<table>
<thead>
<tr>
<th>Title</th>
<th>Late-interseismic state of a continental plate-bounding fault: Petrophysical results from DFDP-1 wireline logging and core analysis, Alpine Fault, New Zealand</th>
</tr>
</thead>
<tbody>
<tr>
<td>Author(s)</td>
<td>Townend, John; Sutherland, R.; Toy, V. G.; Eccles, J. D.; Boulton, C.; Cox, S. C.; McNamara, David D.</td>
</tr>
<tr>
<td>Publication Date</td>
<td>2013</td>
</tr>
<tr>
<td>Publisher</td>
<td>American Geophysical Union (AGU)</td>
</tr>
<tr>
<td>Link to publisher's version</td>
<td><a href="http://dx.doi.org/10.1002/ggge.20236">http://dx.doi.org/10.1002/ggge.20236</a></td>
</tr>
<tr>
<td>Item record</td>
<td><a href="http://hdl.handle.net/10379/6722">http://hdl.handle.net/10379/6722</a></td>
</tr>
<tr>
<td>DOI</td>
<td><a href="http://dx.doi.org/10.1002/ggge.20236">http://dx.doi.org/10.1002/ggge.20236</a></td>
</tr>
</tbody>
</table>
Late-interseismic state of a continental plate-bounding fault: Petrophysical results from DFDP-1 wireline logging and core analysis, Alpine Fault, New Zealand

J. Townend
School of Geography, Environment, and Earth Sciences, Victoria University of Wellington, PO Box 600, Wellington 6012, New Zealand (john.townend@vuw.ac.nz)

R. Sutherland
GNS Science, Lower Hutt, New Zealand

V. G. Toy
Department of Geology, University of Otago, Dunedin, New Zealand

J. D. Eccles
School of Environment, University of Auckland, Auckland, New Zealand

C. Boulton
Department of Geological Sciences, University of Canterbury, Christchurch, New Zealand

S. C. Cox
GNS Science, Dunedin, New Zealand

D. McNamara
GNS Science, Lower Hutt, New Zealand

[1] We present a geophysical characterization at 0.1–100 m scales of a major plate-bounding continental fault in a late-interseismic state. The Alpine Fault produces Mw≈8 earthquakes every 200–400 years and last ruptured in 1717 AD. Wireline geophysical logs and rock cores extending from one side of the Alpine Fault to the other were acquired in two boreholes drilled in 2011 at Gaunt Creek during the first phase of the Deep Fault Drilling Project (DFDP-1). These data document ambient conditions under which the next Alpine Fault earthquake will occur. Principal component analysis of the wireline logging data reveals that ≈80% of the variance is accounted for by electrical, acoustic, and natural gamma properties, and preliminary multivariate classification enables the lithologies of sections of missing core to be reconstructed from geophysical measurements. The fault zone exhibits systematic variations in properties consistent with common processes of progressive alteration and comminution near the principal slip zone, superimposed on different protolith compositions. Our observations imply that the fault zone has the opposite sense of elastic asymmetry at 0.1–100 m scales to that of the crustal-scale orogen imaged by remote geophysical methods. On the basis of the fault-zone scale asymmetry, the bimaterial interface model of preferred earthquake rupture directions implies a northeastward direction of preferred Alpine Fault rupture. On-going characterization of the structural and hydraulic architecture of the Alpine Fault will improve our understanding of the relationship between in situ conditions, earthquake rupture processes, and the hazards posed by future earthquakes.

Components: 12,939 words, 10 figures, 2 tables.
Keywords: Alpine Fault; DFDP-1; fault zone structure; earthquake cycle; principal component analysis; wireline logging.

Index Terms: 8034 Rheology and friction of fault zones: Structural Geology; 8002 Continental neotectonics: Structural Geology; 8010 Fractures and faults: Structural Geology; 8107 Continental neotectonics: Tectonophysics; 8163 Rheology and friction of fault zones: Tectonophysics; 5104 Fracture and flow: Physical Properties of Rocks; 7209 Earthquake dynamics: Seismology; 1242 Seismic cycle related deformations: Geodesy and Gravity.

Received 15 April 2013; Revised 31 July 2013; Accepted 31 July 2013; Published 23 September 2013.


1. Introduction

1.1. Motivation and Background

[2] The relationship between fault zone structure and earthquake rupture is of fundamental scientific and societal interest, and understanding how geological structures accommodate and modulate earthquake processes requires integration of theoretical, field, and laboratory results [Rice and Cocco, 2007; Sibson, 1992]. Outstanding questions include how physical (tectonically driven) and chemical-hydrological processes interact and affect fault structure and thus rheology at different times in the seismic cycle [Hacker, 1997; Tullis et al., 2007]. The state of the fault zone—its integrated properties—is described by macroscopic parameters as well as microscopic. Many assumptions must be made about how the physical state of a fault evolves, and these are codified in phenomenological models such as the rate and state description of friction [Marone, 1998a]. However, few direct observations have been made of the physical state of an active fault in advance of its rupture. Knowing the physical properties and ambient conditions prevailing immediately prior to a large earthquake would provide primary observational constraints on what processes occur at different times and on different scales during the seismic cycle [Zoback et al., 2010].

[3] In this paper, we describe the current state of a fault that is understood to be late in its cycle of great earthquakes, the Alpine Fault. We consider what may happen in the next earthquake, for which our observations on scales of ~0.1–100 m constitute initial conditions.

[4] The Alpine Fault forms an ~850 km long portion of the Australia-Pacific plate boundary in the South Island of New Zealand (Figure 1) and in the central section has slipped during the Late Quaternary at an average rate of 27 ± 5 mm/yr horizontally and 6–9 mm/yr vertically [Norris and Cooper, 1997, 2001]. Paleoseismic observations reveal that three earthquakes of Mw 7.6–8.2 have ruptured the central Alpine Fault since 1430 AD, the most recent being in 1717 AD [De Pascale and Langridge, 2012; Sutherland et al., 2007]. Recent trenching studies from the southern onland portion of the fault reveal a 24 event record of earthquake activity since 6000 BC with an average, quasi-periodic recurrence interval of 329 ± 68 years [Berryman et al., 2012]. No historic earthquakes larger than ~Mw 6.5 have occurred but the fault exhibits low levels of microseismicity [Boese et al., 2012] and seismic tremor [Wech et al., 2012]. The almost 300 years that have elapsed since the last known rupture of the central section of the Alpine Fault constitute a large fraction of the average recurrence interval. In other words, the Alpine Fault appears to be late in the cycle of stress accumulation that will lead to a future large earthquake [Berryman et al., 2012; Sutherland et al., 2007] and poses the greatest present-day earthquake hazard in southern New Zealand [Stirling et al., 2012].

