas-solid-fluid system during seismic faulting (Lin, 1996, 1997). Classification of the physical origin of pseudotachylyte into a melt- or crushing-origin is difficult due to its extremely fine-grained nature, and the common obscuring effects of recrystallisation, deformation, alteration and metamorphism – these being the major reasons for such controversy over the origin of pseudotachylyte (Lin, 2007).

The current, general consensus regarding pseudotachylyte origin is that it is a product of frictional melting on a fault plane, produced by movement at seismic slip rates. This consensus was arrived at on the basis of being supported by a considerable amount of literature that clearly demonstrates pseudotachylytes have been through a melt phase (Maddock, 1983; Spray, 1987, 1995; Lin, 1991, 1994a, b; Lin et al., 2002). It has taken about a century to reach this consensus, with the major debate being whether pseudotachylyte is the product of intense crushing and cataclasis, or is the product of melting. Holland (1900) was the first worker to suggest pseudotachylyte (or ‘trap-shotten gneiss’ as he referred to it) was a product of melting of the rock by mechanically generated heat. Shand (1916) interpreted the ‘type’ pseudotachylytes he described as being products of melting, and due to the lack of evidence of a fault relation, he concluded they were shock related. Many authors up until the late 1970’s interpreted pseudotachylyte to have a melt-origin based on glassy textures, amygdules, vesicles (Scott & Drever, 1953); and spherulites, radial microlites, devitrification textures, irregular and embayed lithic clasts, and recrystallisation (Park, 1961; Philpotts, 1964). Following important, highly acclaimed publications that concluded pseudotachylyte is, or can be produced by frictional melting on a fault plane, based on both field and petrological studies (Sibson, 1975, 1977), and theoretical investigations (Jeffreys, 1942; Mckenzie & Brune, 1972; Cardwell et al., 1978), the frictional melting hypothesis was widely accepted by researchers in the field during the 1970’s (Lin, 2007). Wenk (1978) and Wenk & Weiss (1982),
suggested a crushing-origin for pseudotachylyte, and that the rarity of glass in a pseudotachylyte indicated the need for this alternative generation mechanism. They contended that previous studies had failed to confirm the presence of glass in any pseudotachylytes and argued for an origin by ultra-comminution, and they pointed out that veins contain many features of intense cataclasis, indicative of formation by what was referred to as an ‘explosive spallation’ mechanism.

This pseudotachylyte origin controversy was mostly resolved by the early 1990’s (Lin, 2007), as many studies demonstrated evidence of pseudotachylyte veins having been through a melt phase (Maddock, 1983; Spray, 1987, 1995; Lin, 1991, 1994a, b). High-velocity rock-friction experiments carried out on friction welding (Spray, 1987, 1988, 1993) and a high-velocity rotary shear apparatus (Tsutsumi & Shimamoto, 1994, 1996, 1997a, 1997b; Shimamoto & Lin, 1994; Lin & Shimamoto, 1998; Hirose et al., 2000; Hirose, 2002; Hirose & Shimamoto, 2005a), have produced friction melts artificially, under similar conditions to those encountered during seismic faulting in the crust, and further confirmed a melt origin. These experiments have also made important findings in regard to pseudotachylyte generation and its effect on seismic slip.

Following on from the initial description of pseudotachylyte at the beginning of this chapter, in this thesis, the term ‘pseudotachylyte’ is used to describe dark coloured, fine-grained veins and injection structures that preserve evidence of having been through a melt phase, as a result of frictional heat generated at seismic slip rates. Therefore the molten material can be used as a paleoseismic indicator, and has the potential to allow constraint of earthquake source parameters, as has been accomplished by many authors to date (e.g. Mckenzie & Brune, 1972; Sibson, 1975; Lin et al., 2003, 2005; Barker, 2005; Hirose & Shimamoto, 2005a, b; Noda & Shimamoto, 2005; Di Toro et al., 2005, 2006a, b, 2008; Pittarello et al., 2008).

1.3 Identifying Pseudotachylyte

As discussed in the previous section, pseudotachylyte often closely resembles, and is also often intimately associated with ultracataclasites and ultramylonites. By the definition of pseudotachylyte used in this thesis, pseudotachylyte must preserve evidence of having been through a melt phase. The following list summarises key indicators within pseudotachylyte veins that provide evidence that indicate a melt-origin of pseudotachylyte, rather than a cataclastic origin, and gives examples of authors who have presented such evidence:


II. Occurrence of vesicles and amygdale structures exsolved from a melt (Maddock, 1983; Maddock et al., 1987; Lin, 1991, 1994a), and also immiscible mineral phases (Magloughlin, 1992, 2005)

III. Presence of chilled margin, indicating rapid quenching of melt in contact with cooler host rock (Philpotts, 1964; Lin, 1994a)

IV. Occurrence of microlites and spherulites that are stable and form at only high temperatures and with rapid quenching (Maddock, 1983; Macaudière et al., 1985; Spray, 1988; Toyoshima, 1990; Lin, 1991, 1994a, b)

V. Flow textures as indicators of a fluidized phase (Lin, 1994a, b)

VI. Certain systematic chemical relations between pseudotachylyte and the host rock, or precursor cataclasites (Maddock, 1992; Magloughlin, 1992; O’Hara, 1992; Curewitz & Karson, 1999; Warr & Pluijm, 2005)

Magloughlin and Spray (1992) suggest the presence of one or more of these features, except V and VII, provide strong evidence of a melt origin, or are more simply explained by a melt origin than other mechanisms. During identification of pseudotachylytes investigated in this study, this suggestion was followed.

1.4 Seismological Origin

To this date, the only fault-related rock that is recognised as being a strong paleoseismic indicator or an unequivocal ‘seismic fossil’, is pseudotachylyte (Cowan, 1999). The generation of pseudotachylyte by seismic faulting was initially proposed by Jeffreys (1942), and has subsequently been supported by many authors (McKenzie & Brune, 1972; Sibson, 1975, 1977, 1980a; Spray, 1987; 1995). The significance of pseudotachylyte as a ‘seismic fossil’ is large, as it can provide a myriad of information about stress levels and energy partitioning during seismic faulting, as well as information regarding earthquake rupture processes and the rheology of pseudotachylyte-bearing fault zones. In the following paragraphs, I present the fundamentals of earthquake mechanics and the basis for frictional melting during seismic slip.

1.51 Stress Theory, Brittle Failure and the Frictional Strength of Faults

In fluid-saturated crust, the three principle compressive stresses \((\sigma_1 > \sigma_2 > \sigma_3)\) are reduced by the pore fluid pressure \((P_F)\), such that their relationship can be re-written as (Hubbert & Rubey, 1959):

\[
\sigma_1' = (\sigma_1 - P_F) > \sigma_2' = (\sigma_2 - P_F) > \sigma_3' = (\sigma_3 - P_F)
\]

The normal and shear stresses resolved on fault plane that contains \(\sigma_2\), can be related to the effective principal stresses as follows:

\[
\sigma_n' = \frac{(\sigma_1' + \sigma_3')}{2} - \frac{(\sigma_1' - \sigma_3')}{2} \cos 2\theta
\]

\[
\tau = (\sigma_1' - \sigma_3') \sin 2\theta
\]

where \(\sigma_n'\) is the effective normal stress acting on the fault plane, \(\tau\) is the shear stress acting on the fault plane, and \(\theta\) is the angle between the plane and the greatest compressive stress, \(\sigma_1\). At a depth, \(z\), the effective vertical stress, \(\sigma_v'\), provided by the overlying rock column of average density, \(\rho\), and the acceleration due to gravity, \(g\) \((9.81 \text{ ms}^{-1})\), is expressed as:

\[
\sigma_v' = \rho g z (1 - \lambda_f)
\]

where the pore fluid factor, \(\lambda_f = P_F/\sigma_v\), is a measure of the pore-fluid pressure level. Brittle failure occurs in intact, isotropic, homogenous rock when the Mohr-Coulomb criterion is satisfied, being that shear failure occurs when:

\[
\tau = \mu_i \sigma_n' + C_0 \quad \text{provided that} \quad (\sigma_1' - \sigma_3') > 4T_0
\]

where \(\mu_i\) is the internal coefficient of friction, \(C_0\) is cohesive strength of the intact rock, and \(T_0\) is the tensile strength of the rock. For most rocks, the cohesive strength \((C_0)\) is approximately twice the tensile strength \((T_0)\), and the internal friction coefficient \((\mu_i)\), lies between 0.5 and 1.0. These shear fractures usually form at an orientation of approximately
22° - 32° to $\sigma_3$. Pseudotachylytes are generally thought to be related to shear fractures, however brittle failure can occur in two further modes: extension fractures, where failure occurs when $P_f = \sigma_3 + T$, and only if $(\sigma_1 - \sigma_3) < 4T$; and hybrid extensional-shear fractures, which both have to satisfy a Griffith failure criterion.

Once a shear fracture has initiated, a state of zero-cohesion along the fault is achieved. The static frictional shear strength of an optimally oriented, cohesionless fault can be expressed by:

$$\tau_f = \mu \sigma_n$$

(6)

where $\mu$ is the static coefficient of friction. This condition will only be maintained if the aggregate (gouge) in the fault zone is not cemented or hardened to regain cohesion. Byerlee (1978) suggested a range for the static coefficient of friction in the brittle crust of $0.6 < \mu_s < 0.85$, based on experimental rock mechanics. This was later supported by field observations (Sibson, 1994), and borehole stress measurements (Townend & Zoback, 2000).

The frictional strength of a cohesionless fault is controlled by the static coefficient of friction, and the effective normal stress acting on that fault, as shown in equation 6. Failure would therefore be expected to occur if the effective normal stress is reduced or the shear stress is increased. Theoretical shear strength profiles for thrust and normal faults, where $\sigma_c = \sigma_3$ for thrust faults, and $\sigma_c = \sigma_1$ for normal faults (Anderson, 1942; Sibson, 1974), are shown in figure 1.01.

As can be seen from figure 1.01, for a thrust and a normal fault at the same depth and under the same fluid pressures, the thrust fault would have a much greater shear strength, or shear resistance to failure. However, field evidence indicating this excessive shear resistance are not observed (Sibson & Toy, 2006). However, fluid overpressures generally associated with thrusting and compressional regimes are likely to lower the frictional strength of thrust faults (Sibson, 2003a).
1.52 Earthquake Mechanics, Seismic Energy and Pseudotachylyte Production

Frictional sliding is assumed to begin when the ratio of the resolved shear ($\tau$) and normal stresses ($\sigma_n$) on a fault surface reaches a yield value for $\mu_s$, when the Mohr-Coulomb criterion for failure in intact or cohesionless rock is met. Once the sliding initiates, the frictional resistance decreases to a lower value, if the kinetic coefficient of friction, $\mu_k$, is exceeded by $\mu_s$ (Scholz, 1990). This dynamic instability results in rapid acceleration of the bounding blocks and an associated drop in the shear stress resolved on the fault plane, known as the static stress drop, $\Delta\tau$. This process is known as ‘frictional stick-slip’ behaviour, and it is widely thought that it is the process by which earthquakes operate. The frictional strength of a fault, as outlined in the last section controls the ease of which stick-slip behaviour can occur, which is in-turn controlled by the normal stress and the static coefficient of friction, as shown in equation 6. The acceleration results in a radiation of seismic energy, as stored elastic strain energy in the surrounding rocks is released during this rapid motion. Once sliding stops, the friction coefficient returns to $\mu_s$, and strain accumulates once more in the period of no motion, until the elastic yield is reached, and the instability once again manifests itself as a sudden slip displacement.

Stick-slip behaviour on faults can be simulated by the block and spring model (e.g. Scholz, 1990), whereby pulling on a block on a horizontal surface via a spring accumulates elastic strain in the spring, and the static friction increases to a point where it meets a yield value and the block suddenly slips, lowering the frictional resistance to a lower dynamic value and releasing some of the stored strain energy in the spring. The movement generates frictional heat at the slip surface, and also radiates energy as seismic waves. Stick-slip experiments show that the stress drop for each slip increment is only a small fraction of the average shear stress, $\tau_{av}$. It is still generally unknown whether the stress drop is only a small fraction of the average shear stress for large crustal earthquakes, although the phenomenon has been observed from mining induced earthquakes (e.g. McGarr, 1994).

An earthquake can be considered to be a dynamically running shear crack through the crust, and slip occurs at any point for a certain duration known as the rise time. The slip on the fault can terminate when the elastic strain is relieved, when material heterogeneities in the surface interfere with movement. Slip can also terminate abruptly when the rupture propagates into a velocity strengthening stability regime, as outlined by Scholz (1998), where the earthquake will produce a negative stress drop and rapidly terminate. The earthquake rupture propagates out from the focus (source of initial slip) at approximately 3 km s$^{-1}$ (Sibson, 1989), and ruptures an area on the fault surface, $A$, before terminating by one of the above mechanisms. The average slip increment $\bar{u}$, is averaged over the rupture area, $A$, and can be related to the seismic moment, $M_0$, by the following equation:

$$M_0 = G\bar{u}A$$ (7)

where $G$ is the shear modulus (typically 3 x 10$^4$ MPa). The rupture area is often estimated by either multiplying the length by width of rupture, or by considering the distribution of microseismicity and aftershock activity on a cross-sectional area of the fault plane (e.g. Sibson, 1989). The seismic moment can be used to derive a moment magnitude, $M_w$, based on the moment magnitude relationship of Hanks and Kanamori (1979), which is:

$$M_w = \frac{2}{3} \log_{10} M_0 - 9.1$$ (8)
When an earthquake occurs the total energy ($E_T$) involved in the process is partitioned into 3 components: the radiated energy released as seismic waves ($E_R$); the frictional energy loss on the surface of the fault ($E_F$), which is all converted to heat, $Q$; and the fracture energy ($E_G$), which is a form of latent heat required to create the rupture surface (part of which is now known to be converted to heat). $E_F$ is generally considered to account for most of the energy budget, and does not directly influence the rupture dynamics (e.g. Beeler et al., 2003). $E_R$ can be measured easily by far field instruments and the ratio $E_R/E_T$ is known as the radiation efficiency, $\eta$, and is an important parameter in earthquake mechanics, although its quantification is difficult to attain due to the uncertainties in estimating the $E_T$. $\eta$ can be related to the average shear stress $\tau_{av}$ on a fault during slip by:

$$\eta \tau_{av} = \frac{GE_R}{M_0} = \frac{E_R}{uA} = \tau_a = \tau_{av} - \tau_{f(\text{av})}$$

(9)

where $\tau_a$ is the apparent stress, which is the fraction of the total work expended that is available for seismic radiation, and can also be estimated from the difference between the average shear stress during slip ($\tau_{av}$), and the average frictional shear resistance ($\tau_{f(\text{av})}$) that resists the slip. $G$, $E_R$ and $M_0$ can all be measured, however, using these to solve equation 9 only gives the product of the seismic efficiency and not the individual values of $\eta$ and $\tau_{av}$. Therefore, $\tau_a$ is easily estimated, whereas $\tau_{av}$ is not, introducing a paradox, and making it extremely difficult to resolve absolute values of dynamic shear stress ($\tau_f$), resolved on a fault during seismic slip.

If $E_T$ were known from measurement of $E_R$, and estimation of $E_G$ and $E_F$, then the stress paradox will be solved, and an estimated magnitude of $\tau_{av}$ could be found. The portion of $E_G$ not converted to heat can be estimated from examinations and measurements of fracture surface area in fault damage zones (Chester et al., 2005). Pseudotachylytes, if assumed as the only product of co-seismic frictional heating on a fault, can be used to estimate the amount of heat energy, $Q$, released during an event. As suggested by Scholz (1990, p. 133), neglecting the $E_G$ and $E_R$ based on the assumption that their magnitudes are negligible with respect to the work done by friction, the general energy balance for seismic faulting becomes:

$$W_f \approx Q$$

(10)

where $W_f$ is the work done by faulting, and $Q$ is the frictional heat generated on the sliding surfaces during slip on a fault. Heat is generated on a fault surface during slip according to:

$$Q = \tau_{f(\text{av})} \bar{u}$$

(11)

If pseudotachylytes are present on a fault surface then, the heat energy generated during the frictional failure at seismic slip rates was not able to diffuse away from the surface fast enough, therefore an associated temperature rise, $\Delta T$, through the thickness of the shearing zone, $z$, occurs according to:

$$\Delta T = \frac{\tau_{f(\text{av})} \bar{u}}{\rho c_p z}$$

(12)
where \( \rho \) is the density of the rock, \( c_p \) is the specific heat capacity, and assuming that heating can be considered adiabatic, based on short rise times and a slip zone thicker than a few millimeters. Once a friction melt layer is formed it may act as a lubricant (depending on its viscosity), and reduce the frictional shear resistance \( (\tau_f) \) of the surface and cause dynamic weakening, which leads to large stress drops. Melt lubrication as a dynamic weakening mechanism can explain the indication that there are low \( \tau_f \) values associated with many earthquakes, based on interpretation of seismological and geophysical data (Heaton, 1990) and heat flow measurements on active faults (Brune et al., 1969).

Di Toro et al. (2005) used the relationship shown in (12) to show that the thickness of a friction melt layer, m, is related to the average frictional shear resistance by:

\[
\tau_f = \frac{c_p \Delta T + \Delta H_{fus} (1 - \phi)}{\mu} m
\]

(13)

where \( \Delta H_{fus} \) is the latent heat of fusion, \( \phi \) is the proportion of unmelted porphyroclasts, and \( \Delta T \) is defined as \( T_m - T_R \), the difference between the initial melt temperature, \( T_m \), and the country rock ambient temperature, \( T_R \). Provided that slip occurs over a localized zone, it can be predicted from equations 11 and 12 that most moderately sized earthquakes in the seismogenic zone will produce friction melts. Sibson and Toy (2006) and Toy (2007) calculated thicknesses of friction melt layers produced by earthquakes, for \( \mu = 1 \) m and 10 m at various depths in dry crust for the various modes of faulting, using the frictional shear resistance profiles shown in figure 1.01. The results of these calculations, which used: \( \rho = 2750 \) kg m\(^{-3}\), \( c_p = 1200 \) J kg\(^{-1}\) °C\(^{-1}\), \( \Delta H_{fus} = 5 \times 10^5 \) J kg\(^{-1}\), \( \Delta T \) - calculated for various depths using a geothermal gradient of 25°C km\(^{-1}\) and \( T_m = 1200 \) °C, and \( \phi = 0.3 \); are shown in table 1 for thrust and strike slip faults for slip increments of 1 m.

<table>
<thead>
<tr>
<th>Depth</th>
<th>Strike Slip Fault</th>
<th>Thrust Fault</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 km</td>
<td>( \tau_{f(\text{av})} = 5.8 ) MPa; m = 1.3 mm</td>
<td>( \tau_{f(\text{av})} = 11.8 ) MPa; m = 2.6 mm</td>
</tr>
<tr>
<td>5 km</td>
<td>( \tau_{f(\text{av})} = 29 ) MPa; m = 6.3 mm</td>
<td>( \tau_{f(\text{av})} = 59 ) MPa; m = 13 mm</td>
</tr>
<tr>
<td>10 km</td>
<td>( \tau_{f(\text{av})} = 58 ) MPa; m = 14 mm</td>
<td>( \tau_{f(\text{av})} = 118 ) MPa; m = 29 mm</td>
</tr>
<tr>
<td>15 km</td>
<td>( \tau_{f(\text{av})} = 87 ) MPa; m = 24 mm</td>
<td>( \tau_{f(\text{av})} = 177 ) MPa; m = 48 mm</td>
</tr>
</tbody>
</table>

Table 1.01: Average shear resistance (\( \tau_{f(\text{av})} \)), associated depth and calculation of the thickness of a pseudotachylyte melt layer (m) for a slip increment of 1 m, for strike slip and thrust faults. After Toy (2007).

Table 1.01 shows that for even a slip increment on a thrust fault of only 1 m, at 5 km depth in the crust the seismic faulting should generate enough heat to form a 1.3 cm melt layer. This indicates that a vast volume of pseudotachylyte must be produced during a faults active life, when the number of earthquakes that take place on a fault, and the range of depths and slip increments associated with the seismicity are taken into consideration. Frictional dissipation estimates (10 – 100MW m\(^{-2}\)) are also such that the production of friction melts should be widespread (Sibson & Toy, 2006). For a relevant example, Sibson and Toy (2006) calculated that a total thickness of pseudotachylyte of well over 1km should be exposed in the Alpine Fault Zone. These vast volumes of pseudotachylyte are not observed in
the real world, and pseudotachylites are thought to be rare occurrences in fault zones all-together. This unexpected scarcity of pseudotachylites raises some interesting questions:

- Are pseudotachylites rarely generated during seismic slip on a fault plane due to some melt-inhibiting mechanism? OR
- Are pseudotachylites commonly generated but rarely preserved in recognizable form? Have most of them been structurally or chemically overprinted during exhumation?

It is thought that certain factors may inhibit melting during seismic slip, with the most likely factors being the presence of a fluid pressure during seismic faulting and dynamic weakening mechanisms that lower kinetic shear resistance. Several postulated mechanisms for dynamic mechanisms exist, with the most notable being thermal pressurization of fluid within the slip zone (Sibson, 1973, 1980b; Lachenbruch, 1980; Mase & Smith, 1987; Andrews, 2002; Noda & Shimamoto, 2005; Rempel & Rice, 2006). Thermal pressurization is the transient increase in fluid pressure in a seismic slip zone due to the thermal expansion of the fluids caused by frictional heating. The increase in fluid pressure lowers the frictional shear resistance, \( \tau_f \), according to (Lachenbruch, 1980):

\[
\frac{\delta P_F}{\delta t} = - \frac{\delta \tau_f}{\mu \delta t}
\]

where \( t \) is the duration of the event. The lowering of frictional shear resistance inhibits further temperature increase during slip, and therefore melting of the rock. However, the magnitude of the pore pressure rise is dependent on the material properties of the rock and the fluids. Lachenbruch (1980) concluded that the most important material properties are the permeability of the rock and the dilatation characteristics of the pores during faulting, and if in excess of \( \sim 100 \text{mD} \) or 2-3%, respectively, the coupling effect of temperature, pore pressure and frictional resistance can be ignored, and thermal pressurization will not occur. It has long been thought that pseudotachylite, due to the above mechanism, cannot be produced in fault zones that contain fluids (e.g. Sibson, 1980b), and should be a dry process. However, it is now thought that thermal pressurization is a common process during seismic slip if it is taken as the weakening mechanism (Noda & Shimamoto, 2005; Wibberley & Shimamoto, 2005). Furthermore, the presence of fluids during pseudotachylite formation is now considered to be possible in the light of results from geochemical and fluid inclusion studies (Parry, 1998; Boullier et al., 2001; O’Hara & Sharp, 2001).