[5] Several factors make the Alpine Fault an important target for scientific drilling [Townend et al., 2009]: these include the well-determined, rapid fault slip rates and regional plate motion history; the oblique fault kinematics, which result in hanging-wall rocks being exhumed from depths of >25 km on time scales of 1–2 Myr [Little et al., 2005; Norris and Cooper, 2007; Vry et al., 2010]; and the factor noted above of the fault being late in the earthquake cycle. Extensive field studies [Cooper and Norris, 1994; Little et al., 2002, 2005; Norris and Cooper, 2003; Norris et al.,
1990; Sibson et al., 1979, 1981; Toy et al., 2008; Toy et al., 2010] and geophysical experiments [Beavan et al., 2007; Boese et al., 2012, 2013; Ingham and Brown, 1998; Okaya et al., 2007; Stern et al., 2007; Wannamaker et al., 2002] during the last 30 years provide a detailed framework for interpreting borehole observations.

This study focuses on data collected in early 2011 from the central Alpine Fault during the first phase of Deep Fault Drilling Project (DFDP-1) operations at Gaunt Creek, a tributary of the Waitangitaoa River located ~6 km SSW of Whataroa [Sutherland et al., 2012]. Data collected during DFDP-1 provide a continuous transect from the hanging wall (Pacific plate) to the footwall (Australian plate) and enable influential models of fault rock structure originally developed for the Alpine Fault [Reed, 1964; Sibson et al., 1979, 1981] to be examined from petrographic and petrophysical perspectives at a single location and late in the presumed earthquake cycle.

The specific objectives of this paper are to (1) describe the key petrophysical characteristics of lithologies identified separately on the basis of core descriptions and petrographic analysis; (2) determine the petrophysical characteristics diagnostic of each lithology and by which they can be discriminated, using a combination of single parameter and multivariate analyses; and (3) consider the implications of these characteristics and the fault’s overall architecture for models of earthquake rupture phenomena and fault zone evolution during and following the next large Alpine Fault earthquake.

1.2. Structure and Seismotectonics of the Alpine Fault

The hanging wall of the central Alpine Fault consists of foliated, predominantly quartzofeldspathic Alpine schists of up to garnet-oligoclase facies; the footwall comprises Paleozoic graywackes intruded by Devonian-Cretaceous granitoids overlain by a Cretaceous-Cenozoic cover sequence [Cox and Sutherland, 2007; Cox and Barrell, 2007]. The architecture of the fault has been described in detail by multiple authors [see Norris and Cooper, 2007, and references therein] and reveals brittle and ductile structures across a zone ~1 km in width [Sibson et al., 1979]. At Gaunt Creek, a sequence of mylonitic schist and cataclasite overthrusts schist-derived Quaternary fluvioglacial gravels and mylonite-derived talus.
breccia along a 30–39° E dipping fault: in the immediate subsurface, the fault’s dip shallows to subhorizontal at the base of thrust sheets due to gravitational collapse [Cooper and Norris, 1994; Norris and Cooper, 1995]. Seismological analysis yields an axis of maximum horizontal compressive stress in the Southern Alps trending ~115°, implying low resolved shear stresses on the southeastward-dipping Alpine Fault as a whole but near-Andersonian geometries for individual strike-slip and reverse fault segments in the shallow subsurface [Boese et al., 2012].

[9] Warr and Cox [2001] described the clay mineralogies of fault rocks exposed along the central Alpine Fault and discussed the effects of clays on fault zone evolution and strength. At Gaunt Creek, the proportion of clay-sized (<2 μm) particles was found to increase toward the fault from <5% by mass in the hanging-wall mylonite, to 20–50% in the cataclasite, and 45–57% in the gouge. Boulton et al. [2012] confirmed that fault gouges exposed in outcrop at Gaunt Creek contain higher proportions of phyllosilicate minerals (30–45%) than the hanging-wall cataclasite (20%), and concluded that the minerals in gouges (smectite + kaolinite + chlorite + illite/muscovite and rare lizardite) are representative of lower temperatures than those in cataclasite (chlorite + illite/muscovite).

[10] Warr and Cox [2001] developed a model of Alpine Fault clay formation consisting of the following three stages: cataclastic fragmentation of amphibolite facies mylonites during anhydrous brittle deformation at temperatures exceeding 350°C; hydrous chloritization at temperatures less than ~320°C (corresponding to depths shallower than ~6 km, assuming a geotherm of 60°C/km); and the formation of swelling clays, notably smectite, within ultrafine-grained gouge at temperatures lower than ~120°C (depths shallower than 2–4 km). Laboratory analysis confirms that the gouges have lower steady state sliding friction coefficients and matrix permeabilities (by three orders of magnitude) than the hanging-wall cataclasite [Boulton et al., 2012].

[11] The first phase of the Deep Fault Drilling Project (DFDP-1) began in 2011 with the drilling of two vertical boreholes at Gaunt Creek [Sutherland et al., 2012]. The boreholes (DDFP-1A—100.6 m depth; DDFP-1B—151.4 m depth) were drilled ~80 m apart using percussion techniques in the uppermost section followed by wireline coring and some rotary drilling when borehole stability precluded coring. Once drilled, each borehole was logged using a suite of wireline geophysical tools, and detailed analysis conducted of the core samples.

[12] An analysis of key lithological and hydraulic observations made during DFDP-1 has been published previously [Sutherland et al., 2012]. That analysis revealed a mineralogically and hydrologically distinct alteration zone extending into the hanging wall more than 30 m from the principal slip zone (PSZ). The alteration zone is formed predominantly of cemented low-permeability fractured cataclasite and obscures the boundary between the damage zone and fault core. The fault core contains a <0.5 m thick PSZ identified on the basis of core descriptions and wireline logging data, near the base of a 2 m thick layer of gouge and ultracataclasite. Laboratory measurements made on hand-sized samples under 30 MPa confining pressures yield permeability estimates for the fault core and the hanging-wall cataclasite of ~10⁻²⁰ m² and 10⁻¹⁶ to 10⁻¹⁸ m², respectively [Boulton et al., 2012]. In contrast, the permeability of the distal damage zone determined from slug tests is approximately 10⁻¹⁴ m² [Sutherland et al., 2012]; in other words, as much as a six order-of-magnitude permeability difference exists across the alteration zone.

1.3. Petrophysical Characteristics of Active Faults and Fault Rocks

[13] The in situ petrophysical properties of several active continental fault zones have recently been investigated via scientific drilling and wireline logging. Wireline logging data from the San Andreas Fault Observatory at Depth (SAFOD) main borehole reveal a ~200 m wide zone of low electrical resistivity and P and S wave velocities, within which are two ~2 m wide zones of even lower resistivity and seismic velocities that coincide with intervals of caving deformation and thus are demonstrably actively creeping [Zoback et al., 2010, 2011]. Jeppson et al. [2010] examined the low-resistivity, low-velocity “damage zone” by comparing wireline logging data with lithologic and microstructural observations [Bradbury et al., 2011] and concluded that the low velocities in this zone were produced by fractionally weak mineral phases associated with a pervasively sheared fabric. Geochemical analysis has revealed the presence of smectite in a spot-core from 3067 m depth and material collected during an unsuccessful coring run at 3436 m [Schleicher et al., 2007] and smectite-bearing illite at 3992 m [Schleicher et al., 2009]. These phases have been interpreted to be
authigenic clays forming by dissolution at temperatures of $<130^\circ$C.