Another, perhaps more obvious factor that could contribute to the inhibition of frictional melting is whether the thickness of the displacement zone is sufficiently thin to allow the heat from frictional dissipation to induce an adequate temperature rise for melting to occur. It is suggested that slip needs to be restricted to zones of \( \sim < 4 \text{cm} \) for frictional heating to be adiabatic, and cause sufficient temperature rise (Sibson, 2003b).

In summary, the presence of pseudotachylites as veins within a fault zone is a strong indicator of paleoseismic activity. They can provide valuable insights into the nature of earthquake processes and rupture propagation, provided they are not structurally or chemically overprinted. To obtain realistic estimates of earthquake source parameters (\( \tau_c \), slip weakening distance), exposures of pseudotachylite must be sufficiently large, and contain offset markers, in order to estimate volume of pseudotachylite and measure slip increments, respectively (Di Toro et al., 2005). A new, multidisciplinary approach for retrieving information on earthquake source mechanics from pseudotachylite-bearing
fault zones has been postulated by Di Toro et al. (2008), where by integrating classical field structural analysis with 3D laser scanning mapping, allows for the reconstruction of the geometry of a fault zone with millimeter scale precision. Understanding the geometry of a fault is important as it controls the nucleation, propagation and termination of earthquakes (e.g. Scholz, 2002), and also allows for realistic estimates of earthquake source parameters.

1.5 Scope and Aims of Research

This research includes a field based, geochemical and microstructural analysis of pseudotachylytes that are exposed along the Alpine Fault zone of New Zealand. Pseudotachylytes from 3 different host rock types are investigated, derived from footwall, hangingwall and core portions of the fault. Estimations, using parameters measured in the analyses, of the physical conditions during seismic faulting are carried out, such as ambient temperature conditions, temperature of the melts and stress conditions.

The main aim of the research is to characterise and compare the different pseudotachylytes hosted in 3 different settings to draw conclusions on: the overall rheology of the brittle portion of the Alpine Fault; the nature and processes that have contributed to pseudotachylyte formation in the 3 settings, especially in the fault core; and the paleoseismic implications for the Alpine Fault based on these ‘seismic fossils’.
Chapter 2

The Alpine Fault Zone, New Zealand

2.1 Introduction

Exhumed rocks of the Alpine Fault Zone host abundant pseudotachylyte, as has been recorded by various previous workers (Wallace, 1976; Sibson et al., 1981; Bossière, 1991; Warr et al., 2003; Warr & Pluijm, 2005; Toy, 2007). This study investigates pseudotachylytes derived from three contrasting host rocks, derived from the core, footwall and hangingwall of the Alpine Fault. The study areas include two localities; Gaunt Creek, in the Waitangi-taona Valley, and Harold Creek, both within the central section of the Alpine Fault (figure 2.02). This chapter aims at providing a field context in which to place these pseudotachylytes, by introducing the regional and local geological settings by reviewing past work on the Alpine Fault.

2.2 Structure, Kinematics and Age

The currently active Alpine Fault, with an average strike of 055°, defines part of a major plate boundary between the Australian plate to its west, and the Pacific plate to its east (figure 2.01). The fault is the principal structural feature of the South Island of New Zealand, extending onshore approximately 400km from Milford Sound in Fiordland to Maruia Valley near Arthurs Pass, forming a sharp northwestern boundary to the Southern Alps. At its northern termination it splays into subsidiary faults that continue northeastward as part of the Marlborough Fault System, while in the south the fault extends offshore and is more or less continuous with the Puysegur subduction zone (Figure 2.01). The fault trace is straight on a regional scale, but locally, in the central section, the fault forms 1-10 km long segments of oblique thrusts that dip moderately (~ 30° - highly variable) to the southeast that are separated by sub-vertical strike-slip segments. This serial partitioning is the result of local, shallow perturbations of the stress field due to variations in the surface topography over time (Norris & Copper, 1995, 1997). The dip of the fault, based on surface structural data of Sibson et al. (1979), is thought to steepen to ~ 45° at around 2-4 km depth, and then shallow again to ~ 30° at ~ 25 km depth, based on seismic reflections (Kleffman et al., 1998). The shallowing of the fault at depth has been suggested to represent a linkage to a basal décollement (Koons, 1990; Little et al., 2002), which may form the boundary between an observed upper, low velocity zone, and a lower, higher velocity zone (Little et al., 2005), that form ~ 35 km and ~10 km, respectively, of a ~ 45km crustal root beneath the Southern Alps.

The Alpine Fault accommodates approximately 70% of the total Pacific-Australian plate motion, with most of the slip occurring within a narrow, <10 – 50 m wide gouge and cataclasite zone (Norris & Cooper, 2001). A ~ <100m wide zone, predominantly within the hanging wall, of secondary faulting and associated gouge zones, defines the faults’ damage zone (Norris & Cooper, 2007). The fault has an apparent dextral strike-slip displacement of 480 km (Wellman, 1955), which offsets a significant marker unit, the Dun Mt Ophiolite Belt (figure 2.01), and a ~ 70km convergent displacement via reverse oblique slip motion (Walcott, 1979; Allis, 1986). Offset late Quaternary markers, such as gravel terraces, have constrained the slip on the fault to be predominantly oblique (e.g. Sibson et al., 1979), with present
Figure 2.01: Map of the South Island of New Zealand and the Pacific-Australian plate boundary. The dextral offset of the Dun Mountain Ophiolite Belt is obvious. Eastern Province and Western Province rocks are juxtaposed along the fault. The fault splays at its northern end, while at its southern end it runs offshore and joins the Puységur Subduction System.
day rates estimated at 23-25 mm yr\(^{-1}\) dextral strike-slip and >10 mm yr\(^{-1}\) reverse dip-slip (Cooper & Norris, 1994; Norris & Cooper, 2001). However, dip-slip rates of >10 mm yr\(^{-1}\) of the fault fall off from this maximum to the north and south of the central section (Norris & Cooper, 2001). The component of reverse slip has resulted in uplift of the eastern side, and exhumation of deformed hangingwall rocks to the southeast of the fault. It is likely these rocks represent the nature of the deformation within the fault zone at depth, in both the brittle and ductile regimes. Uplift rates of 8-10 mm yr\(^{-1}\) have been calculated, based on offset river terraces, and raised marine benches (Bull & Cooper, 1986). 1.1 Ma pseudotachylytes (Warr et al., 2003) that do not overprint any brittle features derive similar uplift rates, if it is assumed they formed at or near the base of the regional seismogenic zone (Toy, 2007).

The Southern Alps are the result of the uplift along the Alpine Fault, and rise to > 3500m its central section (figure 2.02). The section between Harold Creek and Fox Glacier, otherwise known as the central section of the fault, coincides with the highest exhumation rates and heat flow, the youngest mineral ages (Little et al., 2005), and the high localisation of convergent strain (Norris & Cooper, 2007). This zone of highly localised strain, in the centre of the oblique interplate collision, is thought to be due to the interplay between high erosion rates and exhumation, leading to thermal weakening and intense localisation (Ellis et al., 2001; Koons et al., 2001). The Southern Alps and a highly asymmetric erosion pattern couple with the Alpine Fault to form a two-sided orogenic wedge (Koons, 1990, 1994), in which deformation yields the distinctive, paired outcrop pattern of a narrow band of high-grade amphibolite facies schists and mylonites beside a broad region of lower-grade shallow crustal schists and greywackes (figure 2.02). The
Alps also help to express the trace of the Alpine Fault in Westland, where it separates their high, dramatic topography from the low, flat topography of the West Coast.

The Alpine Fault is thought to have originated as a <100 km wide zone of dextral shearing in the Oligocene-Miocene (Carter & Norris, 1976; Cooper et al., 1987; Sutherland, 1999; Little & Mortimer, 2001). This was followed by convergence across the Pacific-Australian plate boundary commencing in the Neogene (Sutherland, 1995), with the onset of uplift between 8-5 Ma (Ar-Ar and fission track ages, Batt et al., 2004), and a marked acceleration in uplift of the Southern Alps at 5 Ma (increased deposition in offshore Waiho-1 core, Sutherland, 1996). This acceleration at 5 Ma is likely to be the point at which strain localised to form the Alpine Fault zone (Toy, 2007). It is inferred that the slip rates in the Alpine Fault zone, discussed above, have remained relatively constant over the last 5 Ma, due to consistency between continuous GPS measurements of plate motions and the predicted NUVEL-1A model, as shown by Beavan et al. (1999).

2.3 Regional Geology

The Alpine Fault juxtaposes basement rocks of terranes of the Eastern Province to the southeast of the fault, with those of the Western Province, to the northwest of the fault (figure 2.01).

The Western Province rocks, at least in the central section of the fault, are composed of a variety of lithologies. Ordovician, greenschist facies metasediments of the Greenland Group form part of the Buller Terrane. The Greenland Group metasediments are intruded by Devonian to Carboniferous granitoids of the Karamea Suite, and Jurassic to Cretaceous granitoids of the Darran, Separation Point, and Rahu suites. A sliver of the Fraser Complex is juxtaposed against the Greenland Group metasediments, some of the Cretaceous granitoids, and the Eastern Province rocks in the hangingwall, between the Wanganui River-Taramakau River section of the Alpine Fault (Rattenbury, 1991). The juxtaposition is attributed to vertical movement along the Fraser Fault and the Bald Hill Range Thrust. The Fraser Complex, studied in detail by Rattenbury (1987a, 1991), is composed of a metamorphic suite of metapelitic, migmatitic and hornblende amphibolite facies gneisses, intruded by granitoids and a more basic set of dykes. Post-Late Cretaceous mylonitisation (Rattenbury, 1987b) that affected the Fraser Complex, is unlikely to be related to the Alpine Fault. The Fraser Complex forms part of a regional basement to the Greenland Group in Westland (Rattenbury, 1991), and both have been considered as possible protoliths to the Alpine Fault zone rocks (e.g. Toy, 2007).

The Eastern Province rocks include the Torlesse, Caples (figure 2.01a), Murihiku and Brook Street Terranes, as well as the Aspiring lithologic association and the Dun Mountain Ophiolite Belt. The Torlesse Terrane greywackes form a wide belt that covers all of Canterbury and most of Otago to where they contact the volcano-sedimentary Caples Terrane along a NW-SE trending boundary in Central Otago (figure 2.01a). Pelites, metacherts and metabasites form the Aspiring lithologic association (Craw, 1981, 1983) in the NW of this boundary, on the Torlesse side. These terranes were metamorphosed in the Mesozoic as they collided with one another, forming the 150 km wide, two-sided, structural arch of the Otago Schists (Mortimer, 2003; figure 2.01). The Otago Schist is a subgroup of the regionally extensive Haast Schists, of which the Alpine Schist is also a part (Mortimer, 2000). The Alpine Schists form a 10-20 km wide strip, running sub-parallel to the Alpine Fault (figure 2.01), of steeply dipping, high-grade schists that decrease in grade from amphibolite facies at the Alpine Fault in the west to prehnite-pumpellyte facies in the east (figure 2.02). The mylonites within 1-2 km east of the fault contain metabasic horizons, and it has been suggested that mylonites of
the fault zone are derived from the Aspiring lithologic association, due to the abundance of metabasite (Little, et al., 2002).

Considering the regional geology is important here, as there is a possibility that the fault rocks in the Alpine Fault zone may be derived from any of the regional lithologies. It is more unlikely, however, at least in the studied central section of the fault, that terranes south of the Dun Mountain Ophiolite Belt are presently incorporated as the belt forms a ‘barrier’ to the central fault zone (figure 2.02).

### 2.4 Fault Zone Geology

As already mentioned, uplift of hangingwall rocks to the southeast of Alpine Fault zone has exhumed deformed fault rocks and deep-seated crustal rocks adjacent to the present surface trace of the fault. A general cross-section of the fault zone from NW to SE (figure 2.03), shows that the fault rock sequence passes through basal gouge and cataclasites in contact with Quaternary gravels or Western Province basement, into ultramylonite, mylonite and protomylonite zones with a distance totaling up to 1 km, before passing into Alpine Schist. The best exposure of a continuous section of these fault rocks is at Gaunt Creek (figure 2.02), which occurs on the longest thrust segment of the fault (Cooper & Norris, 1994). There are also numerous creeks that cross the fault that expose good sections of most or parts of the deformed zone. The following paragraphs briefly describe the main fault rock types within the Alpine Fault zone, as observed by previous field workers. Most of these rocks, especially the cataclasites and pseudotachylytes, will be described in more detail from my own field observations in chapter 3 as they play host to pseudotachylyte veins. The nomenclature used for the fault rock descriptions, and for the thesis a whole, is that of Sibson (1977).

![Figure 2.03](image-url)
2.41 Cataclasites and Gouges

Cooper & Norris (1994) provided a description of the cataclasite zone of the Alpine Fault, by examination of the cataclasites at Gaunt Creek in the central section of the fault. They recorded a pale green cataclasite zone, 30 m thick, composed of cataclased mylonite and crushed vein quartz, immediately overlying ~ 50-70 cm zone of indurated grey gouge, marking the Alpine Fault principal slip surface. They also noted a black heterogenous unit structurally overlying the green cataclasite, which they classified as bands of ultracataclasite alternating with lenticular zones of cataclased mylonite, and containing lenticular shear zones decorated with pseudotachylyte glass. Based on compositions of garnet fragments they concluded that the cataclasite has been derived from hangingwall Alpine Schist-derived mylonites, however, the possibility of footwall granitoids of the Western Province as protoliths to the zone cannot be overlooked. A later, more general account of the Alpine Fault by the same authors, (Norris & Cooper, 2007), provided general thicknesses of the cataclasite zone of 10-50m, and noted that the thickness is highly variable due to internal imbrication in order to achieve critical thicknesses and strength necessary for continued westward overthrusting.

A detailed petrographic and microstructural analysis of deformation mechanisms in the cataclasites and gouge zones was carried out by Warr and Cox (2001), who concluded that progressive stages of anhydrous cataclasis and frictional melting, hydrous chloritisation and growth of swelling clays in the matrix lead to significant weakening of the fault during its exhumation. Toy (2007) re-classified the green coloured cataclasites at Gaunt Creek as protocataclasites on the microstructural basis that no distinct fine-grained ‘matrix’ exists, and also re-classified the overlying ultracataclasite of Cooper & Norris (1994), as partially cataclased ultramylonites owing to a strongly disjunctive cleavage offset by gouge filled shears. The cataclasite zone is of particular interest to this study as it is hosts abundant pseudotachylyte veins, therefore it will be described in much more detail in chapters 3, 4 and 5.

2.42 Mylonites

Reed (1964) first described mylonites exposed in the hangingwall of the Alpine Fault in the upper Kokotahi Valley. Sibson et al. (1979, 1981) described the fault zone at many localities and created a composite cross-section across the fault from granitoids in the footwall through cataclasites and into the mylonites. In this, they differentiated the mylonites into augen, green and schist-derived mylonites (figure 2.03), across their section, and followed Reed (1964) in stating that the augen mylonites were derived from granitoids in the hangingwall. Sibson et al. (1979, 1981) were unsure of the protolith of the green mylonites, however they state the presence of feldspar porphyroclasts can be used as evidence for a granitoid protolith. Subsequent studies have found that the green mylonites are strongly and locally retrogressed and altered (e.g. Campbell, 2002), and observations suggest that they are derived from both Western Province granitoid and Eastern Province schist protoliths (e.g. Prior, 1988; Toy, 2007). Augen mylonites have been petrographically, geochemically and microstructurally analysed by Toy (2007), in which it was concluded they are derived from a Western Province granitoid protolith. Reed (1964) considered augen and green mylonites to be distinct groups of mylonites, and Sibson et al. (1979, 1981) extended their distribution to the entire fault zone. However, these authors’ descriptions were of the northern section of the fault zone (e.g. Saddle Creek, Douglas Creek and Harold Creek sections), and these mylonites are now known to be relatively rarely exposed in the central and southern sections of the, where Alpine Schist-derived mylonites volumetrically dominate along, and across strike (Toy, 2007).
The Alpine Schist-derived mylonites described by Sibson et al. (1979, 1981), have been re-classified more broadly into ultramylonite, mylonite and protomylonite subzones by various workers (Prior, 1988; Wright, 1998; Campbell, 2002; Norris & Cooper, 2003, 2007; Toy, 2007). Within these mylonites, there are various lithologies due to the heterogeneous nature of the protolith to the Alpine Schists. The lithologies include abundant metabasite horizons, and a few metachert horizons within a dominantly quartzofeldspathic block, that in itself contains quartz-feldspar rich and mica-rich end-members, of which many mylonites are intermediary (Toy, 2007). Data from various mapping projects of the mylonite zone along the Alpine Fault show that it can generally be considered to be a 1-1.5 km wide zone (figure 2.02) of a southeastward progression from the fault of ultramylonite, mylonite and protomylonite dipping ~33° to the southeast, essentially parallel to the Alpine Fault plane (Norris & Cooper, 2007). Two types of differently derived mylonites that host pseudotachylytes are examined in this study, and will be described in more detail in chapter 3.

2.43 Pseudotachylytes

Pseudotachylyte was first described in the Alpine Fault zone by Wallace (1976), 6 km south of the Moeraki River, in the southern section of fault. Wallace (1976) preferred the term ‘hyalomylonite’ for the dark veinlets he described and analysed. Reed (1964) described dark, flinty ‘ultramylonite’ veins in the Alpine Fault zone, later confirmed by Sibson et al. (1981) to be friction melts, therefore were the first records of pseudotachylyte in the fault zone, although unknown to the author at the time.

It is now known that pseudotachylyte is widely distributed within the Alpine Fault zone, and is hosted in all the fault rocks. The best, easily accessible exposure at present of pseudotachylytes is at Harold Creek (figure 2.02) where there are voluminous, thick veins hosted within the mylonites. However, many of them are in the form of fallen blocks, making it impossible to place them within a well-defined field relationship. Studies focused on local pseudotachylyte occurrences have subsequently been carried out, especially at a Harold Creek. Adams (1981) obtained a K-Ar age of 9.8 Ma for pseudotachylyte from Harold Creek, to help estimate uplift rates and the thermal structure of the fault zone. Seward and Sibson (1985) carried out fission-track dating on a pseudotachylyte from NE of the Haast River, obtaining an age of 0.43 Ma. Bossière (1991) described, structurally and in detail, in situ pseudotachylytes at Harold Creek and the Wanganui River, and also carried out microprobe analyses on pseudotachylytes from two different host rocks (green and schist-derived mylonites). Warr et al. (2003) dated pseudotachylyte in a fallen block from the south bank of the Wanganui River, and arrived at an age of ~1.1 Ma, concluding that the vein was composed of multiple melt layers indicating cyclic frictional heating and pulses of pseudotachylytes occurring during a single, large earthquake episode. Warr & Pluijm (2005) analysed the same pseudotachylyte hosted in mica-rich mylonite, and another pseudotachylyte hosted within a more mafic, green coloured mylonite. They concluded that the two mylonites were most likely derived from Alpine Schist, and that the compositional variation of the veins can be explained by crystal fractionation of the melts. Cooper and Norris (1994) described fresh pseudotachylyte glass (as mentioned in section 2.41) from the ultramylonite zone at Gaunt Creek, and reported an analysis in Norris and Cooper (2007), concluding that it was clearly not an equilibrium minimum partial melt, characterised by a K-rich and Fe-Mg-poor composition. The most comprehensive macro- and microstructural investigation of the Alpine Fault pseudotachylytes to date is by Toy (2007), in which she described in detail the field relationships of the various types of pseudotachylytes from various locations along the Alpine Fault, and included estimations of ambient crustal conditions during seismic slip and implications for the rheology of the fault zone. Although this project has a different scope to the comprehensive field and petrographic
investigation of the pseudotachylytes by Toy, it hopes to add to her findings and answer certain questions arising from her work.

2.5 Conceptual Fault Zone Models

A general, conceptual fault zone model of how processes of deformation interact and vary with depth in the crust has been compiled from data obtained from regions containing exhumed active and inactive shear zones, termed the ‘Sibson-Scholz’ fault zone model. Sibson (1977) first proposed the model and it was later refined in a more site-specific manner by Scholz (1988). The model suggests that in the upper, brittle crust, localised zones deform by elastico-frictional mechanisms producing fault rocks with randomly oriented fabrics (e.g cataclasites, pseudotachylytes), and with depth there is a gradual transition to crystal-plastic deformation in the ductile crust. According to the model, at depths of 10-15 km, between the brittlely deforming upper crust and the ductily deforming lower crust, there exists a transition zone commonly referred to as the brittle-ductile transition zone, where semi-brittle deformation mechanisms operate.

This model can be applied to the Alpine Fault zone, and has been by various authors. Based on the results of Toy (2007), the latest Alpine Fault zone model is shown in figure 2.04. Firstly, the models shows the fault zone dipping at approximately ~45° to the southeast, and widening at depth which is suggested by the thickness of the uplifted mylonites compared to the exposed cataclasites. The model also shows incorporation of both footwall and hangingwall rocks into the fault zone, as evidenced by slivers of the different protoliths juxtaposed against one another at some surface exposures (e.g. Harold Creek), indicating the shear zone is an anastomosing network of high strain zones (Toy, 2007). These mylonites have been exhumed from depths of ~25-33 km (Norris & Cooper, 2007; Toy, 2007), from the
base of the crustal root where the action of a basal décollement is proposed (Koons, 1990; Little et al., 2002, 2005; figure 2.04). The figure shows elevation of the isotherms and therefore the temperature-dependent brittle-viscous transition in the hangingwall, which was first proposed by Koons (1987), and further constrained to ~6-10 km by Toy (2007) in this model. The ‘fault-locking’ depth proposed by Beavan et al. (1999) is also shown in figure 2.04 and is, in accordance with their modeling, the depth below which stable slip or shearing is occurring, and above which, material is storing elastic strain. The progressive development of the localisation of strain is shown well in figure 2.04, by fact the schist foliation is progressively being rotated into parallelism with the shear zone walls at high strain. A crustal strength profile that shows the changes in the strength profile during the interseismic period is also shown as an inset in figure 2.04. Of particular importance, pseudotachylyte formation distribution is also shown in the figure, at depths, with a voluminous group in the cataclasite zone at shallow levels, and another group at depth in the brittle-viscous transition zone.

2.6 Seismicity on the Alpine Fault

As pseudotachylytes are the products of earthquakes in the brittle crust, a general account of the seismicity and paleoseismicity of the Alpine Fault is necessary, and provides a seismological background, and possible distribution of pseudotachylytes that have formed during seismic activity of the fault.

![Figure 2.05: Seismicity on the Alpine fault as recorded by NZNSN (1990-1997), after Leitner et al. (2001). A relatively low-activity zone in the central section of the fault is obvious.](image)

The Alpine Fault is a seismically active fault, with repetitive seismic and interseismic periods due to the frictional stick-slip behaviour of the fault zone in the brittle crust. The seismicity map of the South Island as recorded by the New Zealand National Seismic Network (NZNSN) between the period 1990-1997, is shown in figure 2.05. The seismicity within the region of the Alpine Fault is generally thought to be moderate (Leitner et al., 2001), and the
central section of the fault is especially known for its low seismicity (Evison, 1971). North and south of this relatively inactive region the seismicity increases markedly, at the Hope Fault intersection in the north, and south of Jackson’s Bay in the south. The low seismicity reported in the central section of the fault (~Hari Hari to Haast) by Eiby (1971), Evison (1971), Scholz et al. (1978), Leitner et al. (2001), coincides with the maximum uplift and exhumation rates. Leitner et al. (2001) carried out the Southern Alps Passive Seismic Experiment (SAPSE) during 1995/1996 and compared their results of the analysis of 195 earthquakes during a 6-month period with the recordings of the NZNSN (1990-1997), and a local network at Lake Pukaki (1975-1983). The study by Leitner et al (2001), indicates that the Alpine Fault releases elastic strain seismically from the surface down to 10-12 km depth between Milford and the Hope Fault junction, with a low seismicity rate indicating the fault is currently locked, and there is the potential for large earthquakes. The results also showed the depth of the seismogenic zone beneath the Alps and adjacent to the Alpine Fault is possibly 3-4 km shallower than the rest of the South Island (12 ± 2 km), consistent with elevated isotherm models east of the Alpine Fault (e.g. Koons, 1987; Shi, 1996).

Modeling of elastic crustal deformation to the South Island plate velocity data by Beaven et al (1999) indicated a shallow locking depth of 5-8 km on the central Alpine Fault, above which the only plate motion occurring is by seismic slip. This finding, combined with the knowledge that there has been no significant earthquakes on the central Alpine Fault in the historical record (~170 years), is evidence that this section of the fault fails in large ($M_w$ ~7.6) earthquakes (Beaven et al., 1999). Paleoseismic studies support this with estimates of moment magnitudes in the range $M_w$ 7.6-7.9, based on 200-600 km rupture lengths (Sutherland et al., 2007) and strike slip offsets of ~8 m (Hull & Berryman, 1986). These studies determined there have been 3 major earthquakes of approximately this magnitude in the last ~800 years (Yetton et al., 1998), with best estimates of the last event occurring in 1717 (Wells et al., 1999). To fulfill a strike-slip rate of 27 mm yr$^{-1}$ the Alpine Fault must rupture with magnitudes of $M_w$ ~7.6, every 296 years. The large rupture lengths as estimated from paleoseismic studies indicate that the fault must rupture along most of its length from north Fiordland to the Hope Fault junction.

2.7 Summary

The Alpine Fault is a major, active oblique strike slip structure that fails in major ($M_w$ ~7.6) earthquakes every 300 or so years. Between these events small earthquakes occur on the fault plane. At these magnitudes, it is likely friction melts are being produced on the fault plane during the events, if the conditions are correct. Studying the already exhumed pseudotachylytes gives insight into not only the seismic history of the Alpine Fault, but also the overall rheological behaviour of this major plate boundary fault zone.
Chapter 3

Pseudotachylyte Field Relationships

3.1 Introduction

Fieldwork for this investigation was carried out at the two well-known and documented Alpine Fault localities: Harold Creek and Gaunt Creek. The localities and their spatial relation with one another and the Alpine Fault is shown in figure 2.02. The grid references (GR) and topographic names used in this section relate to the NZ Topographic Map 260 Series. Previous studies, as mentioned in section 2.43, have noted the presence of pseudotachylytes in all the fault rocks in the Alpine Fault zone, and these observations are confirmed by this study.

This chapter describes the pseudotachylyte field relationships observed from examination of outcrops and float rocks at Gaunt Creek and Harold Creek, with the aim of placing the pseudotachylytes analysed in chapters 4 and 5 within a field context. Three broad groups of host rock for pseudotachylyte are identified – footwall (Western Province-derived augen mylonites), hangingwall (Alpine Schist-derived mylonites) and fault core (cataclasites), and the pseudotachylytes within them are characterised. Field relationships of pseudotachylytes within Western province granitoid-derived mylonites are described from Harold Creek only, while pseudotachylytes hosted in cataclastic rocks in the fault zone are described only from Gaunt Creek. Pseudotachylytes hosted within Alpine schist-derived mylonites are described from both locations. The chapter begins with sections describing the field areas and the rocks that host pseudotachylyte.

3.2 Field Areas

3.21 Harold Creek

Harold Creek, a tributary stream to the Wanganui River, on the SW side of the river’s valley is located at approximately GR2315000E, 5779500N, near the township of Hari Hari (figure 2.02), and runs approximately NE for ~2.5 km before turning sharply NW as it exits its valley and enters an alluvial fan. The lower ~700 m were examined in this study, where scattered outcrops of various sizes are exposed by relatively recent rock falls. The Harold Creek section has been described previously by Sibson et al. (1979, 1981), Prior (1988), Wright (1998) and Toy (2007). Wright (1998) mapped the creek in detail, and recognised that the section contained varied mylonite lithologies, and that the Alpine Fault is exposed at the lower end of the creek just before the sharp NW turn (location A, figure 3.01), and mapped the fault trace to follow the creek bed upstream for ~1 km. A detailed map of the lower creek, the same portion that has been examined in this study, was mapped by Toy (2007) and is reproduced below in figure 3.01. The map shows different rocks of schist and granitoid protoliths are juxtaposed against one another as slices, with the trace of the Alpine Fault represented by the red dashed line. The granitoid-derived mylonites are composed of augen and banded mylonites, and the schist-derived mylonites are composed of dark coloured quartzofeldspathic mylonites, with some metabasite horizons. The intercalated slices of differently derived mylonite lithologies are thought to represent the
fault zone at depth as being an anastomosing network of high-strain zones, therefore incorporating both footwall and hangingwall rocks (Toy, 2007). The average foliation in the mylonites at Harold Creek dips at ~45° to the SE.

![Figure 3.01: Map of the lower section of Harold Creek after Toy (2007).](image)

3.22 Gaunt Creek

Gaunt Creek is a 4km long WNW flowing stream that feeds into the west bank of the Waitangi-taona River, approximately 4 km south of the SH6 road bridge. On the true left of lower Gaunt Creek, a ~700m long, ~100 m high slip outcrop exposes the recent trace of a thrust fault; this is known as the best exposure of the Alpine Fault in Westland (Cooper & Norris, 1994). The slip exposes a section through the fault zone, from footwall gravels in the NW, through the core of the fault and into the mylonites and Alpine Schist in the hangingwall to the SE.

The fault, dipping at ~40° southeast, is marked by a pale green, basal cataclasite zone overthrusting recent fluvio-glacial gravels. The fault surface shallows to the NW as a result of progressive shortening and overthrusting over an irregular talus fan surface (Cooper and Norris, 1994). The Gaunt Creek section has been mapped and the rocks within described by Cooper and Norris (1994), and later by Toy (2007), in which some of the zones were reclassified. Cooper and Norris (1994) described the footwall block at Gaunt Creek as being composed of schist-derived fluvio-glacial gravels overlain by mylonite-derived talus fan deposits and then mixed schist and mylonite-derived fluvial outwash gravels. Toy (2007) described the sequence immediately overlying the fault plane to the SE, as a ~15 m thick zone of mint green cataclasite overlain by a ≥15 m thick zone of olive green and black cataclasited ultramylonite. Overlying this, Toy (2007) mapped a ~100 m section of ultramylonites, followed by ~400 m of mylonites then ~500 m of protomylonite before transitioning into Alpine Schist. The mylonites were all found to be schist-derived, except for a
small sliver of augen mylonites found immediately above the cataclasite zone. Toy (2007) also mapped a near vertical, strike-slip fault zone, striking at ~070°, approximately 200 m from the trace of the fault plane. The average foliation of the mylonites at Gaunt Creek dips ~50° to the SE.

3.3 Host Rocks

3.31 Western Province granitoid-derived augen mylonites

Pseudotachylytes are hosted in augen mylonites at Harold Creek, and it is thought these mylonites are derived from footwall, Western Province rocks. The only mylonites in the hangingwall of the Alpine Fault zone of which there are strong lines of evidence suggesting derivation from a Western Province protolith are the augen and banded mylonites (Toy, 2007), exposed at Harold Creek. Banded mylonites are not known to, and were not observed to host pseudotachylytes, so not be focused on in this study and only augen mylonites will be the ‘footwall’ pseudotachylyte host rocks. Distinguishing between hangingwall- and footwall-derived protoliths for the mylonitic host rocks is important for the comparative aspect of this study; therefore, a justification for inferring that the augen mylonites are derived from Western Province granitoids is highly necessary.

Past workers (e.g. Reed, 1964; Sibson et al., 1979, 1981) inferred that the augen mylonites were derived from footwall granitoids of the Western Province, due to the presence of large porphyroclastic feldspar augens. Based on petrographic, geochemical and microstructural evidence Toy (2007) concluded the augen mylonites were derived from Western Province granitoids:

i) **Petrographic evidence**: the HREE epidote mineral allanite, although present in small quantities in the Alpine Schist, it is far more abundant in the augen mylonites.

ii) **Microstructural evidence**: schist-derived mylonites commonly contain large biotite fish, while the biotite in the augen mylonites is extensively recrystallised and comminuted to a finer grain size; and, amphiboles have undergone more grain-size reduction and retrogression in augen mylonites than in schist-derived mylonites. It is likely these reflect the different grain-sizes and microstructures of the protoliths rocks, and also different deformation conditions.

iii) **Geochemical evidence**: The augen mylonites from Harold Creek have similar Sr/Y ratios and aluminium saturation indices to Cretaceous Western province granitoids from the Petlab database. In addition, the augen mylonites show similar trends to the Petlab Western Province granitoids on Ce vs. \([\text{Al}_2\text{O}_3/(\text{Na}_2\text{O}+\text{K}_2\text{O}+\text{CaO})]\) and Zn vs. SiO\(_2\) plots. The results of the geochemical study by Toy (2007) should not be used as conclusive evidence for protoliths, however it has been suggested that titanium-in-quartz (TitaniQ) geothermometry (Wark & Watson, 2006) could be used to better constrain the protoliths of the mylonites (M. Palin pers. comm., 2009).

Abundant allanite was also documented by Toy (2007) to occur in some retrograde metabasite mylonites at Gaunt Creek, indicating that they are derived from Western Province rocks. Thin (~1 cm) metabasic layers do occur within augen mylonites, and may once have been intrusive dykes within the granitoid protoliths, and it is therefore possible that the abundant, large, retrogressed metabasite mylonite boulders in the creek bed at Harold Creek are derived from the Western Province rocks. However, although possible, I believe it is more likely the metabasic mylonite boulders in Harold Creek are from Alpine Schist-derived mylonite outcrops that do contain metabasite (Toy,
This is supported by the fact that metabasite layers, of variable, but significant thickness are known to occur in schist-derived mylonites, and are thought to represent metabasic lithologies of an Aspiring Terrane protolith. I will not use metabasites as footwall-derived host rocks in this study. However, as they do host particularly abundant and voluminous pseudotachylyte veins, they are of interest and are therefore included in this chapter as an additional host rock. The possibility of a contribution of the metabasite chemistry to pseudotachylyte veins in adjacent lithologies also cannot be ruled out.

**Field Appearance**

Augen mylonites at Harold Creek are a brown colour on weathered surfaces, and dark grey on fresh surfaces. They contain cream or white coloured feldspar porphyroclasts that stand out against the dark, fine grained matrix. They generally have a weak, spaced foliation, defined by mica or quartz and feldspar-rich layers. The augen range from ~1 mm up to 10 mm in length, however are generally about 2-5 mm, and are elongate in the direction of the foliation. Augen form 5 to 30 % of the rock mass. Some of the augen mylonites at Harold Creek were observed to be interlayered, on a 10 cm scale, with dark green, mica rich layers (metabasic layers – Toy, 2007), generally about 1-2 cm thick.

### 3.32 Alpine Schist-derived mylonites

Alpine-schist derived mylonites outcrop at Harold and Gaunt Creeks, and host ample pseudotachylyte veins. The schist-derived mylonites occur as 3 main lithologies, with metabasite and minor metachert horizons occurring in a predominantly quartzofeldspathic block. It is clear the ~1 km thick sequence of mylonites at Gaunt Creek are derived from the Alpine Schist as there is a gradual transition from them to the schist. Harold Creek mylonites are more difficult to distinguish, as the Western Province-derived and schist derived mylonites occur as intercalated slices in the hanging wall, as discussed before and shown in figure 3.01, however, similarities between Harold Creek and Gaunt Creek appearances can be used to distinguish them.

The quartzofeldspathic mylonites have quartz-feldspar-rich or mica rich endmembers (Toy, 2007). Mica rich mylonites tend to have a grey purple tinge, and a strong foliation, while the more quartz-feldspar rich members have a weaker foliation and a dark grey colour. Quartzofeldspathic mylonites contain well-developed shear bands cross cutting the foliation (Toy, 2007). Metabasic mylonites are often inter-bedded with quartzofeldspathic mylonites and the contacts are commonly gradational. The metabasic mylonites range from pale to dark green, and contain mm-thick layers of quartz, plagioclase and hornblende alternating with layers of biotite and hornblende, defining the foliation. The foliation varies in penetrative strength based on mica content, and in some cases, especially in quartz and feldspar-rich varieties, the foliation is defined by slightly elongate hornblende porphyroclasts. Garnets porphyroclasts are also commonly present in the metabasites. The metabasites are mostly strongly retrogressed to greenschist facies assemblages containing chlorite, epidote, calcite and opaques, and host particularly voluminous pseudotachylyte veins in boulders at Harold Creek. Minor metachert mylonites that are often associated with the metabasites are also present in schist-derived mylonites. Metacherts are pale cream and yellow colours, with a weak foliation, defined by garnet and mica layers (Toy, 2007). This study has not recognised pseudotachylytes in metacherts horizons, probably due to their relatively low volume as a lithology in the schist-derived mylonites.
3.33 Cataclastic host rocks at Gaunt Creek

Cataclastically deformed rocks within ~40 m of the Alpine Fault plane in the hangingwall, compose the ‘core’ of the fault. In the fault core, brittle deformation mechanisms have dominated producing abundant pseudotachylytes hosted in cataclasites. Pseudotachylytes were observed in the pale green cataclasites and the overlying partially cataclased ultramylonites at Gaunt Creek.

Figure 3.02: Photos of the pale green cataclasite zone at Gaunt Creek. a) Along strike photo of the Alpine fault with gravels in the hanging wall overlain by the pale green cataclasite unit. The Alpine Fault is the sharp contact that can be seen dipping towards the right (SE) of the photo. The grey gouge at the base of the pale green unit can be seen in the foreground of the photo. b) General nature of the pale green cataclasite showing its incoherent, pervasively fractured texture.
3.33.1 Pale green cataclasites

At Gaunt Creek ~15 m thick zone of pale green cataclasites occurs immediately above the fault plane at Gaunt Creek (figure 3.02a). At its base, on the fault plane is a ~1-5 cm thick clay gouge, above which rests a 50-70 m thick zone of indurated, grey-green fine-grained gouge (figure 3.02a). In outcrop, the thicker gouge unit is mostly homogeneous, with rare white quartz/feldspar clasts 1-3 cm in diameter, broken into chains that run for up to 10 cm. There are also some darker, wavy bands, with amorphous boundaries in the gouge layer.

The upper contact of the gouge is a distinct planar surface, upon which the ~15 m thick pale green cataclasite unit lies. The pale green cataclasite unit is composed of a cemented, highly sheared and incoherent material (figure 3.02b). Broken pieces of white quartz and feldspar are commonly observed standing out from the pale green material, with sizes ranging from 1-10 cm. These relatively rare clusters of quartz and feldspar fragments are the only clasts in the pale green material that are visible in hand specimen, as it is so fine-grained. The zone is pervasively hydrothermally altered and retrogressed, with clay alteration and chloritisation of minerals throughout the cataclasite. Silica and calcite cementation is also pervasive, however, some areas are more highly cemented (or less fractured), than others. The zone also contains white coloured pods or lozenges, ranging in length from 10cm to 1.5 m, composed of quartz and feldspar. The quartz-feldspar pods are generally more indurated than the surrounding cataclasite and therefore standout and form pillars and ridges in the deeply eroded zone.

The zone contains abundant clay-rich shears, and thin discontinuous veins, and is highly fractured. Overall this pervasively altered and cemented fine grained material is fragmented into ~1-20 mm blocky pieces that, depending on the level of cementation, tend to crumble when crushed in the hand, releasing dusty (if it is dry), fine-grained clays. The pale green cataclasite, especially in some places looks to have possibly preserved a mylonitic fabric.

Microstructural observation by Toy (2007) reveals that these cataclasites are in fact protocataclasites, as no fine-grained matrix is obvious. Detailed microstructural observations in chapter 5 of this thesis support and add to this. The contact between the mint green cataclasite and the overlying cataclased ultramylonite is defined by an undulating shear surface, which was observed to contain pseudotachylyte.

3.33.2 Cataclased ultramylonites

Approximately ≥15 m of dark green and black, coherent cataclased ultramylonite overlies the pale green cataclasite at Gaunt Creek. This was originally interpreted by Norris & Cooper (1994) to be ultracataclasite of obvious schist origin. Toy (2007) reclassified this material as cataclased ultramylonite.

In outcrop, this material exhibits a strange and complex inter-fingering of the black and dark green layers (figure 3.11a). The material is much more cohesive than the underlying cataclasites and shows an obvious cleavage which is offset by regular, cm-spaced microfaults and fractures, resulting in a blocky outcrop appearance (figure 3.11a). The cleavage fabric is ultramylonitic, with many mm-scale laminations of light and dark minerals. Toy (2007) found that the colour difference between the interleaving materials is due to a higher proportion of opaque minerals and titanite in the black compared to the green cataclased ultramylonites. Shear surfaces tend to separate the green from the black material, and these surfaces host gouge, cataclasite and/or pseudotachylyte veins.
3.4 Pseudotachylytes

The pseudotachylytes were identified in the field as any relatively planar, dark coloured, flinty vein. The veins were further classified into fault veins or ‘generation surfaces’, and if present, injection veins, both of which are typical structures of pseudotachylytes as described in Chapter 1. Determining a melt origin of the flinty veins is not possible in the field, however any veins, whether they be pseudotachylyte or ultracataclasite, described below have been sampled and a melt origin has been confirmed from microscopic studies, which are described in chapters 4 and 5.

3.41 Pseudotachylytes hosted in augen mylonites

Pseudotachylytes were observed in augen mylonites at Harold Creek in large quantities in float boulders in the creek bed, and more rarely in-situ. This observation has been made previously by other authors (Sibson et al., 1979, 1981; Toy, 2007), and apparently there were once voluminous, in-situ pseudotachylytes visible in a part of the creek bed, that have since been obscured (V. Toy, pers. comm., 2009). With a scarcity of good outcrop, and the numerous float boulders present in the creek bed, the descriptions of the pseudotachylytes hosted in augen mylonites are therefore largely derived from these boulders.

Pseudotachylytes fault veins and injection veins were observed in the in-situ augen mylonites, but were very difficult to recognise in the mossy, dark coloured outcrop, and the full nature of the veins was in some cases only revealed after slicing the rocks at a later date. Two outcrops of pseudotachylyte veins hosted in augen mylonite were found, and a sketch of one of the outcrops observed approximately at location B in figure 3.01, is shown in figure 3.03. Although pseudotachylytes were not well exposed in outcrop, the abundance of veins in the small area of the outcrop in figure 3.03, suggests that veins are possibly quite common in augen mylonites.

Figure 3.03: Sketch of in-situ pseudotachylyte veins hosted in augen mylonite at Harold Creek. Parallel generating surfaces that are sub-parallel to the mylonitic foliation are linked by injection structures. The dashed lines represent the mylonitic foliation. The paired generating surfaces at the top of the image were continuous for ~4 m along their strike.
Fault veins are generally 1-5 mm thick, often varying along strike, thinning to <1 mm in some parts, making them difficult to trace, and then thickening again to where the trace of the vein can be re-established. Veins are either parallel to, or cross-cut the foliation at low-moderate angles, and are generally continuous along strike for 1-5 m. Injection veins are injecting into the mylonite in a highly erratic, irregular manner. The injection veins range from 2mm to 30 mm thick and from 5mm to 300mm long, often cross-cutting the mylonitic foliation at high angles and tapering from the base of the injection to the tip (Figure 3.03, 3.04). Fault veins were generally observed to be planar on one side, and more irregular on the side that has injection structures (figure 3.03).

![Figure 3.04](image)

**Figure 3.04:** Pseudotachylyte veins hosted in an augen mylonite boulder at Harold Creek. Parallel generating surfaces can be seen diagonally running from the bottom right to the top left of the photo across the boulder. The paired generating surfaces cross cut the foliation and are linked by a voluminous injection structure. A third pseudotachylyte vein cross cuts these generating surfaces, but is difficult to see in the photo, it is approximately parallel to the string of the compass. The compass, less the string, is 10 cm long.