[14] Low electrical resistivities and seismic velocities associated with an active fault have also been detected as part of the Taiwan Chelungpu Drilling Project (TCDP) [Wu et al., 2008]. The shallowest fault zone recognized in the core from TCDP hole A is inferred to have been the locus of slip during the 1999 Chi-Chi earthquake. The fault zone (“FZ1111”) comprises 1.12 m of “ultrafine grained” clayey gouge that becomes progressively foliated with proximity to a 12 cm thick basal layer of indurated black material of similar mineralogical composition that is interpreted to be the PSZ of the Chelungpu Fault and to have accommodated slip in 1999 [Wu et al., 2008]. The FZ1111 breccia exhibits anomalously low resistivity (reduced by 40% with respect to the nonbrecciated lithology), density, and seismic velocities (reduced by 20–25%) [Wu et al., 2008]. Similar patterns of electrical resistivity, P wave velocity, and neutron porosity were observed at three inferred shear zones in the Nojima-Hirabayashi borehole drilled into the Nojima Fault following the 1995 Kobe earthquake [Ikeda, 2001].

[15] Wireline logging within an inactive strand of the Median Tectonic Line also reveals the fault zone to have distinctive petrophysical properties [Shigematsu et al., 2012]. In this case, the lithologic boundary between mylonitized granitoid rocks of the low-P/high-T Ryoke granitoid terrane in the hanging wall and schistose rocks of the high-P/low-T Sambagawa metamorphic terrane in the footwall coincides with the upper extent of a $\sim$80 m wide zone of reduced electrical resistivity and seismic velocity extending into the footwall. Within the hanging wall, shallow-dipping fault structures defined by thin breccia zones and low electrical resistivity can be identified.

2. Data Acquisition and Analysis

2.1. Core Collection and Lithologic Interpretation

[16] The first paper describing findings from DFDP-1 [Sutherland et al., 2012] employed a preliminary lithologic interpretation of the cores. We present the analysis below with reference to a refined lithologic model [Toy et al., 2012] comprising eight lithologies defined initially from hand specimen-scale core descriptions and informed by subsequent petrographic and geochemical analyses:

[17] Unit 1: Gray-green ultramylonites—medium to dark gray (quartzofeldspathic) or dark green (metabasic) mylonite to ultramylonite. This is the dominant lithology present in the >1 km thick outcrops of shear zone rocks at most locations along the central Alpine Fault zone [Toy et al., 2008]. The mineralogy is quartz + oligoclase + biotite + muscovite ± calcite ± hornblende + accessory minerals.

[18] Unit 2: Brown-green-black ultramylonites—brown, olive green and locally black, very fine-grained, planar foliated or millimeter-diameter feldspar augen-bearing ultramylonite. The mineralogy is quartz + plagioclase + chlorite + epidote + accessory minerals.

[19] Unit 3: Unfoliated cataclasites—brecciated or more highly comminuted schist-derived ultramylonite (Unit 1) and brown-green-black ultramylonite (Unit 2). The extent to which original intact rock has been fractured and sheared varies so they span the spectrum from “fractured protolith” to “ultracataclasite” in a strict sense [Sibson, 1977]. Carbonate forms a pervasive cement, so the rock mass is cohesive despite extensive fractures.

[20] Unit 4: Foliated cataclasites—cataclasites similar to those of Unit 3, but with planar to locally anastomosing fabrics defined by millimeter-spaced solution seams of aligned phyllosilicates. The extent of tectonic comminution increases with proximity to the PSZ; the most highly comminuted materials are barely cohesive and clay rich, and equate to ultracataclasite in a strict sense [Sibson, 1977].

[21] Unit 5: Gouges—medium brown or gray, clay-rich ultrafine-grained gouge containing clasts of Units 1–4, as well as recycled laminated gouges, and rare fragments of footwall lithologies.

[22] Unit 6: Granitoid-gneissic cataclasites—cataclasites containing various proportions of fragmented, quartz-rich granitoids and metabasites, both foliated and unfoliated. The mineralogy is quartz + potassium feldspar + plagioclase ± biotite ± chlorite ± muscovite + accessory minerals.

[23] Unit 7: Breccias—cemented protocataclasite to breccia composed of clasts of quartz + potassium feldspar + amphibole and their retrograde equivalents. This lithology was not encountered in the wireline-logged intervals of either borehole and is not referred to further below.
24 Unit 8: Fluvio-glacial gravels—gravel composed predominantly of Alpine Schist clasts. This lithology was encountered in the basal ~1.5 m of the logged interval in DFDP-1A and is not included in the lithological analysis below.

25 The mineralogical compositions of a small number of core samples were determined using X-ray diffraction (XRD) analyses of random powder disc mounts, providing an indication of gross variations in the distributions of minerals present in abundances exceeding ~5%. All the samples analyzed contain quartz, plagioclase, muscovite, and chlorite. Carbonates (calcite + ankerite) are most prevalent in hanging-wall samples (Units 1–5) and are only sporadically present in Units 6 and 7 beneath the 128 m PSZ in DFDP-1B. Potassium feldspar is only found in the few meters immediately above the PSZ in DFDP-1A (Units 4 and 5) and beneath the 128 m PSZ (Units 6 and 7) in DFDP-1B.

26 The major element geochemistry of the core samples analyzed with XRD was determined using X-ray fluorescence (XRF) or inductively coupled plasma-acoustic emission spectroscopy (ICP-AES). The loss on ignition (LOI) was also measured by determining the mass lost as a result of heating 1.0 g of each core sample to 1000°C. LOI quantifies volatile loss during progressive heating, and thus represents the combined proportion of organic carbon, carbonate, and other hydrous phases (e.g., clay or micas), and interstitial water in a sample.

27 The lithologic model identifies a 0.2 m thick gouge unit (Unit 5) at slightly shallower depths in DFDP-1B than the ~0.3 m wide zone of particularly anomalous wireline log responses. In this paper, we presume that the gouge is centered on the wireline log anomaly, and adjust the DFDP-1B core depths by 0.2 m downward for comparison with the geophysical data. No alteration is made to the DFDP-1A core depths, as the distinctive wireline log response presumed to represent the PSZ coincides with the depth of the gouge in the core reference frame.

2.2. Wireline Logging Data Sets

28 Wireline logs spanning most of the open-hole interval between the casing and bottom of the hole were collected in both DFDP-1A (below 30 m) and DFDP-1B (below 48 m). Only the interval in each borehole logged with multiple tools is discussed here. For DFDP-1A, the interval is 31–92 m: borehole instability precluded logging much beyond the PSZ at 91 m. For DFDP-1B, the logged interval analyzed here is 60–140 m. The deeper of two PSZs intersected by DFDP-1B at ~143.8 m could not be wireline logged due to adverse borehole conditions.