Most of the veins observed were paired with parallel fault veins which bound faulted or cataclased mylonite, and are linked by injection structures and reservoir zones (figure 3.03). The network of pseudotachylyte between the paired fault veins can become quite complex, especially at mm-cm scale, with clasts of mylonite or feldspar set in a pseudotachylyte matrix (figure 3.03). The separation between the fault veins ranges from 2-20 cm, but is generally observed to be ~ 10cm. A particularly well-exposed paired vein network was observed in a float boulder of augen mylonite in the creek bed (figure 3.04), with an injection vein that crosscuts older veins, which suggests multiple melt generation events. Many boulders in the creek bed, and a boulder of augen mylonite from Harold Creek that is
displayed in the foyer of the Geology Department at the University of Otago, show large volumes of pseudotachylyte within reservoir zones that are within the fractured mylonite between paired generating surfaces.

The general orientation of these pseudotachylyte veins, as measured in outcrop, is a high angle (40-70°) dip to the SE, similar to the local mylonitic foliation. Offsets of fault veins were not observed, mostly due to poor continuity and veins being subparallel to the foliation.

3.42 Pseudotachylytes hosted in Alpine Schist-derived mylonites

The structural relationships of pseudotachylytes hosted in schist-derived mylonites from Gaunt Creek and other exposures along the fault, have been described in detail and characterised by Toy (2007), and her findings form the basis of this section. Outcrop and thin section observations are included in this section to gain an idea of the structural relationships involved. Either foliation parallel or foliation oblique pseudotachylyte veins occur in the Alpine Schist-derived mylonites. Foliation oblique veins crosscut the foliation at low angles, and no oblique veins were observed to crosscut the foliation at high angles. Thin ultracataclasite layers are observed in the schist-derived mylonites. These are commonly parallel to the mylonitic foliation, and were not observed to crosscut any of the pseudotachylyte veins.

3.42.1 Pseudotachylyte veins parallel to mylonitic foliation

Foliation-parallel veins can be traced for up to 10 m across an outcrop, and are commonly very thin (1-2 mm), but can thicken in places up to 10 mm where they form lenses that taper at each end (figure 3.05). At outcrop scale it is difficult to see injection veins, but petrographic observations in chapter 4 reveal they are present. Adjacent to the veins there is often fractured and faulted blocks of mylonite, with thin pseudotachylyte-filled slip surfaces that either join or are parallel to the main fault vein. Space created by fracture of the mylonite can act as reservoir zones for pseudotachylyte.

Flow fabrics observed in some pseudotachylyte vein in thin section indicate top the NW shear, consistent with shear sense indicators at Gaunt Creek and Alpine Fault kinematics (Toy, 2007). The orientations of foliation parallel veins are the same as the mylonitic foliation at Gaunt Creek.

Figure 3.05: Outcrop sketch of foliation parallel pseudotachylyte veins hosted with Alpine Schist-derived mylonites. Dashed lines represent mylonitic foliation.
3.42.2 Pseudotachylyte veins oblique to the mylonitic foliation

Foliation-oblique pseudotachylyte veins cross cut the mylonitic foliation at angles generally less than 30°. The oblique veins tend to be thicker than the parallel veins, commonly at around 1 cm thick, and can get up to 2 cm. Often the foliation on one side of the fault vein converges toward the vein, which is parallel to the foliation on the other side (figure 3.06a). This relationship, like that depicted in figure 3.06a, commonly occurs on the contact between quartzofeldspathic and metabasic horizons in the mylonites at Gaunt Creek. At one Gaunt Creek outcrop, it was observed that a ~1.5 cm thick vein separated blocks of quartzofeldspathic mylonite, with foliation on either side having a different orientation (figure 3.06b). Injection veins are more commonly associated with oblique veins than parallel veins, and they can be seen easily in outcrop to inject along foliation planes as shown in figure 3.06a and b.

![Figure 3.06: Outcrop sketches of foliation parallel pseudotachylytes. Note injection structures along foliation planes.](image)

3.43 Pseudotachylytes hosted in metabasite mylonites

Particularly voluminous pseudotachylyte veins were observed to be hosted in numerous metabasic mylonite boulders at Harold Creek. The metabasic lithologies were not observed in situ, and therefore the following descriptions are drawn entirely from the boulders in the creek bed. Fault veins vary in width from 1-2 mm to 20 mm (figure 3.07), and form very complex networks with one another, and the injection veins (figure 3.07). Injection veins vary from 1-10 mm thick and 1-15 cm long (figure 3.07), and often link fault veins. Injection veins are commonly observed to be at moderate to high angles to the fault vein.

Fault veins have varied relationships with the surrounding foliation orientations, exhibiting parallel and oblique arrangements. Often a vein will run parallel, then jog across the foliation for distances up to 10 cm, and then continue to run parallel to the mylonitic foliation. Paired fault veins are also commonly observed, and may not necessarily be parallel to one another (figure 3.07). A single fault vein may bifurcate at a point, and rejoin some distance later. Large, voluminous reservoir zones are observed in the metabasite hosts, with brecciated, and rounded pieces of mylonite set in a pseudotachylyte matrix (figure 3.07). Cataclastic textures are commonly associated with the
veins, with pseudotachylyte in some veins transitional into cataclasites. Cataclastic textures and microfaulting also disrupt some of the veins, suggesting a later stage of cataclasis. However, other veins in the same boulder may not be cataclastically deformed which suggests multiple stages of pseudotachylyte production.

![Figure 3.07: Hand specimen photo of a pseudotachylyte bearing metabasite boulder from Harold Creek. A voluminous network of pseudotachylyte can be seen infiltrating the rock mass. The compass is 10cm long.](Image)

### 3.44 Pseudotachylytes hosted in cataclastic rock at Gaunt Creek

Pseudotachylytes were observed to be hosted in pale green cataclasites and dark green and black cataclased ultramylonites within ~ 40 m of the Alpine Fault at Gaunt Creek.

#### 3.44.1 Pseudotachylytes hosted in pale green cataclasites

Pseudotachylyte veins in this host rock are very difficult to recognise in the field, as they are highly fractured and discontinuous, and many of them are only slightly darker than the surrounding cataclasite. In addition, the veins and the cataclasite are very incohesive, therefore making sampling the veins very difficult. Only a few veins were traced, difficultly, for up to a maximum of 2 m, while most of the veins could only be traced for ~10–30 cm. Some areas of the pale green cataclasites zone are relatively well cemented and coherent, therefore provide good, relatively continuous exposures of pseudotachylyte veins.

The veins, when able to be differentiated, are observed to be grey-blue or brown in colour, and have a fine-grained, aphanitic appearance (figure 3.08a, b). The veins are on average ~5 mm thick, and generally do not exceed thicknesses of ~10 mm. The pseudotachylyte veins have very irregular orientations, often curving around inhomogeneities or abruptly splitting into other veins. Some veins have alternating colour bands parallel to the vein walls. Distinguishing between fault veins and injection veins in some cases is difficult due to limited continuity of the
recognised vein. Injection veins that can be linked to a fault vein are generally found to be 1-40 cm long, and have approximately the same width of the fault veins at their base, and taper out along their length. The most continuous veins were commonly observed on the contact surfaces between the cataclasite and the numerous quartz-feldspar pods (figure 3.09), and injection structures of these veins were directed into the pods (figure 3.09). Fragmented pseudotachylyte veins in the cataclasite were often observed to be within the vicinity of quartz and feldspar rich areas in the zone.

Figure 3.08: Photos of pseudotachylyte veins hosted in the pale green cataclasite zone. a) Brown pseudotachylyte vein that is has been fractured and offset by a gouge filled shear. Note the colour banding in the vein, and its close association with a quartz rich area. b) Blue pseudotachylyte vein.
Networking veins are also common, with paired generating surfaces linked by injection structures. Blue and brown veins do not tend to cut one another, and in one case a brown and blue vein intersect, but it cannot be deduced which is the later. The orientation of the veins is variable ranging in strike from approximately 040° to 110°, with a dip range towards the south of approximately 45° to 85°. An outcrop was observed close to the boundary between the pale green cataclasite and the overlying cataclasized ultramylonite. This outcrop was well-cemented, contained pseudotachylyte veins that were offset by later conjugate shears, and had textures resembling a mylonitic foliation.

**Figure 3.09:** Sketch of quartz lozenge in the pale green cataclasite showing pseudotachylyte generating surfaces on either side, with injection structures directed to towards the centre of the pod. Blue tinged pseudotachylytes cross cut the pod along through going shears.

**Figure 3.10:** Photograph of the flame-like injection structure near the top of the pale green cataclasite unit. The center of the structure is green in colour while the rim is dark blue.
A particularly interesting structure was observed at the very top of the pale green cataclasite. The structure is photographed in figure 3.10, and it can be seen that the flame like, erratic has a khaki green core and is surrounded by a blue-tinged, 1-2 mm wide rim, that is very similar in colour to the blue tinged veins observed in other parts of the cataclasite. The flame-like structure is connected to a through-going vein, and has the appearance of an injection vein. The vein, which is a blue-tinged pseudotachylyte vein, loses continuity and is obscured after ~10 cm either side of the injection vein.

3.44.2 Pseudotachylytes hosted in cataclasised ultramylonites

Pseudotachylyte veins in this host rock have a grey-blue colour, much like those in the underlying cataclasite, and exhibit a highly aphanitic appearance. The veins occur along most of the shear surfaces between the dark green and black material (figure 3.11a). Fault veins thicknesses range from 1-10 mm, with most of them being very thin (1-2 mm). Injection veins are commonly observed protruding from the fault veins at high angles into the surrounding ultramylonite, and may be up to 15 mm wide at their base and up to 20 cm long (figure 3.11b).

The veins are in some cases offset by the numerous high angle microfaults that disrupt the cleavage of the ultramylonite. These faults may be filled with gouge or cataclasite, and can be decorated by pseudotachylyte as documented by Toy (2007). The photographs of figure 3.11 show the typical relationships and pseudotachylyte structures apparent in the cataclasised ultramylonite.
3.5 Summary

Pseudotachylytes were observed to occur in 4 different host rocks at Harold Creek and Gaunt Creek: Western Province granitoid-derived augen mylonites, schist-derived mylonites, cataclasites, and metabasic mylonites of an unknown protolith. The main relationships between pseudotachylytes and their host rocks can be drawn from the field component of this study:

1. Relatively thin (~1-10 mm) and continuous fault veins occur in Alpine Schist-derived mylonites. These veins are either parallel or oblique to the foliation and generally do not have large injection structures. Injection veins are generally controlled by planes of weakness along foliation or along mylonitic shear surfaces.

2. Relatively voluminous and numerous pseudotachylytes occur in less intact cataclasites, metabasites and in the augen mylonites, with large amounts of pseudotachylyte in injection structures and reservoir zones. Paired generating surfaces may exist, linked by large injection structures.

3. In the pale green cataclasite zone the most continuous pseudotachylyte veins are commonly observed on the surfaces of quartz-feldspar pods, and around more cemented parts of the cataclasite. Away from these areas, veins are largely broken apart into fragments, but may still be closely associated with areas rich in quartz and feldspar.
Chapter 4

Petrography and Geochemistry

4.1 Introduction

Petrographic and geochemical investigations of pseudotachylyte veins allow for the further understanding of processes involved during the frictional fusion of the host rock to form a melt. Optical and electron microscopic investigations have allowed characterisation of the Alpine Fault pseudotachylytes. The material that will be covered in this chapter will be:

- Host rock mineralogy is described, and metamorphism and deformation are discussed
- Petrographic observations of the mineralogy and the textures that indicate a melt origin of the veins are summarised, for the three different host rock settings.
- The methodology and results of the geochemical analyses carried out on the pseudotachylyte veins and their host rocks are presented in this chapter, with the aim of further establishing evidence of a melt origin of the Alpine Fault pseudotachylytes, and to attain matrix compositions for use in later calculations of their physical properties.

Numerous mylonite samples and pseudotachylyte veins hosted in mylonite collected by V. Toy over the period 2004-2007 were used in this study, and some of these are used for analysis in this chapter and in chapter 5.

4.2 Petrographic Methods

Optical microscopy and electron microscopy were both used to obtain petrographic information. Most of the mineralogical descriptions are based on transmitted light optical microscopy observation of thin sections. Where possible, thin sections were cut perpendicular to the veins.

Back-Scatter Electron (BSE) imaging was carried out on the JEOL JXA-8600 Superprobe in the Otago University Geology Department. Samples were prepared as standard petrographic thin sections, and polished on a rotary cloth lap using a 1 µm diamond paste. The thin sections were then carbon coated and microprobed using a beam current of \(-2 \times 10^{-11} \text{ A}\), and an acceleration voltage of 15 keV. The highest magnification obtained with these samples was approximately 500X, above which the images become progressively lower in quality. BSE imaging is particularly useful for distinguishing between quartz and feldspar, especially crystals that are fine grained and cannot be resolved with an optical microscope. The image produced displays a different grey shade for each mineral, the shade being a function of the amount of backscattered electrons, which is in turn a function of atomic number. The order of darkest to lightest grey shades for minerals in the samples imaged is quartz-feldspar-chlorite-micas-amphiboles-garnets-opaques. Not all of these are present in most of the images taken, they are labeled wherever they are presented in this thesis.
Energy-dispersive spectrometry (EDS) spot analysis was also used to identify the mineralogy and composition of some materials for petrographic purposes, this method will be described further in section 4.6.

Pseudotachylyte veins hosted in the cataclasites produced poor quality polished thin sections. This was due to the very high proportion of soft clay minerals in the samples. The poor quality of polishing in turn affected the quality of many of the images. However, the poor quality of the polish did in some cases highlight the mineralogy and texture of a certain area, therefore this was also used to differentiate minerals or bulk mineralogy of a given area. The fine-grained and highly altered nature of some of the pseudoatchylyte veins and also the pale green cataclastic host rocks made it difficult to resolve the their mineralogies with both petrographic methods.

4.3 Host Rock Petrography

Petrographic descriptions of the host rocks are briefly summarised below, in order to characterise their mineralogy for understanding melting relationships during pseudotachylyte paragenesis. The mineralogy and metamorphism of the Alpine Fault mylonite lithologies has been described by many previous workers (e.g. Reed, 1964; Sibson et al, 1979, 1981; Rattenbury, 1987a; Prior, 1988; Cooper & Norris, 1994; Read, 1994; Grapes, 1995; McClintock & Cooper, 2003; Toy, 2007), and is therefore generally well known. Petrographic descriptions of the cataclasite zone rocks have been carried out by Warr and Cox (2001) and Toy (2007). The following descriptions of the mylonites include some microstructural observations, while the microstructures in the cataclasites are described in chapter 5.

4.3.1 Augen mylonites

4.3.1.1 Mineralogy and metamorphism

The augen mylonites have a porphyroclastic morphology, with large feldspar porphyroclasts in a very fine-grained matrix. A typical section of augen mylonite has the mineral assemblage quartz-feldspar-chlorite-epidotitannite-carbonate-biotite=hornblende=muscovite. Modal abundances are difficult to estimate due to the fine-grained matrix, however it is obvious there is a relatively high proportion of fine grained phyllosilicate minerals present in the augen mylonite matrix. The augen are generally microfractured or well-rounded plagioclase porphyroclasts that can be up to 5 mm in diameter and can comprise up to 50% of the rock. The microfractures are commonly infilled with calcite (figure 4.01b). The matrix is largely fine grained and composed of sericite, chlorite, plagioclase, quartz, and calcite with high mica proportions defining a foliation that anastomoses around porphyroclasts. The hornblendes are pseudomorphed or completely replaced by chlorite and chlorite-calcite-stilpnomelane (figure 4.01a), and rare biotites are being replaced by chlorite. Quartz and calcite veins cross cut the foliation, often forming networks that are joined by discrete shear surfaces. Epidotes are present as allanite, often with clinozoisite rims. It should be noted that feldspars are only weakly altered in the augen mylonites, except when closely (~1-2 m) adjacent to fracture zones and large pseudotachylyte veins, where they are largely saussuritised and often also variably cataclased.

The strongly retrogressed nature of biotite and hornblende, replaced by chlorite, stilpnomelane and calcite are indicative of lower greenschist facies conditions. Most of the feldspar augens have been found to be albite endmember-rich (Toy, 2007). Albite, epidote and chlorite are all index minerals of the greenschist facies. Therefore the augen mylonites have mineral assemblages indicative of equilibration at lower greenschist facies conditions.
4.31.2 Microstructures and deformation

Feldspar porphyroclasts are microfractured and fine grained quartz commonly infills the cracks and occupies pressure shadows created by the porphyroclasts. Quartz is generally fine grained, and slightly elongate in the direction of foliation. In rarely recognised quartz aggregates, the grained boundaries are generally seriate in nature. Larger quartz grains show undulose extinction.

Toy (2007) summarised the deformation of the augen mylonites, concluding that the rocks deformed predominantly by grainsize sensitive creep and pressure solution processes, while feldspar porphyroclasts underwent brittle microfracturing.

4.32 Alpine schist-derived mylonites

4.32.1 Mineralogy and metamorphism

The typical mineral assemblage of a quartzofeldspathic Alpine Schist-derived mylonite is quartz-plagioclase-biotite-muscovite with lesser garnet-chlorite-epidote-calcite-titanite-graphite. Mineral abundances estimated from thin section have found quartz and plagioclase form most of the rock (~60 %), followed by brown biotite and muscovite forming ~35 %, with the rest of the phases composing the remaining ~5 %. A dominant S-C’ fabric (Bérthe et al., 1979) is obvious in thin section (figure 4.02), with the C’ planes containing fine grained micas (mostly biotite), feldspar and quartz.

The bulk of the plagioclase feldspars in the schist-derived mylonites have been found to be of oligoclase composition (e.g. Cooper, 1980; Read, 1994). The presence of Ti-rich brown biotite in the mylonites has also been noted (Toy, 2007). These are known to be index minerals of the amphibolite facies, indicating the schist-derived mylonites largely equilibrated under amphibolite facies conditions. The presence of chlorite and epidote in some sections indicates that the mylonites underwent some retrogression as greenschist facies conditions were experienced.

Figure 4.01: Photomicrographs of augen mylonite sample OU77737. a) View of feldspar porphyroclasts with mica rich layers anastomosing around the clasts b) Microfaulted feldspar porphyroclast with fractures in filled with calcite. Note fine grained nature of the matrix.
The Alpine Schist-derived mylonites have undergone less retrogression and alteration than augen mylonites, and the cataclasites.

![Figure 4.02: Typical thin section view of Alpine Schist-derived mylonite. C' shear surfaces cross-cut the image from bottom right to top left composed of fine-grained phyllosilicates and quartz. The foliation is horizontal and defined by quartz and felspar rich layers alternating with mica rich layers.](image)

### 4.32.2 Microstructures and deformation

As already mentioned a S-C' fabric is obvious in thin section (figure 4.02), with alternating large biotite grains and quartz-feldspar rich layers defining the S foliation, while aggregates of fine grained biotite and probably fine quartz and feldspar are contained in the C' plane. Toy (2007) noted quartz grains have well-developed grain boundaries that may be interlobate and commonly bulge into pinning structures, and have only weak undulose extinction and no deformation bands. A grain shape fabric is also obvious within quartz rich bands.

Toy (2007) proposed that dynamic recrystallisation of quartz dominantly occurred by dislocation creep, and that there was intense, pervasive development of a shear fabric.

### 4.33 Cataclastic host rocks

#### 4.33.1 Mineralogy and metamorphism

**Pale green cataclasites**

The pale green cataclasites and cataclasized dark green and black ultramylonites in the fault core have essentially the same mineral assemblage, although the modal abundances vary between the two rocks and within them. Accurate mineral identification, and therefore modal abundance estimations are difficult in these rocks due to the fine-grain size of many of the minerals, however, some broad observations and estimations are possible. Photomicrographs are not used here to illustrate the descriptions due to the very fine grain size and highly altered nature of the cataclaseite.
The mineral assemblage that can be resolved from thin section observation is quartz-feldspar-chlorite-calcite-titanite-clinozoisite plus rare but optically resolvable biotite and muscovite. Feldspar porphyroclasts are heavily saussuritised and are also altered to chlorite and clay minerals. The matrix is cemented mostly by abundant fine-grained by silica and also calcite. The cementing texture is shown very well in the BSE image of figure 4.03a. Chlorite and epidote are present throughout the matrix also. Clasts containing a remnant mylonitic fabric are observed in some sections, although they are highly altered to fine-grained chlorite and clays. Limonite staining in very fine-grained matrix areas, and along the foliation of the existing mylonite clasts is common. The large, white coloured pods were found by thin section observation and BSE imaging to be composed of predominantly feldspar cemented by quartz (figure 4.03b).

Figure 4.03: a) BSE image of the pale green cataclasite showing the highly cemented texture. b) BSE image of a quartz-feldspar pod, showing quartz cementing feldspar.
Discussion and Interpretation

One relatively intact, unaltered plagioclase feldspar found in the pale green cataclasites was observed to have zoned reverse extinction. This may indicate the feldspar crystallised from a melt, and is not a result of metamorphic crystallisation. If this is the case, it is likely that at least part of the pale green cataclasite unit is derived from a Western Province granitoid-derived mylonite. This would not be surprising given that a thin slice of augen ultramylonite mylonite was mapped just overlying the pale green cataclasite by Toy (2007), and that apparently in-situ granitoid material was once visible in the creek bed at Gaunt Creek (R. Norris pers. comm., 2008).

By centrifuging well-disaggregated samples of the pale green cataclasites from Gaunt Creek, Warr and Cox (2001) found that they are intact composed of ~20-60wt % of clay-sized particles (<2 µm in diameter). X-ray Diffraction (XRD) analyses of these samples by Warr and Cox (2001) further revealed the clay-sized particles are composed of (in order of approximate decreasing abundance), chlorite, illite-muscovite, biotite, smectite and vermiculite. The clay minerals illite, smectite and vermiculite are probably derived from the alteration of plagioclase feldspar.

The presence of abundant chlorite and also some epidote indicates metamorphic equilibration at greenschist facies conditions. Furthermore, Warr and Cox (2001) found that pervasive solid-state transformation of biotite, dissolution and neocrystallisation formed Mg-chlorite at sub-greenschist facies conditions (<320°C), and that the clay minerals grew at lower temps (<120°C). The presence of abundant cementing calcite further indicates metamorphic equilibration at greenschist facies conditions, as carbonate is known to be scarce in Alpine Schist amphibolites (Cooper, 1970). If these cataclasites are derived from an Alpine Schist protolith, they have experienced a much larger amount of retrogression and alteration than the weakly retrogressed mylonites that lie above them.

Cataclased ultramylonites

As mentioned above the green and black cataclased ultramylonites and the pale green cataclasites that lie directly below them have essentially the same mineral assemblages. The main difference between them is that the pale green cataclasites contain much more quartz and chlorite. The difference between the green and black colours of the cataclased ultramylonite is a result of it having a higher proportion of opaques and titanite. This cataclasite is also pervasively cemented by quartz and calcite. Plagioclase feldspars are still altered to saussurite and clay minerals, however the effect has not been so intense. Felted masses of sericite are more common in the cataclased ultramylonites and the matrix generally has a more optically distinguishable appearance.

The cataclased ultramylonites have probably experienced a broadly similar metamorphic history to the pale green cataclasites, with an assemblage indicating greenschist facies equilibration. The pale green cataclasites have also experienced much more alteration and cementation by infiltrating fluids.

4.33.2 Microstructures and deformation

Pale green cataclasites

The pale green cataclasites are mostly composed of porphyroclasts of internally microfractured plagioclase feldspar and quartz. The nature of the cataclasites, however, is very variable from one area to the next and they present
a wide range of grainsizes. In most areas it is dominated by large clasts (>1 mm) composed of randomly oriented, cemented polymineralic material separated by fractures filled with very fine cements and micas, and with no visible fine-grained matrix. In other areas, the cataclasite may be of a porphyroclastic texture with larger porphyroclasts of altered feldspar and quartz set in a finer grained matrix or cement composed of predominantly quartz and mica, with thin shear surfaces anastomosing around the porphyroclasts and into finer grained areas, this type of texture generally occurs within the vicinity of strongly localised shearing adjacent to pseudotachylyte or ultracataclasite veins. There are also areas where clasts of chloritised and altered ultramylonite are dominant, and strongly resemble the texture of the overlying cataclasite ultramylonite unit. Some large clasts may contain within them quartz-cemented fragments of earlier cataclasite material. Quartz and calcite cementing is pervasive through the cataclasite, and some of the quartz grains exhibit undulose extinction and some aggregates of fine quartz grains have interlobate grain boundaries and a slightly elongate shape. Quartz veins also exhibit undulose extinction.

Discussion and interpretations

Accurate grain size and volume fraction analysis is difficult in this cataclasite due to its pervasively cemented nature and multiple episodes of fracturing, but estimates from thin section are adequate. The microstructural observations of the pale green cataclasites have recognised no consistent fine-grained matrix that is not quartz or calcite cement. The bulk of the unit is therefore not a cataclasite at all, but a protocataclasite, or possibly a fine crush breccia or crush microbreccia. This was also found to be the case by Toy (2007), in which she classified the pale green unit as a protocataclasite.

As the quartz and calcite cement is so dominant and appears to be the ‘matrix’ containing the quartz and feldspar porphyroclasts, the fluids that deposited them may have contributed to the fragmentation, as proposed by Toy (2007). If this was the case, the fluid pressures must have exceeded lithostatic pressure and promoted the brittle failure, or further brittle fragmentation of the rock mass. The quartz in the cement appears to have been deformed by dislocation creep, and it has been suggested by Toy (2007) that transiently high pore fluid pressures promoted failure by brittle mechanisms in temperature and stress conditions that would usually result in dislocation creep of the quartz lattice. The transiently high fluid pressures may have been attained during seismic rupturing, and then fallen again, following a typical ‘fault valve’ cycle (Sibson, 1988). Warr and Cox (2001) proposed that transiently high fluid pressures may be able to be attained due to clay mineral transformations that reduce permeability and build up the fluid pressure that reaches its peak during seismic slip.

Later displacement of the cemented mass was probably along the localised, shears and fractures that separate the large internally cemented fragments. Except for on discrete shears and in, or immediately adjacent to ultracataclasite or pseudotachylyte veins, the bulk of the pale green protocataclasite did not deform by traditional cataclastic mechanisms of failure, attrition and wear. They were probably coseismically fragmented at depth (~5 km) by transiently enhanced fluid pressures, and then cemented by these silica rich fluids. Re-fragmentation probably occurred in these relatively strong, cemented masses, during multiple cycles of enhanced fluid pressures and cementation. Brittle failure on discrete shears that occur as anastomosing networks throughout the pale green unit later produced pseudotachylyte and ultracataclasite.
Based on mineralogical and microstructural characteristics of the clays in the pale green cataclasite, Warr and Cox (2001) defined three successive modes of deformation and alteration that formed the cataclasite; anhydrous cataclasis and frictional melting in the fault zone at 6-12 km depth, followed by pervasive hydrous chloritisation of mafic constituents resulting from coseismic meteoric fluid infiltration at sub-greenschist facies conditions, further followed by reaction weakening in the upper crust (2-4 km) due to the growth of swelling smectite and vermiculite in the gouge matrix at relatively low temperatures. Some of my observations support Warr and Cox’s interpretations and possibly allude to differences that will come to light later in this thesis.

Green and black cataclased ultramylonites

The green and black ultramylonites exhibit a strong disjunctive cleavage in thin section, with discontinuous layers often offset by conjugate microfaults. The microfaults contain foliated gauge and calcite and quartz cement. The foliation, defined by layers of fine-grained titanite and opaques alternating with quartz-muscovite-sericite-feldspar layers, anastomoses around internally fractured feldspar porphyroclasts <3 mm in diameter. Rare feldspar porphyroclasts exhibit a weak, reverse zoned extinction. Strongly anastomosing, thin shear surfaces may be continuous with the mica-rich layers. In some areas, especially near ultracataclasite and pseudotachylyte veins, a cataclastic texture may be developed with very high clast to matrix ratios.

Discussion and interpretations

This unit has, from field observations, been considered to be an ultracataclasite (e.g. Cooper & Norris, 1994). Microstructural observation reveals that the green and black unit is not pervasively cataclased at all, and exhibits a relatively intact fabric that has developed a disjunctive cleavage. Toy (2007) made the same observation and renamed the unit a partially cataclased ultramylonite, and concluded that the dominant deformation mechanism in the ultramylonites was pressure solution facilitated by grain boundary sliding in the mica-rich layers. The microstructural observations made of the ultramylonite at Gaunt Creek in this study, support this conclusion.

4.4 Pseudotachylyte Petrography

Two materials are recognised as components of the pseudotachylyte veins observed. These are:

1. Fragments of host rock and minerals that were not fully incorporated into the melt, but may have been partially melted around their edges. These fragments did not crystallise from the melt phase, as indicated by their remnant textures and are referred to here as ‘lithic clasts’ or just ‘clasts’.

2. A very fine-grained matrix representing the solidified equivalent of the pseudotachylyte melt. The nature and mineralogy of this matrix were irresolvable by the methods used in this study.

The pseudotachylyte veins observed in this study all possess general similarities regarding their clast mineralogy and morphology, and matrices. To avoid repetition, generalisations on the nature and mineralogy of the clasts and the nature of the matrix can be made for the three differently hosted pseudotachylytes. These generalisations are summarised in the following subsection, followed by petrographic descriptions and the features that are characteristic for each specific set of pseudotachylyte veins.
4.41 Features common to all pseudotachylyte veins observed

Petrographic observations in this study have recognised features common to all the pseudotachylyte veins collected from the Alpine Fault zone host rocks. These are:

- Quartz and to a lesser extent feldspar comprise most of mono-crystalline clasts in the pseudotachylyte veins. Composite fragments are also observed commonly composed of a quartz-feldspar aggregate from the host rock, with other accessory minerals. Much lesser amounts of mica are present as clasts.
- Quartz clasts are generally larger and intact, while feldspar grains are smaller, more corroded, and fractured and often infiltrated by the matrix material.
- The smaller clasts are generally more rounded and circular than larger clasts, which are commonly angular and contain embayments into which matrix intrudes.
- Clasts commonly have blurry margins visible in under an optical microscope.
- The veins commonly contain different coloured banding that are often parallel or sub-parallel to the vein walls, that reflects may reflect different matrix compositions.
- The matrix of the veins typically has dirty yellow-brown or dark brown or dark grey colour in plane-polarised light, with no optically resolvable crystals. These matrix colours vary between samples of the same host rock type, and also within the same vein. In cross-polarised light the matrices are dark brown or black, and have a mottled appearance in thin section.
- A few small patches of isotropic material have been observed in some of the veins, however, confirmation that these are glass is not possible without Transmission Electron Microscopy (TEM) analysis.
- The veins form either: sharp boundaries with the intact host rock or cataclasites; or grade into true cataclasites that eventually grade into relatively intact host rock. In most cases the veins will either have two sharp boundaries, or one sharp boundary with the other boundary being gradational with cataclasite. Sharp boundaries are usually undulating at the thin section scale with maximum undulation amplitudes of ~1 mm, and have injection veins protruding from them.
- Injection veins that link parallel generation surfaces and form networks and reservoir sites are commonly very clast rich, and contain abundant and angular lithic fragments.

4.42 Pseudotachylytes hosted in augen mylonites

The pseudotachylyte veins in the augen mylonites described in section 3.41 have a relatively large volume of plagioclase feldspar clasts (figure 4.04a), that preserve the altered texture that they exhibit as porphyroclasts in the fractured and cataclased mylonites that are adjacent to many of the veins. Quartz clasts exhibit undulose extinction. Calcite clasts are also sometimes present, and are especially dominant in one sample (figure 4.04b). The veins range from being relatively clast rich (figure 4.04b) to clast poor (figure 4.04a). Polymineralic lithic fragments are generally observed in thin, clast rich veins and injection structures.

One sample (OU77805) contains abundant vesicles (figure 4.05a) that dominate the resolvable material in the vein, with a very low proportion of lithic fragments. The volume fraction of the vesicles is approximately 10-15 %, although this is based on broad estimations from thin section. Quartz fills most of the vesicles, although calcite fills a number of them, and some are filled by a mixture of the two phases. The vesicles are highly circular or slightly
ellipsoidal in shape and have blurry and cracked margins (figure 4.05a). The cracks in the margins often tend toward the centre of the vesicle. In areas of high ellipsoidal vesicle concentration, there emerges an orientation of their long axis parallel to the direction of flow. The vesicles, especially the larger ones, have been fractured by later fractures and veining. The quartz in many of the vesicles exhibits sweeping, smooth undulose extinction. Very fine-grained epidote and chlorite was also observed around the rims and in fractures of the vesicles. The pseudotachylyte matrix in some areas of this sample has been altered to a brown-yellow colour and the vein has been pervasively fractured and veined. Calcite veins cross cut this sample and some of the vesicles (figure 4.05b). Textures indicate calcite later replaced some of the quartz in many of the vesicles, as shown in figure 4.05b. The vesicular sample is from a voluminous injection structure found in a boulder at Harold Creek.

Figure 4.04: Photomicrographs of pseudotachylyte hosted in augen mylonite. a) OU77780. Plagioclase and quartz clasts set in the fine grained, dark matrix. b) OU77702. Calcite clast in pseudotachylyte matrix exhibiting distinctive twinning.

Figure 4.05: Photomicrographs of vesicles in augen mylonite-hosted pseudotachylytes from sample OU77805. b) Quartz has precipitated around the edges of the vesicle and calcite in the middle. The vesicle has also been faulted by later fracturing of the pseudotachylyte vein.
The matrix of the pseudotachylytes in the augen mylonites ranges from being dark grey/black (figure 4.05a) to dark brown (figure 4.05a) to light brown-yellow. The yellow coloured matrix is composed of extremely fine-grained alteration products, and possibly also fine sericite and calcite. In some areas of the veins the matrix has been replaced by chlorite. Late stage quartz and calcite mineralization is found to commonly occur in the same veins, which are further cross-cut by thin, anastomosing chlorite filled veins.

**Discussion and interpretations**

The presence of the vesicles in the one sample of augen mylonite-hosted pseudotachylyte is not only a very strong indicator of a melt origin for the pseudotachylyte, but their presence is also testament to the exsolution of a vapour phase from the melt under low lithostatic pressures (Maddock et al., 1987). By determination of wt % of H$_2$O and CO$_2$ phases present in the host rock and the volume fraction of vesicles, estimations of lithostatic pressure and therefore depth of formation can be attained, as has been carried out by Maddock et al. (1987). This option will be discussed further in chapter 6.

Quartz that infills the vesicles exhibits undulose extinction. This means the vesicular pseudotachylyte underwent some further deformation after its formation. The elongation of some of the vesicles could be attributed to this later deformation, however, it could equally be attributed to flattening and alignment experienced immediately after their formation due to flow in the surrounding melt. Undulose extinction indicates that dislocation creep may have been operative in the quartz filled vesicles, and the presence of undulose extinction alone suggests the quartz was undergoing recovery suggesting temperatures of ~300-350 °C (Passchier & Trouw, 1996). Down dip of the Alpine Fault, the onset of quartz plasticity at these temperatures occurs at approximately 5 km, according to the fault zone model of Toy (2007).

The pseudotachylytes hosted in augen mylonites were probably formed either after or were coeval with cataclasis. The pseudotachylyte was produced in relatively large quantities, and infilled large reservoir sites, that are particularly voluminous compared to the fault veins. Calcite veining and precipitation through the rock occurred before and after cataclasis and pseudotachylyte formation, as is evident by the presence of both clasts of calcite in the fault veins and cross-cutting calcite veins. Calcite veining was possibly pseudo-synchronous with quartz veining, however quartz did infill the vesicles in sample OU77805 first. Chlorite veining and alteration of the pseudotachylyte matrix occurred later.

### 4.43 Pseudotachylytes hosted in schist-derived mylonites

#### 4.43.1 Pseudotachylyte veins parallel to mylonitic foliation

Thin section observation reveals that injection structures in these pseudotachylytes are common. They are regularly controlled by microfaults at acute angles to the fault vein that bound them on one side, forming a wedge-like shape. The microfaults are formed along the C’ shears in the mylonites. These injection structures are generally short and broad, however can taper out over quite some length as the pseudotachylyte is injected up the microfault. Very small, less common injections structures that are not associated with the microfaults are also present, and are short, broad embayments into the wall rock at high angles to the fault vein, and commonly taper out over a short distance (<1 mm).
Clasts in the veins are predominantly quartz and feldspar, with some lithic fragments, especially in clast-rich veins. The quartz clasts exhibit undulose extinction and the feldspars are largely plagioclase. The matrix is generally a brown or black cryptocrystalline matrix, and in some places sericite can be seen replacing it. These pseudotachylytes are commonly associated with cataclasites and ultracataclasites, and will commonly lie adjacent to cataclasite (figure 4.06a), or as in many cases, the pseudotachylyte veins will grade into a cataclastic texture (figure 4.06b). When this is the case, the pervasion of the pseudotachylyte into the reservoir of the cataclased material creates a largely clast rich pseudotachylyte-cataclasite mixture, with large lithic fragments (figure 4.06c). Veins containing a dominantly cataclastic texture can have patches of dark material within them that may be pseudotachylyte (figure 4.06d). In some cases the pseudotachylyte patches may be cataclased themselves, or be coherent and appear to have formed in the cataclasite along shear surfaces (figure 4.06e) and contacts between the cataclastic fragments (figure 4.06f). The cataclasites adjacent to the veins can have an anastomosing network of thin shear surfaces containing fine-grained opaque material parallel or sub-parallel to the pseudotachylyte vein walls. Late calcite veins and fractures disrupt the veins. Rare calcite has also been observed as clasts in the pseudotachylyte veins from Gaunt Creek (Toy, 2007).

Calcite and quartz filled vesicles are present in one sample that range in size from about 10 to 100 µm. The quartz infilling the vesicles exhibits undulose extinction and sub-grain development. This sample also possibly has at least two generations of pseudotachylyte veining the first being the vesicular generation that is cross-cut by a more clast-poor generation that exhibits very visible flow banding and viscous folding.

4.43.2 Pseudotachylyte veins oblique to the mylonitic foliation

Unsurprisingly, this group of veins yields largely identical microscopic observation to the foliation parallel veins described above. The main difference is that they are not as closely related to cataclasites as the foliation parallel pseudotachylytes. It should also be noted that these veins also exhibit some small fault-bounded injection veins, but these are barely significant as reservoir zones compared to the large injection veins that occur along foliation planes in the mylonite.

Discussion and interpretations

From the apparent presence of calcite clasts in pseudotachylyte veins hosted in schist-derived mylonite, it has been suggested that calcite veining occurred before cataclasis and pseudotachylyte generation (Toy, 2007). The formation of mixed cataclase and pseudotachylyte veins occurred before further cataclasis and shearing on anastomosing shear surfaces in some of the veins. Secondary formation of pseudotachylyte may have occurred forming veins that have very few but small clasts in them, that may exhibit well-preserved flow banding and viscous folding in the matrix. Undulose extinction and sub-grains in the quartz that infills the vesicles of one sample, indicates that the quartz deformed by dislocation creep after this pseudotachylyte was generated. Quartz deforms by dislocation creep at temperatures of 300-350°C, which corresponds to ~4.5 km depth using the down-dip geothermal gradient of Toy (2007). This generation of pseudotachylyte could therefore have formed at or below this depth range.

From thin section observation it appears pseudotachylyte formation, especially in the foliation-parallel veins, was closely related to cataclastic deformation. The cataclasite veins that contain bits of pseudotachylyte within them, or
Figure 4.06: Photomicrographs in PPL illustrating the association between cataclasites and pseudotachylytes in the schist-derived mylonites. a) top left. OU77809. Cataclasite layer in the top of the image lies adjacent to the pseudotachylyte at the bottom of the image. b) top right. OU77901. Pseudotachylyte at bottom of image grades into cataclasite at the top of the image. c) OU77845 and d) OU77834 middle row. Clast rich pseudotachylyte cataclasite mixtures. e) and f) bottom row. OU77834. Pseudotachylytes along grain contacts and shear surfaces in cataclasite-pseudotachylyte mixtures.
‘mixed’ cataclasite-pseudotachylyte veins (figure 4.09c, d, e, f), suggest that either; pseudotachylyte veins formed and then were later cataclased, or pseudotachylyte melt ‘blebs’ formed within the cataclasite along shear surfaces or between grain contacts during cataclastic deformation. Evidence for both of these processes exists. Some pseudotachylyte is fragmented, with angular clasts of pseudotachylyte in the cataclastic matrix, that may be the remains of earlier formed pseudotachylyte veins or patches, while some of the pseudotachylyte is relatively coherent and appears to have formed as in-situ patches. Pseudotachylyte along grain contacts in these mixed veins, such as that shown in figure 4.06e/f could have formed before or during cataclasis. Therefore it is uncertain whether some of the pseudotachylyte fragments formed as patches in the cataclasite vein, or are remnants of an earlier pseudotachylyte vein that has been subjected to cataclasis.

Nevertheless, it is apparent pseudotachylyte melt patches have formed in these pseudotachylyte-cataclasite veins, due to sufficient temperature increases realised by slip between cataclasite fragments, or on localised shear surfaces in the cataclasite at seismic slip rates. This localised production of pseudotachylyte material was observed by Hirose and Shimamoto (2005a) in experimentally produced pseudotachylytes in gabbro. Toy (2007), calculated that the slip increment on the shear surface in the mixed pseudotachylyte-cataclasite needed to create a 200 µm thick isolated pseudotachylyte ‘bleb’, was only 5 mm. Equation 13 was used for this calculation, assuming a total frictional resistance drop (i.e. ignoring the part of E that is converted to heat), and used \Delta T = -800^\circ C, \phi = 0.2, \tau_f = -135 \text{ MPa}. Toy (2007) also calculated the total slip required on the boundary of the cataclasite layer to cause melting was ~0.4 m, which was similar to typical seismic slip increments found by other workers (Sibson, 2003b; Di Toro et al., 2005), and therefore concluded that the isolated blebs of pseudotachylyte were probably formed during seismic events. In addition, it is also likely the associated cataclasites were deformed at the seismic slip rates that produced pseudotachylyte on the localised shear surfaces.

4.44 Pseudotachylytes hosted in cataclasites

4.44.1 Pseudotachylytes hosted within pale green cataclasites

Petrographic observations reveal the clasts within the brown pseudotachylytes hosted in the pale green cataclasites are mostly quartz and feldspar (figures 4.07a, b, c). Quartz clasts dominate in abundance and show undulose extinction. Feldspar clasts preserve their highly saussuritised appearance that they display in the host rock, and some of them can be quite large, but are small and difficult to recognise. A relatively large amount of Ti-rich mica phase is also present as very small clasts in many of the veins, especially in the sample OU7780610, and these can only be seen with BSE images as shown in figure 4.10c. An EMP analysis of this phase is included in section 4.52. Composite clasts composed of both quartz and feldspar are also commonly observed in the matrix, exhibiting the quartz-cemented texture inherited from the host rock (figure 4.07b, c). In many of the composite clasts, especially the smaller ones, it is observed that quartz is often in the centre and is surrounded by a rim of feldspar.

The matrix of the pseudotachylytes veins in the pale green cataclasites is generally a dirty brown colour that has been largely replaced by chlorite or sercite in some places. Thin quartz, chlorite, calcite and very thin veins of a high relief second order birefringent mineral that could be epidote, cross cut the pseudotachylyte veins in many places. Veins also are present cross-cutting the host rock, but not the pseudotachylyte veins.
A melt origin could not be confirmed for any of the blue-tinged veins collected. It was found from thin section and BSE imaging that three of them were intact ultracataclasites, based on a large abundance of angular fine-grained clasts of most of the mineral phases observed in the host rock, and lacking melt textures. Photomicrographs showing the ultracataclastic texture are shown in figures 4.08a, b, c and d for samples OU80611, 80613 and 80614. The latter two samples were collected from the strange, flame-like injection structure described in section 3.44.1, which means that it could be a fluidized injection of very fine-grained ultracataclasite.