29 This paper utilizes data collected with five logging tools in DFPP-1A and DFDP-1B with 0.5–2 cm sampling, and focuses primarily on the following seven parameters: natural gamma (γ); neutron porosity (Φn); compensated density (ρc); P wave velocity (Vp) and impedance (Zp = Vp × ρc); short-guard resistivity (ρE); single-point resistance (R0), spontaneous potential (SP); and borehole diameter (D). Preliminary processing of each log to adjust for depth errors was undertaken by the logging contractor. Data from repeated runs of the same tool were amalgamated, if necessary, to form a single log for each tool, and the data inspected to identify any spikes or other artifacts.

30 Natural gamma (γ, expressed in American Petroleum Institute units, “API”) is a measure of the amount of naturally occurring radioactivity in the rock mass and is commonly recorded by all logging tools as a means of amalgamating different data sets [Ellis et al., 2007]. The tools used during DFDP-1 measured the total intensity of the gamma ray flux. This means that the natural gamma signal cannot be interpreted simply in terms of the concentrations of specific radioactive elements (most commonly 40K, 232Th, and 235,238U).

31 The neutron porosity log (Φn) is obtained by interpreting the flux of neutrons through the rock mass in terms of elastic scattering [Dewan, 1983]. Although presented as and commonly interpreted as a measure of interstitial water content, the neutron porosity log is also affected by rock type, hydrous minerals, and some trace elements [Ellis et al., 2007]. We have sufficient data to confirm that hydrous minerals and clays, in particular, are present in our samples, so it is not appropriate to interpret the neutron porosity response solely in terms of pore fluid. Clay minerals are also associated with specific trace elements that may affect this log, but we have not yet been able to confirm geochemically how significant this is in our DFDP-1 boreholes. Caution should thus be applied during interpretation of these data.

32 Rock density (ρc) was measured in the DFDP-1 boreholes using a single-source, dual-receiver density tool, which measures gamma ray backscatter and compensates for the effects of changes in borehole diameter.

33 Two estimates of electrical resistivity, short-guard resistivity ρE and single-point resistance R0, were obtained in each borehole. The short-guard
resistivity measurement yields higher resolution than the single-point resistance and is less affected by conductive drilling fluids [Dewan, 1983; Ellis et al., 2007]; we refer mainly here to the resistivity data for those reasons.

[34] Spontaneous potential (SP) is a measure of electrical effects produced by interaction of borehole fluids with variably permeable wallrocks [Dewan, 1983; Ellis et al., 2007]. Assuming negligible flow rates (i.e., neglecting electrokinetic effects), the two major components of the spontaneous potential are a liquid junction potential arising due to ionic concentration gradients at the interface between the invaded zone and the formation fluids and a membrane potential (which exceeds the fluid junction potential by a factor of ~4) resulting from the distribution of negatively charged clay particles [Ellis et al., 2007].

[35] P wave velocity (VP) was measured using a single-source, dual-receiver logging sonde that yields an estimate of P wave slowness over a 0.3048 m (1 ft) interval. Slowness data from the sonic tool were converted to P wave velocity (km/s) and filtered to remove picking (cycle skipping) errors. The P wave velocity and compensated density data have also been multiplied to obtain P wave impedance.

[36] The televiwer tool provides an acoustic image of the borehole wall (represented in terms of the amplitude of a reflected acoustic pulse) and a three-dimensional map of the borehole radius (computed from the two-way pulse propagation time).

2.3. Multivariate Analysis

[37] The individual wireline logging data sets reveal several pronounced differences between the hanging wall, fault core, and footwall of the Alpine Fault and systematic changes in some parameters with depth. However, individual hanging-wall lithologies generally exhibit very similar petrophysical properties, making their discrimination on the basis of average properties difficult. We have, therefore, conducted a preliminary principal component analysis [Benaouda et al., 1999; Davis, 2002] of the DFDP-1B wireline logging data to characterize each lithology with respect to axes defined by the dataset as whole.

[38] The principal component analysis is here applied to an \((n \times p)\) data matrix \(D\) consisting of \(n\) observations of \(p = 7\) parameters (\(\gamma, \rho_c, \Phi_N, \rho_E, R_E, S_P,\) and \(V_P\)). We omit P wave impedance from this analysis, as it is an exact function of other parameters. This analysis enables us to find a small number \((r < p)\) of orthogonal axes that collectively account for as much of the variance in the data matrix as possible; in other words, we project the data matrix from the \(p\)-dimensional space defined by the logging parameters onto the \(r\)-dimensional subspace that optimally accounts for the variation along individual axes [Krzanowski, 1988]. We identify these axes by computing the eigenvectors and corresponding eigenvalues of the covariance matrix of \(D\), 

\[
\text{cov}(D) = D^T D / (n-1)
\]

However, as the input parameters have widely differing numerical ranges, we first normalize each parameter by subtracting its mean and dividing by its standard deviation, yielding a normalized data matrix \(d\) whose covariance matrix \(\text{cov}(d) = d^T d / (n-1)\) is the correlation matrix of \(D\). The \(j\)th principal component of the data set is identified as the eigenvector \(u_j\) corresponding to the \(j\)th largest eigenvalue \(\lambda_j\) of \(\text{cov}(d)\). The proportion of the variance accounted for by the \(j\)th principal component is given by \(\lambda_j / \sum(\lambda_j)\), and the \(n\) principal component scores representing each measurement projected onto this component are given by \(bu_j\).

[39] We have also undertaken a preliminary classification analysis in order to examine whether the wireline logging data can be used to infer lithology in the absence of core. In brief, this involves computing the distance in \(p\)-dimensional space between each measurement (i.e., row of \(D\)) of unknown lithology and all the measurements corresponding to a specific lithology, for all lithologies [Davis, 2002; Krzanowski, 1988]. For each of the \(i = 1, 2, \ldots, 6\) lithologies recognized in the DFDP-1B cores, we compute the mean \(\mu_{(i)}\) and covariance matrix \(C_{(i)}\) of the seven logging parameters used for principal component analysis (\(\gamma, \rho_c, \Phi_N, \rho_E, R_E, S_P,\) and \(V_P\)). We then infer the lithology of a wireline observation \(x\) for which no core exists by computing the scaled (Mahalanobis) distance between \(x\) and each of the six lithologies,

\[
d_{M(i)} = \sqrt{\left( x - \mu_{(i)} \right)^T C_{(i)}^{-1} \left( x - \mu_{(i)} \right)} , \quad i = 1, 2, \ldots, 6
\]

[40] The lithology ascribed to \(x\) is whichever of the six lithologies yields the lowest value of \(d_M\).

3. Results

[41] As noted above, stability problems near the bottom of both boreholes prevented us acquiring wireline logging data fully spanning the PSZ or sampling the footwall in DFDP-1A, or spanning the deeper
PSZ in DFDP-1B. We focus here on the larger DFDP-1B wireline logging data set, and confine the analysis of DFDP-1A that follows to only the most notable features.