Interestingly, isolated patches of more fine-grained dark coloured material were found in the ultracataclasite OU80611 (figure 4.08 b). The patches seem, from thin section observation, to be intact and coherent, so may have formed in-situ by shearing along localised surfaces, or along grain contact surfaces in the cataclasite, similar to those...
described in section 4.43. Flattened vesicles observed in one sample are filled with calcite and quartz. The vesicles have their long axis aligned in the direction of flow banding, so it was likely the viscous flow of the melt was responsible for their flattening.

Cataclasites that bound many of pseudotachylyte veins exhibit a true cataclastic texture, as opposed to the protocataclastic texture of the pale green unit as a whole. The pseudotachylyte veins are often gradational with these cataclasites perpendicular to the vein walls. If a true cataclastic texture has not developed adjacent to a pseudotachylyte vein, there is often a zone adjacent to the vein cut by thin, anastomosing, through-going shears that are approximately sub-parallel to the vein walls, that may, or more often not have pseudotachylyte in them. Many of the cataclasites adjacent to the pseudotachylyte veins have been cemented by quartz.
Discussion and interpretations

The presence of quartz-calcite-chlorite-filled veins that both cross-cut and are cross-cut by the pseudotachylyte veins indicates that fluids were present before and after pseudotachylyte generation. Further cementing of the cataclasites after pseudotachylyte generation, as indicated by cemented cataclasites that appear to have been coeval with pseudotachylyte formation, is further evidence of fluids being present after pseudotachylyte generation.

The isolated patches of pseudotachylyte in some of the ultracataclasites appear to be relatively coherent indicating they have formed in the ultracataclasite due to sufficiently high temperature increases realised on slip surfaces, and between cataclasite fragments as discussed in section 4.43. The presence of these isolated melt ‘blebs’ indicates that pseudotachylyte and cataclasite formation was coeval, and highlights that the two processes are closely related, in which cataclasis is the precursor to frictional melting. Toy (2007) observed similar melt bl’eb’s in the pale green cataclasite and calculated, using equation 13 with $\tau_f \approx 150$ MPa and $\Delta T \approx 1300^\circ$C, that it would take a slip increment of 10 mm to melt a layer 200 $\mu$m thick. This slip increment is only a small fraction of the total slip across the $\sim 15$ m thick pale green cataclasite zone, which is thought to experience displacements of $\sim 8$ m during seismic events (Berryman et al., 1998).

Most of the pseudotachylyte veins observed within the pale green cataclasite are $\sim 5$ mm thick, and occur as non-planar, discontinuous, effectively anastomosing surfaces throughout the unit. In addition to these surfaces, very thin ($<10$ $\mu$m), anastomosing, through-going shear surfaces are visible in all the thin sections observed. These surfaces have been observed to disrupt the bulk of the pseudotachylyte veins, and may contain pseudotachylyte themselves. The relatively thick pseudotachylyte vein therefore formed before the thin shear surfaces that accommodate most of the slip during a seismic event, during which production of either ultracataclasite or pseudotachylyte occurs along them. Slip is highly localised along these thin ($<10$ $\mu$m) shear surfaces that accommodate all the slip during a seismic event. The thicker pseudotachylyte veins would have accommodated earlier seismic slip with the zone, and may possibly be still accommodating seismic slip at depth.

The most pervasive cataclasis may have occurred before pseudotachylyte generation, during the cementation-failure cycles discussed in section 4.33.2. Pseudotachylytes and ultracataclasites were generated in voluminous quantities along non-planar shear surfaces that often occur around material inhomogeneities in the pale green cataclasite, such as relatively well-indurated quartz and feldspar pods. These veins were then overprinted by later brittle deformation along thin, anastomosing shear surfaces.

4.44.2 Pseudotachylytes hosted in cataclasized ultramylonites

At thin section scale, many more, thin pseudotachylyte veins are visible that intrude between foliation layers or along shear surfaces and microfaults. The pseudotachylytes hosted in the cataclasized ultramylonites have essentially the same petrography to those in the pale green cataclasites, except have mica clasts absent, and overall have less clasts (figure 4.09a). Some veins also contain abundant chloritised microspherulites (figure 4.09b). The matrix is commonly replaced by chlorite and sericite. Pseudotachylyte veins are commonly cross-cut by numerous chlorite, quartz and calcite veins, and disrupted by networks of microfaults.
These pseudotachylites formed by the shearing together of the green and black ultramylonites after the formation of their disjunctive cleavage, and were coeval with the formation of microfaults that they intrude along. The pseudotachylites were then slightly cataclased and rotated along later microfaults, and thin, anastomosing shear surfaces, similar to those in the pale green cataclasites.
4.45 Conclusions from petrography

The following conclusions can be drawn from petrographic observation of the Alpine Fault pseudotachylytes regarding their formation:

- Pseudotachylytes hosted in augen mylonites were generated at the same time as cataclasites, with more than one episode of melting and cataclasis being evident. One generation of pseudotachylyte was vesicular and contained plastically deformed quartz infillings indicating pseudotachylyte generation at least at depths at or below ~4.5 km at temperatures of 300-350°C. Later veining cross-cut the pseudotachylyte and alteration occurred within their matrices.
- Multiple generations of pseudotachylyte production occurred in the schist-derived mylonites, and were all coeval with cataclasis. One generation was vesicular, with plastically deformed quartz infilling the vesicles indicating a depth of generation of ≥5 km. Planes of weakness developed along C’ shears in the mylonites were exploited by the pseudotachylytes for reservoir sites.
- Pervasive, high fluid pressure-assisted cataclasis probably occurred before the bulk of the pseudotachylyte generation in the pale green cataclasites. The pseudotachylytes were generated along non-planar shear surfaces, often around coherently cemented or well indurated quartz-feldspar pods, and are disrupted by through-going anastomosing, thin shear surfaces. The thin shear surfaces may have been accommodating slip at the same time as the pseudotachylyte production, or they are possibly shallower brittle overprints.

4.5 Evidence for melting

All the veins that will be further analysed in the proceeding sections preserve evidence of having been through a melt phase i.e. evidence that the matrix of the pseudotachylyte was once molten and quenched to an aphanitic glass. The evidence for a melt origin in the veins can only be observed by petrographic methods. Except for possible microspherulites in one pseudotachylyte vein, no newly crystallised material was observed in the Alpine Fault zone pseudotachylyte samples, at least with the methods used in this study. The key indicators of a melt origin for pseudotachylyte were summarised in section 1.4, with references to previous studies that have used them. Some of these indicators were realised from the petrographic observation of the three groups of veins in the Alpine Fault zone:

- Quartz and feldspar clasts in the pseudotachylytes commonly have irregularly embayed and interlobate boundaries, but often smoothly rounded boundaries (figures 4.10a, b). This indicates they were partially melted around their edges by the enveloping friction melt.
- The matrices of the veins are generally composed of dirty brown-yellow material that cannot be resolved by the petrographic methods used in this study (figure 4.10b, c, d). It could be that this brown material is altered basaltic glass that has been converted to smectites, ferric oxides and chlorite under hydrothermal conditions (McPhie et al., 1993). Evidence for pervasive hydrothermal alteration is abundant throughout the Alpine Fault zone rocks examined in this study, and in some veins the matrix has a slight green tinge which may be fine-grained chlorite replacing the glass. If the brown material is palagonitised, devitrified glass, another indication that the material has been through a melt phase arises.
The material along the margins of many of the veins has finer grained clasts suspended in a matrix that is commonly darker, more opaque or altered differently than in the vein centres (figures 4.10b,c). These could represent chilled margins of a molten vein that was in contact with relatively cool host rock.

The presence of vesicles in two samples (figure 4.05a, b) is unequivocal evidence of a melt origin, as they form only by exsolution of volatiles from a melt (Maddock et al., 1987). However, there is some disagreement whether pseudotachylyte melts could exsolved gases forming vesicles within the very short timeframes that they are molten. This will be discussed further in chapter 6.

Viscous folding textures, highlighted by bands of different coloured matrix observed in some of the more matrix-rich veins indicate flow of a liquidised phase. Alignment of flattened vesicles also indicates viscous flow.

Figure 4.10: Melting relationships in Alpine Fault pseudotachylytes. a) top left, OU80607. A large clast of quartz is corroded and rounded around its edges, as well as many other well rounded clasts present in the image. b) top right irregularly embayed quartz clast in center of image. c) left OU77804 and d) right OU80607, bottom row, chilled margins with different colours and less clasts than the vein centres.
4.6 Host Rock and Pseudotachylyte Geochemistry

The aims of the geochemical analysis is to highlight certain systematic chemical relationships between host rock and pseudotachylyte that can be used as further evidence of a melt origin (section 1.4), and to attain compositions of the host rock and matrix for later calculations. Both analysis of individual components in the pseudotachylyte veins by electron microprobe, and whole rock powder X-ray Fluorescence (XRF) analysis of one of the host rocks was undertaken.

Several workers have previously carried out geochemical analysis of pseudotachylytes (e.g. Spray, 1987, 1992, 1993; Maddock, 1992; Magloughlin, 1992; O’Hara, 1992; Curewitz & Carson, 1999; O’Hara & Sharp, 2001; Barker, 2003), and have found that pseudotachylytes are chemically distinct from their host rocks with respect to certain chemical component. Generally, pseudotachylyte matrices are commonly more basic than their host rocks and may show enrichment in K₂O, Al₂O₃, MgO, TiO₂ and Fe.

4.61 Methods and analytical rationale

Energy dispersive spectrometer (EDS) Electron microprobe (EMP) analyses were carried out on polished, carbon-coated thin sections in a JEOL JXA-8600 Superprobe, in the Otago University Geology Department, under the direction of Dr. Kat Lilly. The operating voltage for the EDS analysis was 15 kV, with a beam current of 20 nA, and a count time of 200 ‘live’ seconds at ~ 2000 counts s⁻¹. Matrix corrections were by ZAF, and well-known, homogenous oxide standards were used.

Pseudotachylyte matrix and clasts were analysed, as well as “bulk” measurements of the host rocks. Measurements of clasts were carried out with a 2 µm (turned to ‘0’ on the control) probe diameter, with the aim of identifying the mineralogy of the clasts. A 15 or 20 µm probe diameter was used for matrix and host rock readings. Some of the host rock analyses were carried out with a 40 µm probe diameter. This method was used to identify “bulk” compositions of the pseudotachylyte matrix in 6 samples, and as the matrix of the pseudotachylyte is assumed to be the solidified melt, its composition should reflect the composition of the melt that formed during frictional fusion. Special care was taken when performing matrix readings to avoid the influence of lithic clasts in the pseudotachylyte, although this influence or the influence of possibly newly crystallised material on the measured compositions is difficult to avoid due to their possible irresolvable grain-size. Prior to the EDS spot analyses backscatter electron (BSE) images were taken for locating spot analysis points, and also for use in image analysis. The BSE imaging method was described in section 4.2. Poor quality polishes on the thin sections limited the quality of many of the BSE images, and therefore the ability to identify matrix areas and carry out EDS spot analysis was hampered for many samples. Three pseudotachylyte veins hosted within the pale green cataclasite were analysed, with very thorough, representative analyses on OU80610 and OU80607. One vein from the cataclasied ultramylonite was analysed, but a poor polish limited the number of analyses carried out. A single vein of pseudotachylyte hosted within an augen mylonite boulder from Harold Creek was analysed, and a single vein of pseudotachylyte hosted in an Alpine Schist-derived mylonite boulder also from Harold...
Creek was analysed. The mylonite samples were easily polished, and many spot analyses of the pseudotachylyte matrix were able to be completed.

Major element whole rock analyses were carried out on three powdered samples of the pale green cataclasite. XRF analysis was carried out on fused glass disks that were prepared with 0.64g of 105°C dried, powdered sample, 6.8g of Type 12:22 XRF flux composed of 35.3% lithium tetraborate and 64.7% lithium metaborate, and 1g of ammonium nitrate. Dr. Kat Lilly performed the disk fusion and the XRF analyses. The Otago University geology department Phillips PW2400 X-Ray Fluorescence Spectrometer was used for the XRF analysis. Loss on ignition values were calculated from 2g of 105°C dried sample powder heated to ~1100°C. XRF bulk rock analyses of Alpine schist-derived mylonites and augen mylonites were taken from analyses of Toy (2007). The bulk rock analyses of the host rocks were carried out for comparison to the pseudotachylyte matrix. However, it should be noted these ‘host’ rocks are not the ‘wall’ rocks of their corresponding pseudotachylyte veins, instead they were collected from chosen ‘representative’ areas of the host rock lithologies as a whole.

4.62 Samples

The pseudotachylyte samples that were collected during fieldwork for this study and analysed are OU80607, 80617, 80616, 80610. Host rock samples of the pale green cataclasite that were also collected and analysed in this study are OU80604 and OU80612. These samples are representative portions of the pale green cataclasite, and do not contain quartz and feldspar-rich pods. Pseudotachylytes hosted in Augen and schist-derived mylonite samples included in the analysis, were collected by V. Toy as boulders from Harold Creek. The augen and schist-derived mylonite samples were collected from Harold Creek and analysed by XRF by V. Toy also, by the same XRF analysis method described above.

4.63 Results

The results obtained for the whole rock major element XRF analyses of the pale green cataclasites from Gaunt Creek are presented below in table 4.1. Table 4.1 also includes whole rock major element XRF analyses of two Western Province granitoid-derived augen mylonites and two Alpine Schist-derived mylonites from Toy (2007). These samples (OU77741, 77755, 77734, 77740) were all collected from in situ outcrops at Harold Creek.

The results of the analysis of the matrices of the pseudotachylytes are presented in table 4.2. The SiO₂ contents of the pseudotachylyte matrices span from approximately 50 wt% to 61 wt% and therefore include upper basic compositions. Sample GC0924 has notably high K₂O and Fe₂O₃ concentrations, and ~10 wt% less SiO₂ than the other matrices. The results of the whole rock XRF analyses of the host rock and the EDS spot analysis of the ‘bulk’ pseudotachylyte matrix, have been presented as silica variation plots (figure 4.11) and column bar graphs of the ratio of pseudotachylyte matrix to host rock compositions (figure 4.12). These diagrams allow for the comparison of the ‘bulk’ geochemistry of the pseudotachylyte matrices and the bulk rock geochemistry of their host rocks. In figure 4.11, the pseudotachylytes (blue diamonds) and their corresponding host rocks (red squares) are linked by tie lines that highlight the variation between the host rock and the pseudotachylyte. Sample OU80616 was not included in these diagrams as only 2 matrix measurements were carried out. The pseudotachylyte sample (OU80617) hosted in cataclasited ultramylonite was not included in these diagrams as no corresponding bulk analysis of the cataclasized ultramylonite was carried out.
Relative to their host rock all of the samples show matrices depleted in SiO2, and enriched in Al2O3 and TiO2. Consistent, steep enrichment in the matrix of Al2O3 is shown in the silica variation diagram, with a similar but not so steep enrichment in metallic oxides, apart from one sample. There is less consistency in the alkali and lime variation plots, with mylonite-hosted pseudotachylytes showing enrichment in these components relative to their host rock, but the cataclasite-hosted pseudotachylytes show more variation. The pale green cataclasite-hosted matrices show enrichment relative to their host rocks for total alkalis (figure 4.11) and K2O, however, Na2O is depleted in three of the four matrix-host pairs, and CaO is depleted in the matrix of two of the matrix host pairs (figure 4.12). There is also depletion of K2O in the matrices of augen-mylonite hosted pseudotachylytes (figure 4.12).

<table>
<thead>
<tr>
<th>Location</th>
<th>Gaunt Creek (this study)</th>
<th>Harold Creek (Toy, 2007)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample</td>
<td>OU80604-Ccl</td>
<td>OU80612-Ccl</td>
</tr>
<tr>
<td>SiO2</td>
<td>71.34</td>
<td>69.37</td>
</tr>
<tr>
<td>TiO2</td>
<td>0.29</td>
<td>0.45</td>
</tr>
<tr>
<td>FeO*</td>
<td>2.49</td>
<td>3.45</td>
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<tr>
<td>MnO</td>
<td>0.09</td>
<td>0.04</td>
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<tr>
<td>MgO</td>
<td>1.13</td>
<td>1.63</td>
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<tr>
<td>CaO</td>
<td>0.97</td>
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<tr>
<td>Na2O</td>
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<tr>
<td>K2O</td>
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<tr>
<td>P2O5</td>
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<td>0.16</td>
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<tr>
<td>LOI</td>
<td>-</td>
<td>1.896</td>
</tr>
<tr>
<td>Total</td>
<td>99.04</td>
<td>99.461</td>
</tr>
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</table>

Table 4.1: Results of the XRF whole rock analysis on pseudotachylyte host rocks from this study and Toy (2007).

<table>
<thead>
<tr>
<th>Location</th>
<th></th>
<th></th>
<th>Pale green cataclasis</th>
<th>Cataclasite</th>
<th>Granitoid-derived augen mylonite</th>
<th>Alpine Schist-derived mylonite</th>
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<tr>
<td>Host Rock</td>
<td>Sample</td>
<td>No. of Analyses</td>
<td>OU80607</td>
<td>OU80610</td>
<td>OU80616</td>
<td>OU80617</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>18</td>
<td>18</td>
<td>2</td>
<td>4</td>
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<td>SiO2</td>
<td>60.365</td>
<td>60.998</td>
<td>60.63</td>
<td>61.27</td>
<td>61.471</td>
<td>60.760</td>
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<tr>
<td>TiO2</td>
<td>0.675</td>
<td>1.027</td>
<td>0.895</td>
<td>1.355</td>
<td>0.925</td>
<td>1.183</td>
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<tr>
<td>MnO</td>
<td>0.049</td>
<td>0.078</td>
<td>0.05</td>
<td>0.090</td>
<td>0.101</td>
<td>0.065</td>
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<tr>
<td>MgO</td>
<td>1.981</td>
<td>4.381</td>
<td>1.89</td>
<td>2.305</td>
<td>1.837</td>
<td>2.812</td>
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<tr>
<td>CaO</td>
<td>2.519</td>
<td>0.483</td>
<td>0.535</td>
<td>0.510</td>
<td>5.235</td>
<td>3.131</td>
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<td>Na2O</td>
<td>3.351</td>
<td>0.543</td>
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<td>1.052</td>
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<td>P2O5</td>
<td>0.084</td>
<td>0.188</td>
<td>0.695</td>
<td>0.600</td>
<td>0.268</td>
<td>0.468</td>
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<td>Cr2O3</td>
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<td>0.031</td>
<td>0.12</td>
<td>0.050</td>
<td>0.070</td>
<td>0.040</td>
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<td>Total</td>
<td>100.257</td>
<td>97.462</td>
<td>96.325</td>
<td>98.365</td>
<td>98.824</td>
<td>100.005</td>
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</table>

Table 4.2: Results of the EMP EDS analysis of the pseudotachylyte matrices.

Primarily H2O and possibly some CO2, SO2 and halides comprise the volatiles that are considered to be ‘lost’ when they are vaporised from the sample by the electron beam during microprobe analysis. Due to the absence of more sophisticated techniques of determining volatile contents in the matrix, the ‘loss’ following microprobe analysis is considered to be the volatiles. Samples OU80607 and OU777809 have pseudotachylyte matrix totals (table 4.2) exceeding 100 wt% by a small fraction, while for the rest of the samples (excluding CO2) the totals indicate approximate H2O losses of 1.6% for OU80617, 3.6% for OU80616, 2.5% for 80610, and 1.2% for OU77805. Using the discrepancy between 100 wt% and totals obtained from the microprobe analyses is probably reasonable as anhydrous
mineral standards were measured at 100% and hydrous mineral phases that occur in all the host rocks (amphibole, chlorite, micas) but generally not as clasts in the pseudotachylyte veins were also analysed accurately.

Some BSE images, especially in the mylonite-hosted pseudotachylytes show areas of light and dark coloured matrix (figure 4.21a, b), and as described in section 4.2, this is due to different atomic numbers between the colours. A lighter matrix would have higher atomic numbers while lower atomic numbers would be associated with darker areas of the matrix. A number of EDS spot analyses of the light and dark areas in the matrix reveal that in general the darker areas of matrix are silica rich with an average of ~68 wt% SiO$_2$, and that the lighter areas are basic in composition with ~52 wt% SiO$_2$. The lighter areas are also have high Fe$_2$O$_3$ and MgO relative to the darker areas (figure 4.21a, b).

**Figure 4.11:** Silica variation diagrams for the alkalis, alumina, lime and the metallic oxides between the host rocks (red squares) and the pseudotachylyte matrixes (blue diamonds).
An EMP analysis of the one of the relatively abundant clasts in the pale green cataclasite-hosted pseudotachylytes that appear bright white in BSE images (figure 4.11c) is shown in table 4.3. This mineral has a very high Ti content, as well as being a K-Fe-Al silicate.

Figure 4.12: Column bar graphs showing the ratio of matrix composition to the host rock composition. For the
Alpine Schist derived mylonites host 1 = OU77741, host 2 = OU77755. For the pale green cataclasites the
pseudotachylytes are 1 = OU80607 and 2 = OU80610, and the host rock samples are 3 = OU80604 and 4 =
OU80612. For the augen mylonites host 1 = 77734, host 2 = OU77740.
4.64 Discussion and interpretations

As the host rock samples were not collected directly adjacent to the pseudotachylyte veins, they may not be entirely representative of the material that melted to form the pseudotachylyte matrix. As pseudotachylyte forms along discrete surfaces in rock, only material comminuted and subsequently melted along this plane will be included in the melt, and the bulk chemistry of that plane will reflect the chemistry of the melt that is formed from frictional fusion. The host rocks in this study have all been found to be heterogeneous, and contain lithological layering and mineralogical banding, therefore sampling of a representative portion of the host rock for geochemical comparison to a pseudotachylyte vein formed along a discrete plane that probably does not include all the heterogeneities, is difficult. For example, one pseudotachylyte vein OU80610 collected from the surface of a quartz-feldspar pod in the pale green cataclasites has incorporated high K₂O into its matrix, possibly from the K-feldspar rich pod. Therefore the ‘corresponding’ host rocks for this particular vein probably do not represent the material that actually melted to form it.