3.1. DFDP-1B

The key wireline logging data sets in DFDP-1B are illustrated in Figure 2 and summarized according to lithology in Table 1. In computing average parameters, only those sections of the borehole in which the diameter was within 2 cm of the nominal bit diameter of 12 cm, and the density compensation was less than 0.1 g/cc have been analyzed; these sections are referred to below as “in-gauge.” In other words, sections of the borehole that exhibited pronounced enlargement have been ignored in determining the characteristic log responses to mitigate the effects of washouts, etc. on log responses [Benaouda et al., 1999]. In doing so, we recognize that borehole enlargement may reflect lithological variations and hence constitute a petrophysical observable in its own right.

The DFDP-1B logs exhibit overall increases with depth in natural gamma and spontaneous potential, and a decrease in electrical resistivity (Figure 2). Natural gamma increases from \( \sim 80 \) API at 60 m depth to \( \sim 160 \) API at 138 m. Over the same interval, resistivity decreases from more than \( 400 \Omega \cdot m \) to \( \sim 120 \Omega \cdot m \), and SP increases from \( \sim 165 \) to 220 mV. The SP increase with depth in the hanging wall is particularly systematic and is \( \sim 1.25 \) mV/m.

As summarized in Table 1, the gray-green ultramylonites (Unit 1) have the lowest median gamma value of all lithologies (90 API) and the smallest interquartile range (20 API), but otherwise exhibits very similar logging properties to the brown-green-black ultramylonites (Unit 2) and unfoliated cataclasites (Unit 3). The median electrical resistivity (180 \( \Omega \cdot m \)) and single-point resistance (610 \( \Omega \)) of Unit 2 are higher than those of the other logged lithologies, but Unit 2 is indistinguishable in terms of average properties from the two cataclasite lithologies (Units 3 and 4), with which it is intercalated. In the 100–110 m interval,
Table 1. Median (Interquartile Range) Logging Parameters According to Lithology for DFDP-1B (Top) and DFDP-1A (Bottom)

<table>
<thead>
<tr>
<th>Lithology</th>
<th>( \gamma_0 ) (API)</th>
<th>( \rho_0 ) (g/cm³)</th>
<th>( \phi_0 ) (%)</th>
<th>( \phi_n ) (%)</th>
<th>( \rho_n ) (Ω m)</th>
<th>( R_0 ) (Ω)</th>
<th>SP (mV)</th>
<th>( V_p ) (m/s)</th>
<th>( Z_p ) (kg/m²/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DFDP-1B</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Not cored</td>
<td>110 (40)</td>
<td>2.4 (0.2)</td>
<td>17 (11)</td>
<td>9 (6)</td>
<td>350 (190)</td>
<td>840 (190)</td>
<td>180 (30)</td>
<td>3400 (600)</td>
<td>8100 (1800)</td>
</tr>
<tr>
<td>Gray-green</td>
<td>90 (20)</td>
<td>2.4 (0.1)</td>
<td>16 (9)</td>
<td>10 (4)</td>
<td>170 (250)</td>
<td>580 (240)</td>
<td>230 (40)</td>
<td>3500 (700)</td>
<td>8300 (1900)</td>
</tr>
<tr>
<td>Unfoliated cataclasites</td>
<td>110 (30)</td>
<td>2.4 (0.2)</td>
<td>17 (9)</td>
<td>8 (9)</td>
<td>180 (130)</td>
<td>610 (230)</td>
<td>230 (30)</td>
<td>3600 (1100)</td>
<td>8600 (3000)</td>
</tr>
<tr>
<td>Ultramylonites</td>
<td>100 (40)</td>
<td>2.4 (0.2)</td>
<td>16 (9)</td>
<td>12 (5)</td>
<td>150 (50)</td>
<td>530 (120)</td>
<td>230 (10)</td>
<td>3500 (700)</td>
<td>8300 (1900)</td>
</tr>
<tr>
<td>Gouges</td>
<td>120 (20)</td>
<td>2.1 (0.1)</td>
<td>32 (8)</td>
<td>23 (4)</td>
<td>30 (10)</td>
<td>330 (40)</td>
<td>280 (20)</td>
<td>2600 (0)</td>
<td>5600 (300)</td>
</tr>
<tr>
<td>Granitoid-gneissic</td>
<td>150 (50)</td>
<td>2.3 (0.1)</td>
<td>23 (9)</td>
<td>10 (4)</td>
<td>140 (40)</td>
<td>520 (50)</td>
<td>210 (10)</td>
<td>3000 (300)</td>
<td>6800 (900)</td>
</tr>
<tr>
<td>DFDP-1A</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Not cored</td>
<td>100 (40)</td>
<td>2.4 (0.2)</td>
<td>14 (10)</td>
<td>12 (5)</td>
<td>490 (180)</td>
<td>920 (80)</td>
<td>80 (20)</td>
<td>3800 (400)</td>
<td>7200 (2200)</td>
</tr>
<tr>
<td>Gray-green</td>
<td>110 (40)</td>
<td>2.4 (0.2)</td>
<td>18 (10)</td>
<td>9 (5)</td>
<td>510 (80)</td>
<td>930 (60)</td>
<td>80 (10)</td>
<td>3500 (500)</td>
<td>7900 (1300)</td>
</tr>
<tr>
<td>Unfoliated cataclasites</td>
<td>120 (60)</td>
<td>2.4 (0.1)</td>
<td>15 (8)</td>
<td>8 (4)</td>
<td>410 (230)</td>
<td>920 (150)</td>
<td>80 (20)</td>
<td>3800 (400)</td>
<td>9000 (1100)</td>
</tr>
<tr>
<td>Ultramylonites</td>
<td>120 (50)</td>
<td>2.4 (0.1)</td>
<td>17 (8)</td>
<td>12 (5)</td>
<td>130 (40)</td>
<td>490 (60)</td>
<td>230 (10)</td>
<td>3500 (700)</td>
<td>8300 (2000)</td>
</tr>
<tr>
<td>Foliated cataclasites</td>
<td>110 (40)</td>
<td>2.4 (0.1)</td>
<td>16 (7)</td>
<td>7 (3)</td>
<td>520 (140)</td>
<td>950 (70)</td>
<td>80 (20)</td>
<td>3800 (700)</td>
<td>9000 (1700)</td>
</tr>
<tr>
<td>Unfoliated cataclasites</td>
<td>130 (30)</td>
<td>2.4 (0.1)</td>
<td>16 (7)</td>
<td>11 (3)</td>
<td>270 (70)</td>
<td>840 (160)</td>
<td>100 (10)</td>
<td>3300 (800)</td>
<td>8000 (2000)</td>
</tr>
</tbody>
</table>

Unit 2 exhibits systematic downward decreases in natural gamma and resistivity, and downward increases in neutron porosity and spontaneous potential; below this point, with progressive transition to cataclasite, they do not covary systematically or depart much from uniform levels until the PSZ is reached at 128 m.