Another important consideration is that the host rocks may have experienced further bulk chemical changes after pseudotachylyte generation, for example further silica addition or hydration, therefore may have a different composition to the rocks that were involved in pseudotachylyte generation. Nevertheless, the overall geochemical relationships between pseudotachylyte and host rock observed in the Alpine Fault pseudotachylytes are in agreement with those obtained by other researchers, in that the matrices of the pseudotachylyte show general depletion in SiO₂ relative to their host rocks and enrichment in Al₂O₃, and metallic oxides.

The observed geochemical relationships between the pseudotachylyte matrices and the host rock can be explained by the interaction of the mechanical and thermal properties of the minerals that comprise the host rock. As cataclasis is commonly considered to be the precursor to frictional melting (e.g. Maddock, 1992; Magloughlin, 1992) the mechanical properties that control the fracture and comminution of the minerals plays a very important role in determining the composition of the melt. The differences in shear yield strength, fracture toughness and melting points of the host rock-forming minerals during high strain-rate, coseismic conditions results in a hierarchy of susceptibility to friction melting (Spray, 1992). In this study, very few amphibole, biotite, muscovite or chlorite clasts are generally observed in the pseudotachylyte veins, and as they are known to occur in the host rocks, they are assumed to have been preferentially incorporated into the melt relative to quartz and feldspar that predominate the ‘survivor’ clast assemblage. This observation cannot be explained by igneous melting processes, in which equilibrium eutectic melting generates a minimum melt. However, non-equilibrium selective melting of the mineral phases of the precursor cataclastic material that possess lower shear yield strengths, fracture toughnesses and melting points relative to survivor clasts, explain the observations well. Amphibole, biotite, muscovite and chlorite all possess lower shear yield strengths (MN m⁻²), fracture toughnesses (MPa m⁰⁵) and melting points than quartz and the feldspars, and the preferential cataclasis and subsequent inclusion of these hydrous phases into the melt phase explains their absence as clasts in the pseudotachylytes observed in this study.

One interesting observation that seems to differ from this is the relatively highly abundant mica mineral present as small clasts in the matrix of the pale green cataclasite-hosted pseudotachylytes, that cannot be resolved with optical microscopy but exhibits bright white colour in BSE images. EMP analysis of this mineral reveals it has a very high TiO₂ component within an otherwise biotite-like composition (table 4.3). This could be interpreted as a mixture of biotite and rutile, with abundant rutile possibly present along cleavage surfaces of the biotite. Its presence in the melt could indicate either the temperature rise and comminution process was not sufficient to fully incorporate the mineral
The Alpine Fault pseudotachylytes have matrices depleted in SiO$_2$ and generally enriched in Al$_2$O$_3$, alkalis (Na$_2$O + K$_2$O) and metallic oxides (Fe$_2$O$_3$ + TiO$_2$ + MgO + MnO). This is inferred to be due to the preferential inclusion of ferromagnesian amphiboles and biotite, as well as alkali-alumina micas and feldspars, into the melt phase and the exclusion of quartz. This selective inclusion is not a result of equilibrium partial melting, but is the result of a hierarchy of frictional melting susceptibilities controlled by shear yield strength, fracture toughness and melting points of the minerals that comprise the host rock. Micas, amphiboles and feldspar possess lower values of the parameters relative to quartz, and are therefore included in the melt phase.
Chapter 5

Microstructural Analysis

5.1 Introduction

Brittle fault rocks that occur along localised zones of slip are composed of fine gouge and mineral and rock fragments, and in the case of pseudotachylytes, of mineral and rock fragments supported within a matrix that cooled from melt formed by selective frictional fusion of certain host rock minerals. Pseudotachylyte clasts, whether monomineralic or polymineralic, exhibit a wide range of sizes and shapes, and are generally not apparent from mesoscopic observations but with optical microscopy or electron backscatter (BSE) imaging, the majority of the clasts stand out prominently from the matrix.

Analysis of clast size and shape within brittle fault rocks gives insights into the genetic processes of fragmentation and frictional melting involved. Several studies have focused on clast size analysis of naturally (e.g. Sammis & Biegel, 1989; Blenkinsop, 1991) and experimentally (e.g. Sammis et al., 1986; Biegel et al., 1989; Marone & Scholz, 1989; Heilbronner & Keulen, 2006) produced fault gouge, indicating log normal or simple power law grain size distributions and an overall decrease in grain size with increasing confining pressure. Fractal geometry can be applied to clast size distributions, as they also have fractal distributions (Blenkinsop, 1991). To overcome the difficulty of making direct links between observation and process, and to characterise the processes of fragmentation, some authors have used fractal dimensions as simple numbers establish mechanisms of fragmentation (e.g. Sammis & Biegel, 1989; Blenkinsop, 1991; Blenkinsop & Fernandes, 2000; Sammis & King, 2007).

Pseudotachylyte clasts, as they are produced, at least initially by fragmentation processes, should have very similar clast size distributions to gouges and cataclasites. Clast size analysis in pseudotachylytes has been carried out in a number of studies (Shimamoto & Nagahama, 1992; Tsutsumi, 1999; Ray, 1999, 2004), and it has been found the size distributions obey a simple power law distribution of the form:

\[ N \approx N' u^{-D} \]  

(15)

where \( N \) is the cumulative number of clasts with sizes greater than \( u \), \( D \) is the power or fractal dimension, and \( N' \) is a constant that depends on the total number of measurements. However, this grain size distribution does not survive in the fine fraction of the clasts in an analysis of a frictionally fused pseudotachylyte, and in a plot of \( \log N \) vs. \( \log u \), the tail of the distribution, at lower values of \( u \), depresses from the slope of the power law (Ray, 1999). Therefore the fine fraction is incompatible with the size distribution of the clasts, and from this arises further evidence that the irresolvable fine-grained material in pseudotachylyte is not only the product of ultracomminution, but of frictional melting. This modified power law size distribution should be characteristic of all pseudotachylytes (Ray, 2004).
This chapter presents the methods and results of the microstructural analysis of clasts within the pseudotachylyte veins collected from within the Alpine Fault zone. Clast size and volume fraction analysis has been carried out with the aim of characterising the pseudotachylyte samples, attaining aspect and clast to matrix ratios to be used in later calculations, further understanding the mechanical and thermal processes involved in pseudotachylyte genesis, and to provide further evidence of a melt origin for the pseudotachylytes.

5.2 Methods

Illustration software (Adobe Illustrator) was used to manually outline clasts of different mineralogies on BSE images. BSE images were obtained in the University of Otago Geology Department by the methods outlined in section 4.2. Except for two 80X magnification images, all the images analysed were at 200X magnification. The clasts were outlined on the images according to their mineralogy and are therefore grouped into quartz, feldspar, composite (lithic polymineralic fragments), and bright white mica mineral (for cataclasite hosted pseudotachylytes), as layers in the illustration program, and then assigned different grey shades and infilled. The clast outlining method was chosen over grey shade recognition software in this analysis because, in many of the samples, the matrix is altered and varies in colour with patches that are obviously matrix but may be a very similar grey shade to many of the clasts. This lack of grey shade consistency in the matrix would make it difficult to select grey shades that are different to the clasts in some of the images, therefore outlining is the preferred method as clasts can be chosen over patchy matrix areas.

The BSE images were removed as a separate layer, leaving the clasts of various grey shades on a white background. These images were then processed by ImageSXM*, in which the different minerals could be selected and subsequently analysed by grey shade density slicing. A scale was referenced in ImageSXM before the analysis for clast cross-sectional area, perimeter length and major and minor axis lengths was carried out on the various minerals. The data output from ImageSXM were then reduced using Excel spreadsheets.

For each clast, the size is calculated as the diameter of a circle of equivalent cross-sectional area to that clast, and these diameters are used to form the grain size distributions. Log-log plots of cumulative frequency (N) of clasts greater than size (u), were made using bin sizes of 0.2 µm over the total range of clast sizes. The grain size distributions were not plotted for each mineral group, but instead as the total of all the clasts measured in a sample. Area fractions for each mineral group in each pseudotachylyte sample are simply ratio of the cumulative area of the clasts of the mineral group to the total area of pseudotachylyte that was analysed in all the images of that sample, the sum of which is taken as being the clast to matrix ratio for the sample as a whole. Aspect ratios (major/minor axes) are simply the average of the aspect ratios of each mineral clast group, as it was observed there was very little scatter amongst the aspect ratios for each mineral.

21 images were analysed in the four samples of pseudotachylytes hosted in the pale green cataclasite, and three images each were analysed for the schist-derived mylonite and augen mylonite hosted pseudotachylytes. No samples of the cataclased ultramylonite were analysed. All the images had their total areas taken up by pseudotachylyte, except for one sample in which the area of the pseudotachylyte was only partial to the total area of the image. The smallest clasts

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* ImageSXM is public domain image processing software developed from NIH image, for MacOSX; public domain: http://reg.ssci.liv.ac.uk/
size that could be resolved and measured by the image processing software is 0.26 µm for the 200x magnification images, and 0.67 µm for the 80x magnification images. Theoretically, the smallest clasts sizes that can be outlined should be the same as these values, however, a single pixel with a different grey shade is easily overlooked, and there is little certainty that they are clasts. The range of grain sizes therefore measured in this analysis is ~1 - 1000 µm.

The pseudotachylyte samples analysed from the pale green cataclasite zone (OU80605, 80606, 80607, 80610), were all collected from the same outcrop at Gaunt Creek, while the mylonite samples analysed (OU77805, 77809) were both collected at Harold Creek from float boulders. Four (OU80607, 80610, 77805 and 77809) of the six samples analysed here were also analysed for their matrix compositions in chapter 4. Two of the samples analysed are injection structures (OU80607, 77805)

Traditional grain size analyses and distributions are measured by weighing and sieving disaggregated particles into size fractions, which gives a distribution of three-dimensional particle sizes. In this analysis, and in other analyses of pseudotachylyte clast size distribution, it is nearly impossible to separate the clasts from the matrix, and also may cause further comminution. Therefore clast analysis is carried out in two dimensions on thin sections, which do not give the same absolute particle sizes or distributions as the three dimensional measurements. This is because there is a greater probability of intersecting larger particles in two dimensions, and also the section through a particle is highly likely to show less than the maximum particle dimensions. It therefore has to be said that this analysis, like most the other coherent rock analyses, derives two dimensional particle size distributions.

5.3 Results

The cumulative frequency (N) vs. clast diameter (u) plots, where N is the number of clasts greater than or equal to u for the six samples of pseudotachylytes are shown in figure 5.1. Both axes of the plots are in a logarithmic scale. All the diagrams show a distribution of points along a relatively straight line, with lower values of u showing considerable depression from the line of best fit. Figure 5.2 shows the same charts with lines of best fit, but with u < 3 µm removed, which is approximately the value under which the points begin to scatter and depress from the best fit line and the bulk of the population in most of the plots, and it is revealed the best fit line has a better fit to the remaining plots. The lines of best fit in figures 5.1 and 5.2 can be described by the power law distribution, with a relationship as outlined in equation 15. Taking logs of both sides of equation 15 yields the straight-line equation of the points on the plots that have the exponent D as their slope. The D values for the plots in figure 5.1, which incorporate the depressed low-u points, are lower than the D values shown on the plots in figure 5.2, which are higher due to steepening of the line of best fit after elimination of the low-u points. With the exception of two of the plots, OU80607 and OU77805 that have comparatively low D values, the remaining four plots of figure 5.1 fall within the range -1.651 to -2.11. The two plots, OU80607 and OU77805 also have lower D values in figure 5.2. In table 5.1, the D values for both the plots are shown, D1 values are those derived from the plots in figure 5.1, while D2 values are derived from the modified plots of figure 5.2.

The additional results from the image analysis are shown in tables 4.1 and 4.2. The quartz volume fraction dominates the volume fraction of the other mineral phases present as clasts. There is a very small area fraction of the bright white mica even though it is abundantly present in all the pale green cataclasite-hosted pseudotachylytes,
especially in OU80610. This will be due to the fact that most of these clasts had only very low diameters. A similar relationship occurs with the feldspar and composite volume fractions, which are higher than the bright white mica

**Figure 5.1:** Grain size distributions of Alpine Fault Pseudotachylytes. The cumulative graphs show frequency of clasts greater than size $u$. The straight lines of best-fit are also shown and give the power law relationship between $N$ and $u$. The lower clast sizes ($< \sim 6 \mu m$) depression from the best-fit relationship. $D$ values for the line are also shown.
Figure 5.1: Continued
### Table 5.1: Details of the results of the clast analysis

<table>
<thead>
<tr>
<th>Sample OU no.</th>
<th>No. of clasts analysed in images</th>
<th>Cumulative Area</th>
<th>Average aspect</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>80605 80606 80607 80610 77805 77809</td>
<td>80605 80606 80607 80610 77805 77809</td>
</tr>
<tr>
<td>Quartz</td>
<td>1267 3094 997 729 804</td>
<td>1400 92567 245621 142954 106714 92076 30550</td>
<td>1.62 1.616 1.62 1.63 1.59 1.63</td>
</tr>
<tr>
<td>Feldspar</td>
<td>21 101 184 158 -- --</td>
<td>-- 30248 12374 12511 51618 -- --</td>
<td>-- 1.8 1.606 1.64 1.92 -- --</td>
</tr>
<tr>
<td>Composite</td>
<td>-- 45 45 -- -- -- --</td>
<td>-- 72486 178552 -- -- -- --</td>
<td>-- 1.617 1.55 -- -- -- --</td>
</tr>
<tr>
<td>Bright White</td>
<td>136 515 87 3085 -- --</td>
<td>2164 5853 881 31144 -- --</td>
<td>1.64 1.57 1.64 1.63 -- --</td>
</tr>
<tr>
<td>Total</td>
<td>1424 3755 1313 3972 804 1400</td>
<td>124979 336334 334898 189476 92076 30550</td>
<td>-- -- -- -- -- --</td>
</tr>
</tbody>
</table>

### Table 5.2: Results of volume fraction and aspect ratio calculations. Not measured = n.m and dashed = negligible (<0.01).

<table>
<thead>
<tr>
<th>Host Rock</th>
<th>Sample</th>
<th>Vol. frac(\text{Qtz})</th>
<th>Vol. frac(\text{Feld})</th>
<th>Vol. frac(\text{Comp})</th>
<th>Vol. frac(\text{BWM})</th>
<th>Clast/Matrix</th>
<th>Aspect Ratio</th>
<th>GSD values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pale green cataclasite</td>
<td>OU80605</td>
<td>0.09</td>
<td>0.03</td>
<td>--</td>
<td>--</td>
<td>0.12</td>
<td>1.69</td>
<td>-1.651</td>
</tr>
<tr>
<td></td>
<td>OU80606</td>
<td>0.11</td>
<td>--</td>
<td>0.04</td>
<td>--</td>
<td>0.15</td>
<td>1.6</td>
<td>-1.77</td>
</tr>
<tr>
<td></td>
<td>OU80607</td>
<td>0.08</td>
<td>--</td>
<td>0.1</td>
<td>--</td>
<td>0.18</td>
<td>1.61</td>
<td>-1.33</td>
</tr>
<tr>
<td></td>
<td>OU80610</td>
<td>0.05</td>
<td>0.02</td>
<td>--</td>
<td>0.015</td>
<td>0.085</td>
<td>1.73</td>
<td>-1.901</td>
</tr>
<tr>
<td>Augen mylonite</td>
<td>OU77805</td>
<td>0.04</td>
<td>--</td>
<td>--</td>
<td>n.m.</td>
<td>0.04</td>
<td>1.59</td>
<td>-1.22</td>
</tr>
<tr>
<td>Schist mylonite</td>
<td>OU77809</td>
<td>0.02</td>
<td>--</td>
<td>--</td>
<td>n.m.</td>
<td>0.02</td>
<td>1.63</td>
<td>-2.11</td>
</tr>
</tbody>
</table>
Figure 5.2: $N$ vs. $u$ plots of figure 5.1 with the fine fraction ($< 6$ µm) removed, revealing less scatter of the remaining points from the line of best fit. Higher $D$ values characterise these power law size-frequency plots.
Figure 5.2: Continued
area fraction, even though relative to the mica, very few of these clasts were present. This can be explained by the relatively large diameters of these clast types. Relatively low clast to matrix ratios characterise the pseudotachylytes hosted within the mylonites.

The aspect ratios show a high level of consistency between not only the different mineral phases analysed in the samples, but also between the samples themselves. In addition 29 vesicles in the augen mylonite-hosted pseudotachylyte were analysed in two images, and they were found to have a volume fraction of ~0.007.

5.4 Discussion and interpretations

The grain-size reduction process of cataclasis is known to be a self-similar process, and that the products of this fracturing show a power law size-frequency distribution (Sammis et al., 1987; Shimamoto & Nagahama, 1992). The grain size distribution analysis of the Alpine Fault pseudotachylytes exhibit a power law size-frequency distribution, and therefore it may be interpreted that the clasts analysed within the pseudotachylyte underwent brittle comminution in a self-similar fashion. From the power law relationship alone, there is no indication of processes involved in pseudotachylyte formation.

As discussed in chapter 1, the origin of pseudotachylyte has been an issue of controversy since Wenk (1978) doubted the matrix of pseudotachylytes is derived from a melt phase, due to its ultrafine grained and often devitrified texture. Microstructural studies have contributed largely to resolving the controversy between a ‘crushing origin’ and a ‘melting origin’ for pseudotachylytes. Studies by Shimamoto & Nagahama (1992), and Tsutsumi (1999) in which it was shown by size distribution analyses that the area of the ultrafine-grained clasts predicted from power law clast size distributions is less by approximately 1-2 orders of magnitude than the actual measured area of ultrafine material, and thus argued that the matrices of pseudotachylytes cannot be produced by comminution alone. Furthermore, Ray (1999) proposed that clast size distributions in pseudotachylytes deviate from the straight-line power law distribution in the finer fraction of grain sizes (a ‘left hand fall off’ pattern), and attributed this to the transformation of cataclastically deformed rocks to pseudotachylytes by the pervasion of melts generated by frictional heat dissipation on localised slip surfaces. Ray (2004) proposed, using numerical modeling, that the modification of the initial power law size distribution generated by cataclasis (pre-melting) occurs by uniform rim melting of all clasts entrained in the melt, and that the modification should be gradual as there is no possibility of abrupt reduction in the ultrafine fraction. Pseudotachylyte is therefore a mixture of very finely crushed material and quenched melt, therefore indicating that pseudotachylyte is generated by a combination of cataclasis and melting of rocks within a slip zone.

The grain size distributions drawn from the results of the clast analysis of Alpine Fault pseudotachylytes show a power law distribution within the coarser grain sizes, while the finer fraction of the clasts (< 6 µm) deviate from the best fit straight line of the power law. It is shown that when this finer fraction is eliminated from the graphs the exponent of the power law increases from the its lower value when the fine fraction is included. This deviation or ‘left hand fall off’ can be interpreted as being the result of melting of some of the finer fraction of cataclastic material by the friction melt, and is in agreement with grain size analysis and modeling by previous workers (discussed above). The fall off of the plots of the size-frequency distributions in this study are quite drastic, which could either indicate that a high proportion of the finely comminuted material was incorporated into the melt phase during the melting event, or that the
fall off has been exaggerated due to the methods used in the image analysis. The former is less likely, as Ray (2004) pointed out that given adequate heat supply to melt grains of size (using his analogy) $\omega$, the grains of size $(\omega + \delta)$ will survive as clasts of size $\delta$, which may have any value above $\omega$. He concluded that it is possible to have ultrafine material present, and that the fall off should be gradual and gentle, as was shown in Ray (2004).

The drastic fall off of the ultrafine fraction observed in the results is therefore more likely to be a result of the image analysis methods. Manual outlining of the clasts on the BSE images is likely to have contributed to exaggerating the fall off, as the very small clast sizes are very easily overlooked. There is a high possibility that many were in fact overlooked as some of the clasts were observed on close inspection to not stand out strongly from the matrix. The resolution and focus of some of the images also limited the quality and consistency of clast outlining. This oversight in the counting could be wholly or partially responsible for the exaggerated drop off. This highlights the need for high quality, high resolution images and consistent processing in clast analysis, that would give unambiguous results. Nevertheless, this modification of the finer fraction although drastic, has not lead to destruction of the power law size distribution created by initial, pre-melting cataclasis.

If it is interpreted that when the fine fraction (< 6 µm) is removed from the graphs and the resulting best fit lines with higher $D$ values represent the size-frequency distributions of clasts produced by cataclasis alone, then the cataclastic processes may be interpreted. It has been shown in various studies (e.g. Sammis & Biegel, 1989; Blenkinsop & Sibson, 1992) that clast size distributions of fragmented materials have fractal distributions, and the exponents of the power law are the fractal dimensions of the distribution. Natural examples of fragmented fault rocks provide evidence for three distinctive, progressive processes of brittle grain size reduction with increasing strain and confining pressure:

1. Alteration-enhanced microcracking that produces mode-I shear fracture fillings, often due to hydrous alteration reactions resulting in volume increases (Blenkinsop, 1991; Blenkinsop & Sibson, 1992).
2. Constrained comminution in which it is thought particles fail by tensile failure so as to minimise the probability of a neighbouring particle having a similar size (Sammis et al., 1987; Sammis & Biegel, 1989).
3. The selective fracturing of larger particles (Sammis & King, 2007).

The three processes are each characterised by a unique range of fractal dimensions (Blenkinsop, 1991; Blenkinsop & Fernandez, 2000; Sammis & King, 2007), which can be used to establish the grainsize reduction mechanism from comminuted rocks where the grain size distributions are known. Adding one to the $D_2$ values from this analysis yields fractal dimensions that describe the three-dimensional grain size distributions, as Sammis et al. (1987) have shown as a solution to the discrepancy between two- and three-dimensional grain size distributions (discussed in section 5.3). The fractal dimensions derived from the $D_2$ values are ~3-3.4 for four of the samples, but are ~2.4 for samples OU80607 and OU77805. The higher fractal dimensions in four of the samples are characteristic of fractal dimensions established for process 3, the selective fracturing of larger particles, which produces cataclasite or ultracataclasite (Blenkinsop, 1991). This process is the final process in the evolutionary sequence of cataclasis, occurring at high strain-rates. Therefore this process should be expected to have produced the pre-melting ultracataclasites, with grain-size distributions described by the four $D_2$ values, at seismic strain rates.

The two samples that yield the lower fractal dimensions are characteristic of process 1 or 2, microcracking and constrained comminution that produce shear mode fractures and breccias. These two samples are injection veins of
pseudoatoclyte. Although it has been established that ultracataclasis immediately precedes friction melting in fault veins, the process is different for injection veins, which are effectively hydraulic extension fractures that emanate from the fault vein and are infilled with friction melt. No ultracataclasis would precede the pervasion of the friction melt, as there is no shearing along the injection vein to reduce the grain size. The grain size reduction process in the injection veins is more likely to be by implosion and fluid-driven fracturing producing a coarse breccia, which would have instantaneously been infiltrated by a fast moving, clast laden melt. This process therefore unsurprisingly yields a lower fractal dimension indicating initial brecciation with little or no further comminution that produced very little ultrafine material. Another point to consider is that two of the three images analysed for sample OU77805 were taken at lower magnification (80x), however, lower magnifications should give higher D values and fractal dimensions as the finer fraction will be under counted. The effect of the magnification is negligible compared to the process that has been interpreted to yield the lower fractal dimension of the sample.

The relatively low clast to matrix ratios of the mylonite-hosted pseudotachylytes may be due to more heat generated within the veins, or a more finely comminuted starting material that may have resulted a higher fraction of clasts being incorporated into the melt.

5.5 Conclusions

1. The clasts in the samples of Alpine Fault pseudotachylytes that were analysed show power law size-frequency distributions, with the power law exponent being lower for the finer size fraction (< 6 µm) than the coarser size fraction.

2. The frictional heat generated during seismic slip along shear surfaces in the Alpine Fault zone has lead to melting of part of the fine fraction, possibly by uniform rim melting of the clasts produced by comminution. This process has lead to the different power law exponents for the fine and coarse size fractions, with the initial power law distribution of the pre-melt cataclasite, interpreted as being the distribution of the coarse fraction (> 6 µm), only being modified and not destroyed by the subsequent melting. The power law size distribution of the clasts in the pseudotachylytes show the modified pattern, with depression of the finer fractions from the best fit lines.

3. In the fault veins sample, the power law size–frequency distributions interpreted as being the products of pre-melting cataclasis (> 6 µm) have fractal dimensions characteristic of ultracataclasis at high strain rates, indicating that they formed during high strain seismic event. Injection vein power law size-frequency distribution yield lower fractal dimensions, indicating a process of rapid brecciation, consistent with fracturing processes that are thought to form the injection veins.
Chapter 6

Pseudotachylyte Formation Conditions

6.1 Introduction

This chapter combines the results of the field observations and the petrographic, geochemical and microstructural analyses for attempts at estimation of the ambient conditions and the conditions encountered during formation of the Alpine Fault pseudotachylytes. Melt temperatures, host rock temperatures and depths of formation are all estimated. These parameters are then used to constrain pre-failure crustal stress levels, and derive implications for paleoseismicity on the Alpine Fault. A discussion of the rheological implications for the fault zone are also included, followed by the main conclusions to summarise the thesis.

6.2 Melt Temperatures

The temperatures attained by friction melts can be broadly constrained by melt temperatures by the observed presence or absence of certain mineral phases that comprise the host rock, as clasts in the pseudotachylytes. An absence of a mineral phase that appears in the host rock would suggest that this mineral has been incorporated into the melt, and that the melt must have been in excess of the minerals melting temperature. However, the liquidus temperatures defined for the minerals are experimentally determined for igneous melting processes, and therefore caution should be taken when making estimates of friction melt temperatures.

In most of the pseudotachylytes biotite, muscovite and chlorite were all absent as clasts, but were observed in abundance in all the host rocks, which indicates they have been melted. Biotite and muscovite are known to have low melting points on the order of ~ 650°C. Toy (2007) measured biotite Fe/(Fe+Mg) ratios of 0.4 in the Alpine Schist host, a composition that melts at ~1000°C (Best, 2003). Amphiboles in the mylonites are not present in the pseudotachylytes, and owing to their relatively low melting temperatures of ~750-850°C, this observation is probably not surprising. Plagioclase feldspars are observed as clasts in some of the pseudotachylytes but their volume fractions are quite low (table 5.2), compared to their high abundance in the host rocks, indicating most of them were melted. The plagioclase clasts that are observed are highly embayed having undergone partial incorporation into the melt. The plagioclase compositions in the Alpine Schist are approximately An_{20} (Toy, 2007), a composition that has an anhydrous melting temperature of ~1325°C (Winter, 2001), while the compositions of the plagioclase porphyroclasts in the augen mylonites were measured by the simple Michel-Levy method to be ~ An_{30-50}, compositions which have an anhydrous melting temperature of ~ 1400°C (Winter, 2001). Best (2003) showed that the hydrous melting temperature of An_{20} plagioclase at 5 kbar is ~950°C Quartz is the most abundant survivor clast, generally comprising ~65 – 70% of the total volume fraction of the clasts, and is commonly observed to be rounded and embayed, suggesting that it has experienced partial rim melting. Quartz has an anhydrous melting point of ~1760°C (Spray, 1993), that can depress to ~1100°C for hydrous melting (Kennedy et al., 1962). Hydrous melting of all these mineral phases will occur at lower temperatures than those stated above, and as the friction melts are likely to be hydrous due to the melting of hydrous phases such as...
micas and amphiboles, the melting temperatures should lie somewhere in between the anhydrous and hydrous melting temperatures of the minerals.

The melt temperature must have been above the melting temperature of micas and amphibole, as they are not present as survivor clasts, and was likely to be closer to the melting temperatures of feldspar and quartz as they are rounded and, especially feldspar, irregularly embayed around their edges. An estimate of the average melt temperature for the Alpine Fault pseudotachylytes of ~1200°C therefore seems reasonable. This agrees with the estimate that Toy (2007) made on pseudotachylytes from the Alpine Fault zone.

6.3 Host Rock Temperature and Depths of Formation

It would be helpful to estimate the ambient crustal temperature in the pseudotachylyte formation environment, and in turn estimate the depth of formation, so that pre-failure stress levels can be found. The estimates of crustal temperature and depth of the pseudotachylytes are shown in table 6.1. Pressure-depth estimates for pseudotachylyte generation have been estimated by a variety of methods in previous studies. Using analysis of fluid inclusions in pseudotachylyte glass, Boullier et al. (2001) estimated depths for pseudotachylytes in the Nojima Fault zone in Japan. This type of analysis could potentially be carried out on glassy pseudotachylytes reported and analysed by Norris & Cooper (2007), which were collected from the Gaunt Creek section.

Vesicles in pseudotachylytes have also been used to estimate the depth of pseudotachylyte formation, assuming the vesicles formed when the melt was still molten. Maddock (1987) used vesicle abundance, melt compositions and pressure-dependent solubility functions of H₂O and CO₂ and calculated lithostatic pressure and therefore the depth at which the vesicles grew. Other studies have used the same principle for pressure estimates in vesicle-bearing pseudotachylyte veins (e.g. Barker, 2005), and could be applied in this study as three of the thin sections were observed to contain vesicles (chapter 4). However, Toy (2007) determined with a model of pseudotachylyte cooling rates, that it would take a 5 cm thick pseudotachylyte approximately 2 minutes to cool from 1200°C to below a probable solidus temperature of 800°C, and it was shown by Dixon and Dixon (1989) that exsolution of a 200 µm bubble from a melt would take approximately 200 minutes. Therefore, vesicles should not be able to form in the short-lived melt, unless volatiles were entrained in the melt immediately following its formation, as was suggested by Dixon and Dixon (1989). Apart from possibly vesicles, no clear, conclusive evidence has been observed in this study to indicate the presence of a free volatile phase during pseudotachylyte formation. Estimation of depth using the presence of vesicles will therefore not be carried out here.

Although the vesicles will not be used to calculate lithostatic pressures, they are helpful for estimating ambient crustal temperature because in two samples many of them were filled with plastically deformed quartz. One of the pseudotachylytes in the schist-derived mylonite contains vesicles that are infilled with quartz that exhibits undulose extinction and sub-grains. One of the augen mylonite-hosted pseudotachylytes contains vesicles infilled with quartz that exhibits smooth, broad undulose extinction, with no sub-grains or deformation bands. This suggest that the pseudotachylyte hosted in the schist-derived mylonite were formed at higher temperatures in the crust, as sub-grain development in quartz is at a higher temperature stage of recovery than the process that forms just undulose extinction. Sub-grain development in quartz occurs due to sub-grain rotation recrystallisation, and as the development of sub-grains is not pervasive throughout the vesicles in the schist-hosted pseudotachylyte, the recrystallisation probably occurred at
the boundary of regime 1 and 2 for the dislocation creep of quartz of Hirth and Tullis (1992). This boundary occurs at around 400°C for a strain rate of $10^{-12}$ s$^{-1}$, which is the typical strain rate estimated from the Alpine Fault mylonites (Toy, 2007). The pseudotachylytes in the schist-derived mylonites therefore probably formed in crust of this temperature, which corresponds to a depth of ~8 km using the down-dip geothermal gradient of the Alpine Fault zone proposed by Toy (2007). The deformation of quartz within the vesicles of the augen mylonite-hosted pseudotachylytes corresponds to regime 1 of Hirth and Tullis (1992), and marks the initial stages of plastic deformation in quartz, which occurs at ~ 325-350°C for the typical Alpine Fault strain rate. Therefore this would be a reasonable temperature range for the pseudotachylyte formation in the augen mylonites, corresponding to depths of ~4-5 km.

A method of estimating ambient crustal temperature from clast to matrix ratios in pseudotachylytes has been proposed by O’Hara (2001), in which it was found that the clast to matrix ratio, or ratio of wear to melt ($W/M$) is independent of displacement, area, stress and mineralogy, based on models of melting and frictional wear processes. It was shown that the thermodynamic conversion of mechanical work to heat corresponds to $W/M$ and can be defined as $w/q$, the ratio of mechanical work done to the heat produced or the thermal efficiency. $w/q$ is known, from simple heat engine models where heat is moved from a ‘cold’ reservoir to a ‘hot’ reservoir and converted to work, as the efficiency of the heat conversion process that is dependent on the temperature difference of the hot and cold reservoirs. Frictional melting can be regarded as the reverse of a heat engine, in which work by the fault is done on the cold reservoir (host rock) and heat is generated to produce a hot reservoir (friction melt). The efficiency of this work to heat conversion can therefore be regarded as the wear to melt ratio, in which work done by comminution processes produces heat for the frictional melt. This principle was used by O’Hara (1992) to derive the simple equation for estimating the ambient crustal temperature during faulting:

$$T_{\text{crust}} = (1 - W / M)T_{\text{melt}}$$  \hspace{1cm} (16)

where $T_{\text{crust}}$ is the host rock temperature, and $T_{\text{melt}}$ is the melt temperature (both in degrees Kelvin). $W/M$, as already mentioned is regarded as the clast to matrix ratio of the pseudotachylyte. Using the values of the clast to matrix ratios determined in the image analysis in chapter 5 (table 5.2), and a melt temperature of 1200°C, equation 15 gives very high values of ambient crustal temperature (~900°C). This high host rock temperature is very unlikely to be accurate because the pseudotachylytes would therefore have been generated in the ductile regime, under granulite facies conditions. Furthermore, they would be heavily overprinted by mylonitic deformation, which they are clearly not. A possibility for the high temperatures calculated could either be due to over estimation of the melt temperature, or and under-estimation of the clast to matrix ratio. The latter is more likely as the melt temperature would have to be reduced by at least 600°C to get even reasonable crustal temperatures expected for this fault zone. The clast to matrix determined in chapter 5 may not be accurate and could be underestimates, possibly as a result of the resluotion of the images used for clast outlining. O’Hara’s geothermometer will therefore not be applied here to estimate crustal conditions.

Estimations of crustal conditions can also be made by consideration of the metamorphic conditions in the host rock. Some of the dominantly amphibolite facies Alpine Schist-derived mylonites show retrogression to greenschist facies conditions, mainly along mica rich layers containing chlorite and altered plagioclase, that are thought to have
formed by infiltration of fluids along these layers (Toy, 2007). Toy (2007) observed that pseudotachylytes associated with these zones do not show signs of retrogression, and suggested that they either formed before the retrogression or were too impermeable to allow retrogression to occur, and therefore set an upper bound of ~400°C for these veins. This correspond to a depth of 8 km. Pseudotachylytes in augen mylonites are cross-cut by chlorite, calcite and quartz veins, and the matrices of the pseudotachylytes is in some places altered by chlorite. This indicates temperatures of <300°C and depths of less than 4 km. No indications of temperature or depth were observed in this study in the pseudotachylytes hosted in the pale green cataclasites, however, Toy (2007) noted plastically deformed calcite cross-cut by pseudotachylyte indicates temperatures of formation in the range ~170-200°C, which correspond to a depth range of 2.3-3.5 km.

The depth distribution of pseudotachylytes derived from results in this study (table 6.1) are consistent with estimates for the thickness of the seismogenic brittle crust in the Alpine Fault zone (e.g. 10-12 km, Leitner et al., 2001; 8-10 km, Toy, 2007).

<table>
<thead>
<tr>
<th>Pseudotachylyte host rock</th>
<th>Host rock temperature</th>
<th>Depth</th>
<th>Method used to estimate conditions</th>
</tr>
</thead>
<tbody>
<tr>
<td>Schist-derived mylonite</td>
<td>400°C</td>
<td>~ 8 km</td>
<td>Undulose extinction and sub-grains in quartz filled vesicles. Greenschist facies host rock association</td>
</tr>
<tr>
<td>Augen mylonite</td>
<td>&lt;300 - 350°C</td>
<td>~ 4 - 5 km</td>
<td>Undulose extinction in quartz filled vesicles and greenschist facies associations</td>
</tr>
<tr>
<td>Pale green cataclasite</td>
<td>170 - 200°C</td>
<td>~ 2.3 – 2.5 km</td>
<td>Plastically deformed calcite associations (Toy, 2007)</td>
</tr>
</tbody>
</table>

Table 6.1: Ambient crustal temperature and depth of formation of Alpine Fault pseudotachylytes

6.4 Stress Conditions and Paleoseismic Implications

Estimating prefailure ambient stress conditions in the crust would be helpful as values of shear stress resolved on the fault surface may be found. The depth estimates outlined in the table 6.1 suggest there are a range of depths over which pseudotachylyte was formed in the Alpine Fault zone, but two main groups can be distinguished. Pseudotachylytes formed in schist-derived mylonites at depths of around ≥8 km, are relatively low in volume, and are controlled by the anisotropy of the mylonite. Another, more voluminous group of pseudotachylytes formed in greenschist facies augen mylonites and cataclasites at shallower levels of ~2.5-5 km depth.

Frictional shear strength conditions of the crust can be estimated for these depths from the crustal strength profile proposed by Toy (2007), reproduced in figure 2.04. The prefailure strength of the crust is lower than the post failure strength for the frictional-viscous zone, which is below a depth of ~7 km in the Alpine Fault zone. Above the brittle viscous transition (~5-7 km), the strength of the brittle crust is higher immediately prefailure. This correlates well with the postulated depths of formation for the two groups of pseudotachylyte. The pseudotachylytes that formed at ~8 km depth are just below the brittle viscous transition (figure 2.04), and the prefailure differential stress of the crust at this depth would be below 50 MPa, which would result in low shear stresses resolved on the fault surfaces. This would result in low average shear stresses (τf) on the fault plane during seismic rupture. The relatively low volume pseudotachylytes formed at this depth are likely to be the result of these low τf resolved on the fault plane during rupture (equation 13), and the correlation of melt thicknesses to frictional heat production by Toy (2007) indicates they
formed under relatively small (<0.4m) slip increments. These small displacement events just below the brittle viscous transition were likely to be rapidly terminated as frictional sliding under these stress condition is relatively unstable and the rupture would have progressed into a velocity-strengthening regime that produced negative stress drop. This is because the strength of the crust at depths of 8km in the Alpine Fault zone is transiently increased during seismic rupture to higher values than in the aseismic period.

The pseudotachylytes that formed at shallower depths of 2-5 km, in the brittle regime, would have had much higher ambient shear stresses resolved on their fault planes, prior to failure. The magnitude of differential stress in the crust at these depths in the Alpine Fault zone is in excess of 100-200 MPa, corresponding to prefailure stresses resolved on the fault planes of ~50-100 MPa. Relative to the deeper ruptures in the schist-derived mylonites, the shallower rupture would therefore have had a higher average shear stresses resolved on the their fault planes during seismic slip.

The high average shear stresses result in the relatively high volume veins of friction melt observed in the augen mylonites and the cataclasites. These ruptures also occurred within the velocity-weakening regime of the crust as suggested by Scholz (1998), and therefore slip would have been relatively stable and stress-drops were likely to be higher than in the deeper events. Melt lubrication at these depths is also likely to have contributed to slip weakening as melts are relatively thick, and would have allowed greater stress drops to occur due to larger displacements along the lubricated, low frictional resistance fault surfaces. These larger strain releasing events at shallow depths are likely to therefore have been larger magnitudes as moment magnitude \( (M_0) \) is directly proportional to average displacement. On the other hand, voluminous injection structures and other reservoir sites are commonly observed in these pseudotachylytes, and would have allowed melt to escape from the fault plane, therefore maintaining high shear stresses on the slip surface. However, for melt lubrication to be effective only a thin continuous film of melt needs to be maintained on the slip surface (Hirose & Shimamoto, 2005a).

No accurate estimations of earthquake source parameters (\( \tau_{f(avn)} \), \( \mu_k \), slip weakening distance) can be carried out in this study as field conditions conducive for the realistic estimate of these parameters are considerably lacking in the Alpine Fault zone pseudotachylyte outcrops. Di Toro et al. (2005) outlined that the field conditions necessary for the accurate estimation of earthquake source parameters are the presence of large outcrop exposures that allow for estimation of the volume of pseudotachylyte produced during an event, and offset structural markers that allow for determination of displacements. Outcrops of pseudotachylyte in the Alpine Fault zone do not contain these features, and therefore only the broad constraints on stress levels in the crust and relative magnitudes of events as discussed above, can be attained here.

6.5 Implications for the Rheology of the Alpine Fault zone

This study has shown that pseudotachylytes formed at crustal depths ranging from 2-8 km, predominantly under greenschist facies metamorphic conditions. As outlined in chapter 4 , Warr and Cox (2001) proposed 3 progressive stages of brittle deformation during uplift in the Alpine Fault zone, in which the initial stage of the localisation of brittle deformation involved significant strain hardening with frictional melting and anhydrous cataclasis under amphibolite facies conditions. This stage is proposed to be followed by reaction weakening stages in the shallow crust driven by the growth of phyllosilicates and clay minerals, with no further development of frictional melts. The results of this study suggest pseudotachylyte formation at shallow crustal levels under greenschist facies conditions, and that they were not formed in amphibolite facies rocks that underwent the significant portion of the strain hardening in
the Alpine Fault zone. My results suggest the pseudotachylytes formed during transient strain hardening, and subsequently continued to deform by crystal-plastic mechanisms during the inter-seismic period. These results are in agreement with those found by Toy (2007).

The widespread formation of pseudotachylytes in the upper crustal region of the Alpine Fault zone has made a large contribution to the weakening of the crust and the localisation of strain. Altered pseudotachylyte matrix that contains fine grained chlorite and muscovite, would contribute to fault zone weakening as the altered matrices are later comminuted by cataclasis and gouge formation, as these phyllosilicates reduce friction coefficients and localise deformation (Warr & Cox, 2001; Toy, 2007).

6.5 Conclusions

The key conclusions drawn from this study of Alpine Fault pseudotachylytes are as follows:

1. Three principle fault rocks host pseudotachylytes in the Alpine Fault zone. These are; cataclasites, Western province granitoid-derived augen mylonites and Alpine Schist-derived mylonites.

2. Pseudotachylytes are relatively voluminous in the host rocks that have equilibrated under greenschist facies conditions, compared to those in the schist derived mylonites that retain a predominantly amphibolite facies assemblage.

3. In the pale green cataclasites, pseudotachylyte veins commonly occur along surfaces of highly cemented and indurated material, and are especially common on the surfaces of large quartz and feldspar pods.

4. Microstructural observation of the pale green cataclasite reveal that it was transiently fragmented by coseismically enhanced fluid pressures, and subsequently cemented by silica rich fluids. Multiple cycles of this type of fragmentation have occurred, and only on through going, anastomosing, discrete shears do true cataclastic deformation mechanisms operate. It is on these shears that pseudotachylyte and ultracataclasite is produced.

5. Multiple pseudotachylyte generating events are likely to have occurred in all the host rocks, with refragmentation of earlier pseudotachylytes veins common.

6. Cataclasis and frictional melting were synchronous in all pseudotachylyte-generating events and most of the samples contain excellent features that highlight the transition from cataclastic deformation to frictional melting.

7. Relative to their host rocks, the Alpine Fault pseudotachylytes all have matrices depleted in SiO$_2$, and enriched in alumina, alkalis, and metallic oxides. This is due to inclusion of micas, amphiboles, and feldspar into the melt and exclusion of quartz, a preferential selection relationship determined by a hierarchy of frictional melting susceptibilities controlled by mechanical and thermal properties of minerals present in the host rock.

8. Clasts in the Alpine Fault pseudotachylytes exhibit power law size-frequency distributions, with a fall off from the distribution in the finer fraction of the clasts (<6 µm), suggesting the finer fraction has been melted and therefore the depletion modifies the power law exponent that otherwise would be produced by cataclasis.

9. Pseudotachylytes formed during seismic rupture in the shallow crust (<8 km). Pseudotachylytes formed in the schist mylonites at ~8 km depth and are relatively low in volume and displacement due to velocity weakening.
conditions encountered during rupture. More voluminous pseudotachylytes formed at depths of 2-5 km in augen mylonites and cataclasites, where slip increments and magnitudes were larger.
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