The gouge (Unit 5) exhibits markedly different properties from the units above and below, particularly with respect to density (lower by ~0.2–0.3 g/cc than in all other units), neutron porosity (a factor of almost two higher than in the other units) and, most notably, the extremely low resistivity (which attains a minimum value of 23 Ω m at 128.4 m), high spontaneous potential (280 mV at 128.4 m), and low P wave velocity (by almost 1000 m/s with respect to the other hanging-wall lithologies).

The sole footwall lithology recorded in the wireline logging data from DFDP-1B, Unit 6, has the highest median natural gamma value of all units (150 API) and the lowest median seismic velocity of all units except the Unit 5 gouge (3000 m/s). Its median density (2.3 g/cm³) is also lower than all but Unit 5.

Using the acoustic televiewer data from DFDP-1B, we have identified and characterized 411 fractures exhibiting acoustic impedance variations with the surrounding rock in the interval between 50 and 141 m. These fractures are illustrated in Figure 3. As a group, the fractures exhibit a bimodal distribution of poles, with one mode subparallel to the pole to the Alpine Fault plane in outcrop at Gaunt Creek and the other approximately conjugate and defining a moderately north-dipping plane. In other words, millimeter-to-centimeter-scale fractures with orientations similar to that of the Alpine Fault at 10–100 m scales predominate throughout the logged interval, with a subordinate set of conjugate fractures.

Figure 4 summarizes the variations in fracture density in DFDP-1B. Core recovery in the 50–100 m interval was poor (cf. Figure 2) and limited data were collected from below the 128 m PSZ. Nevertheless, below 100 m, the quality of the borehole imagery is reasonably uniform and we thus conclude the overall downward decrease in fracture density to be real. Between 100 and 140 m (i.e., across the PSZ), the cumulative number of fractures is well represented by a quadratic function of depth \( N = -0.05z^2 + 16z - 1100 \), \( r^2 = 0.99 \), meaning that the fracture density decreases with depth \( dN/dz = -0.10z + 16 \) within the hanging wall and across the 128 m PSZ.

3.2. DFDP-1A

As in DFDP-1B, there is a downward increase in gamma and a downward decrease in electrical resistivity over the entire logged interval in DFDP-1A (Figure 5). The SP data reveal a downward increase in SP to ~80 m, where the curve abruptly shifts by ~100 mV, likely because the surface electrode was moved. The data below this depth show a peak in SP at the PSZ, similar to the feature observed in DFDP-1B, and a distinctly lower SP in the gravel below.
3.3. Multivariate Analysis

[51] As described in section 2.3, we have undertaken two types of multivariate analysis to first characterize each lithology using a smaller number of parameters than represented by the wireline logging data set as a whole (principal component analysis) and then examine whether lithologies can be accurately predicted on the basis of the wireline logging data alone (classification analysis).

[52] Principal component analysis is applied to the \( n = 2314 \) in-gauge data from the 60–138 m depth interval in DFDP-1B; the deepest 2 m section of the wireline logging data set is excluded from this analysis because of artifacts in the natural gamma (\( \gamma \)) and \( V_P \) data near the bottom of the logged interval. In DFDP-1B, more than 80% of the variance can be explained using only the first three principal components (and more than 90% using the first four components; Table 2). The first principal component is dominated by the electrical parameters (resistivity, resistance, and spontaneous potential, with loadings of 0.49, 0.51, and \(-0.48\), respectively) and accounts for 50% of the total variance. The second principal component is strongly related to density and \( P \) wave velocity (loadings of 0.65 and 0.53, respectively), and...
accounts for a further 17% of the variance. These first three components, therefore, correspond to bulk electrical properties (dominated by phyllosilicate content), acoustic properties (dominated by fracturing and cementation), and natural gamma (dominated by phyllosilicate content and protolith mineralogy).

[53] The DFDP-1B data are illustrated with respect to the first three principal components in Figure 7. The two mylonitic units, Units 1 and 2, differ in terms of principal component 2 (i.e., their acoustic properties), and the mylonitic and cataclastic units, Units 1 + 2 and Units 3 + 4, in terms of principal component 1 (electrical properties). Unit 5 is clearly delineated from the other lithologies on the basis of its low-resistivity/high-SP electrical properties (i.e., low score on principal component 1) and secondarily its low seismic velocity (low score on principal component 2).

[54] Figure 8 illustrates the results of inferring lithologies based on a classification of the in-gauge DFDP-1B wireline observations in the 60–140 m depth interval. The left-hand column shows the lithologic section constructed on the basis of core samples alone, with missing core represented as Unit 0; the central column illustrates the results of inferring lithology from the wireline log data in intervals where the lithology is already known from core; and the right-hand column is a composite lithologic section formed by augmenting sections of known lithology with lithologies inferred...
in sections of missing core from the wireline logging data.

[55] The proportion of known lithologies that is correctly inferred from the logging data is 73% (second column of Figure 8). The composite lithologic section constructed from known and inferred lithologies (right-hand column of Figure 8) is, therefore, imperfect, but it nevertheless exhibits geologically plausible features. In particular, at depths shallower than ~95 m, the logging data tend to imply intercalated intervals of unfoliated cataclasites (Unit 3) and gray-green ultramylonites (Unit 1), with subordinate brown-green-black ultramylonites (Unit 2); granitoid-gneissic cataclasites (Unit 6) are indicated to be present only below the PSZ; no new instances of gouge (Unit 5) are indicated; no new instances of brown-green-black ultramylonites (Unit 2) are indicated below where they are recorded in core (~116 m) or foliated cataclasites (Unit 4) above where they are recorded in core (~110 m); and narrow intervals of missing core (thin white stripes in the first and third columns of Figure 8) are generally indicated on the basis of the wireline logging data to be of the same lithology as the core surrounding them (e.g., Unit 3 near 104 m).

[56] Separate analyses, not illustrated, show that a threefold lithologic model consisting of hanging-wall, gouge, and footwall lithologies can be reliably classified, with an accuracy of >99% using the Mahalanobis algorithm. Similarly, a fourfold model in which the two ultramylonite lithologies (Units 1 and 2) and the two cataclasite lithologies (Units 3 and 4) are combined yields an accuracy of 90%.

[57] These preliminary results show that the logging data can be used to provide an informative albeit imperfect model of the lithostratigraphy in

![Figure 6. Comparison of wireline logging data in DFDP-1A (blue) and DFDP-1B (red) at equivalent depths with respect to the 91 m PSZ in DFDP-1A and the 128 m PSZ in DFDP-1B. The SP data from DFDP-1A are offset by +150 mV (i.e., shifted rightward) to facilitate comparison between the two boreholes.](image-url)

| Table 2. Principal Component Analysis of the Wireline Logging Data From the 60–138 m Interval of DFDP-1B$^a$ |
|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|
| Parameter      | PC1     | PC2     | PC3     | PC4     | PC5     | PC6     | PC7     |
| $\gamma_0$ (API) | -0.20   | -0.01   | 0.93    | -0.16   | 0.25    | 0.05    | 0.08    |
| $\rho_0$ (g/cm$^3$) | 0.19    | 0.65    | -0.12   | -0.69   | 0.23    | -0.03   | -0.03   |
| $\phi_N$ (%)     | -0.36   | -0.31   | -0.30   | -0.05   | 0.82    | -0.13   | 0.01    |
| $\rho_e$ (g/cm$^3$) | 0.49    | -0.26   | 0.02    | -0.05   | 0.24    | 0.74    | -0.31   |
| $R_e$ (g)        | 0.51    | -0.22   | 0.00    | -0.06   | 0.11    | -0.11   | 0.82    |
| $SP$ (mV)        | -0.48   | 0.30    | -0.15   | 0.10    | -0.05   | 0.65    | 0.48    |
| $V_p$ (m/s)      | 0.26    | 0.53    | 0.09    | 0.69    | 0.38    | -0.09   | -0.03   |
| Eigenvalue       | 3.53    | 1.21    | 0.93    | 0.67    | 0.52    | 0.10    | 0.04    |
| Cumulative percentage explained | 50.43   | 67.69   | 80.98   | 90.61   | 98.04   | 99.45   | 100.00  |

$^a$The table expresses each of the seven principal components (PC1–7) in terms of the original logging parameters.
sections of DFDP-1B in which core was not recovered. Future work in this area will need to address the question of what wireline log parameters are most predictive, based on a more extensive examination of principal components or other multivariate representations of the logging data, and consider how different choices regarding the algorithm used to measure similarity affect the results [Krzanowski, 1988].

4. Discussion

[58] The data from DFDP-1 reveal the fault zone of the Alpine Fault at Gaunt Creek to have similar petrophysical properties to fault zones intersected in other active fault drilling projects. Specifically, the fault core and the PSZ are low-resistivity, low-\(P\) wave velocity features, as observed for the San Andreas, Nojima, and Chelungpu faults and the Median Tectonic Line (see section 1.1). In this section, we consider what the petrophysical, geochemical, and mineralogical characteristics of the Alpine Fault imply in terms of long-term and short-term (earthquake cycle) fault zone evolution and behavior.

4.1. Lithologic and Fault Zone Controls on Petrophysical Properties

[59] Figure 9 illustrates the spontaneous potential and resistivity data from DFDP-1B, colored according to depth and differentiated by symbol according to lithology. Several features are evident:

[60] 1. A strong, negative correlation in the data set as a whole between SP and the logarithm of resistivity;

[61] 2. Different SP/resistivity trends in the hanging wall (depths 60.0–125.0 m; \(R_{\times SP}^{-3.1}, r^2 = 0.87\)) and the footwall (depths 129.0–139.0 m; \(R_{\times SP}^{-2.6}, r^2 = 0.57\));

[62] 3. A systematic increase in SP and decrease in resistivity with depth in the hanging wall;

[63] 4. A slight overall decrease in SP and increase in resistivity with depth in the footwall;

[64] 5. Subordinate clustering of data from each lithology;

[65] Similar features are evident in a plot of spontaneous potential versus single-point resistance (not shown).

[66] We interpret the first of these features to indicate the opposing effects of increasing clay content on membrane potentials (which increase, causing an increasing in spontaneous potential) and resistivity (which decreases due to the higher proportion of clay-bound water). The second feature likely represents bulk differences in mineralogy and microstructure between the hanging wall and footwall. The dependence of both parameters on proximity to the PSZ is consistent in the hanging wall and the footwall, although the latter relationship is much less pronounced; in other words, the closer material is to the PSZ, the higher its spontaneous potential and the lower its resistivity. The final factor highlights gross differences between lithologies in terms of these two parameters. Overall, we interpret the electrical data illustrated in Figure 9 to represent common processes of progressive comminution and alteration affecting mineralogically distinct protoliths on either side of the PSZ.

[67] Figure 10 illustrates the variation of three geochemical parameters obtained via geochemical analysis of DFDP-1B core samples near the 128 m PSZ: potassium content (\(K_2O\)), loss on ignition (LOI), and the chemical index of alteration (\(\text{CIA} = \frac{\text{Al}_2\text{O}_3}{\text{Al}_2\text{O}_3 + \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}}\)), a geochemical proxy for alteration originally developed for the analysis of feldspar weathering.
Also plotted are natural gamma, SP, and resistivity, which exhibit systematic variations with one or more geochemical parameters. In both the hanging wall and the footwall, natural gamma and SP are positively correlated with K2O, SP is negatively correlated with CIA, and resistivity is positively correlated with CIA. The peak in SP at the 128 m PSZ corresponds to local minima in CIA and LOI, consistent with elevated levels of carbon-bearing or water-bearing volatile minerals in the fault core. K2O and CIA are higher in the footwall than the hanging wall, whereas LOI is lower. LOI correlates less well with the wireline logging parameters illustrated in Figure 10, although a weak negative correlation with SP and positive correlation with resistivity is evident.

The strong correlation of natural gamma and SP with K2O and the absence of potassium feldspar in the hanging wall, as well as the downward increase in hanging-wall clay content documented from outcrop samples [Sutherland et al., 2012; Warr and Cox, 2001], support our interpretations of the electrical data in terms of geochemical variations. The downward increase in the hanging wall of natural gamma is also consistent with these interpretations, assuming that the natural gamma signal is dominated by potassium and that the potassium is present mainly in phyllosilicates. In the footwall, the negative correlation between K2O and LOI and the low SP values suggest that most potassium is present in Unit 6 within nonphyllosilicate minerals. We thus infer that higher potassium content and natural gamma values beneath the 128 m PSZ are governed principally by the potassium feldspar present in Unit 6, likely derived from a granitic protolith, rather than by authigenic phyllosilicates.

The general decrease in CIA with depth in the hanging wall and the high LOI likely represent a downward increase in the proportion of carbonate minerals. As noted above, calcite and ankerite are present in the hanging-wall lithologies (Units 1–5), but much sparser in the footwall cataclasite (Unit 6), and those observations agree with the high CIA and low LOI in the footwall.

Despite the hanging wall’s higher fracture density, it has higher P wave velocity (by ~0.5 km/s) and higher density (by ~0.1 g/cc) than the footwall immediately beneath the PSZ (Figure 2): in other words, the hanging wall is less compliant (stiffer) than the footwall. These observations and the low permeability of the hanging-wall alteration zone...
Boulton et al., 2012; Sutherland et al., 2012] suggest that the hanging-wall fractures are at least partially cemented, and based on the petrophysical and geochemical data, we hypothesize that the infilling material is dominated by phyllosilicates, notably clays and micas, and carbonate minerals, particularly calcite and ankerite. The alteration zone and processes of fracturing and sealing within it are hydraulically important components of the fault’s overall permeability structure, and distinct from the damage zone and fault core recognized in other studies [e.g., Caine et al., 1996; Faulkner et al., 2010].

4.2. Fault Zone Architecture and Implications for Earthquake Rheology

Our observations enable us to hypothesize that the Alpine Fault has a northeastward preferred rupture direction. We elaborate on the basis for this hypothesis and important qualifications below.

As discussed in section 4.1, the hanging-wall rocks intersected by the DFDP-1 boreholes are of higher seismic velocity and density, and lower permeability, than the footwall immediately beneath. With the important caveat that our observations of the footwall are more limited than those of the hanging wall (and the total extents of the footwall alteration and damage zones remain to be determined in the next phase of DFDP), the available data thus indicate that the hanging wall is less compliant (stiffer) than the footwall in spite of its higher fracture density.

Differences in compliance on either side of a fault (“bimaterial interfaces”) have been suggested to imply a preferential rupture propagation direction (i.e., unilateral rupture) for vertical faults [Ben-Zion and Shi, 2005; Dor et al., 2006]. According to rupture pulse models, normal stress changes near the tip of a crack traveling along a bimaterial interface are compressive ahead of the rupture front when the crack propagates in the same direction as the more compliant side of the fault, and tensile when the crack propagates in the same direction as the less compliant (stiffer) side of the fault. We note, however, that the effects of compliance contrasts on fault rupture have been disputed by other workers based on simulations incorporating slip-weakening frictional behavior [Harris and Day, 2005] or consideration of the effects of the initial stress geometry on off-fault plastic deformation [DeDontney et al., 2011; Templeton and Rice, 2008].

One consequence of repeated rupture in a preferred direction may be greater cumulative damage on the stiffer side of the fault due to reinforced damage within the dilatational quadrant [Ben-Zion and Shi, 2005]. Asymmetric damage patterns consistent with this effect have been described in several places, notably San Jacinto, San Andreas, and Punchbowl faults at scales of millimeters to meters [Dor et al., 2006], the North Anatolian Fault [Dor et al., 2008], and the Arima-Takatsuki Tectonic Line in southwest Japan [Mitchell et al., 2011].

Ma and Beroza [2008] investigated the effects of a dipping bimaterial interface on earthquake rupture dynamics and the corresponding levels of ground shaking. In the case of a dipping reverse fault with a more compliant footwall than hanging wall (similar to the situation observed at Gaunt Creek), normal stress changes during up-dip rupture propagation counteract the effects of the fault’s dip and produce a broadly symmetric distribution of ground motions. Subsequent work has shown that asymmetric damage surrounding a dip-slip fault can be produced in the absence of a compliance contrast as a result of the asymmetry that arises from interaction of a dipping fault with the Earth’s surface [Ma, 2009]. According to this analysis, the effects of inelastic damage are to reduce

Figure 9. Graph of resistivity ($\rho$) versus spontaneous potential (SP) analysis for DFDP-1B; lithologies are differentiated by symbol shape, and the data are colored according to depth. “Hanging-wall alteration” and “Footwall alteration” illustrate schematically the progressive changes in electrical properties occurring on either side of the PSZ.
ground motions in the hanging wall more than in the footwall.

At first glance, the fracture density data and velocity data from DFDP-1B are consistent with both the bimaterial interface and the dipping-fault models of off-fault damage. Both models imply greater hanging-wall fracturing; moreover, in the case of the bimaterial interface model, the fracture sealing within the hanging wall, we have documented provides a mechanism for reinforcing the original compliance contrast and, in principle, entrenching a preferential rupture direction.

In the case of the Alpine Fault, the bimaterial interface model implies a northeastward direction of preferred rupture (i.e., the direction of motion of the lower-velocity footwall). However, no model yet takes into account topographic effects [Norris and Cooper, 1995, 1997] and dissipative factors such as poroelastic processes [Dunham and Rice, 2008], off-fault heating [Ben-Zion and Sammis, 2013], and the along-strike position of earthquake nucleation presumably also governs rupture propagation. How inelastic processes govern rupture and seismic radiation [Ben-Zion and Ampuero, 2009], and thus seismic hazard, remains a topic of crucial importance. The Alpine Fault provides an opportunity to inform numerical models with direct field observations.

The measurements reported here have been made close to the surface and may not necessarily pertain directly to seismogenic depths; this is particularly true for the footwall, for which we have only limited data. What is important to emphasize, however, is that the scale of our observations is much smaller and more directly relevant to processes operating within the fault zone during earthquake rupture than can be imaged using remote geophysical methods. The mechanism envisaged here by which low electrical resistivities are produced in the alteration zone cannot be directly related to the midcrustal to mantle electrical structure of the orogen [Wannamaker et al., 2002]. Similarly, seismological studies reveal the hanging wall of the Alpine Fault to be of lower $P$ wave velocity [Stern et al., 2007] and higher attenuation than the footwall to depths of $\sim$30 km [Eberhart-Phillips et al., 2008]. Our observations from DFDP-1 suggest that the fault zone has the opposite sense of velocity asymmetry than exists for the orogen as a whole at much larger scales, but this finding is preliminary because we sample only a small part of the footwall at shallow depth.

Figure 10. Illustrative comparisons of wireline logging parameters (black) and geochemical data (red) at the same depths near the 128 m PSZ in DFDP-1B. The gray curves show the continuous wireline logs, which are averaged over 10 cm intervals at the depth of each geochemical measurement to produce the black curves.
4.3. Implications for Changes in Fault Zone State During the Seismic Cycle

Our observations of the Alpine Fault’s petrophysical and hydraulic structure on scales of 0.1–100 m and interpretations regarding the role of phyllosilicates and carbonates in fracture sealing suggest cyclic interplay between processes of fracture creation and sealing, and thus between coseismic and interseismic properties of the fault [e.g., Gratier and Gueydan, 2007; Gratier et al., 2011; Sibson, 1990, 1992; Tullis et al., 2007]. The mechanisms by which fault zone healing takes place presumably vary spatially, temporally, and lithologically [Boulton et al., 2012; Marone, 1998b; Renard et al., 2000], and understanding how their kinetics relate to or govern the seismic reloading process remains a challenging but provocative avenue of future research.

We conclude this discussion by noting that the quasi-periodic recurrence of large Alpine Fault earthquakes documented by Berryman et al. [2012] is consistent with that expected of an evolved fault characterized by a narrow range of size scales, in the context of Ben-Zion’s [2008] model of fault systems’ dynamic modes. According to that model, the range of size scales acts as a tuning parameter that governs the frequency-magnitude versus characteristic earthquake modes of behavior. Accordingly, geological processes (e.g., mode of origin of the crustal-scale fault zone, lithological and hydrological changes during the seismic cycle, fault geometry, off fault damage accumulation) that produce spatiotemporal variations in fault zone strength are intrinsically related to the seismic potential of faults and fault networks. By studying these processes on the Alpine Fault, we can directly observe and quantify the parameters thought to govern its documented characteristic earthquake behavior. Determining which faults exhibit characteristic behavior, and which do not, and describing the first-order geological processes that control this, will ultimately improve our seismic hazard models and probabilistic seismic hazard analysis.

5. Conclusions

The wireline logging data and petrological and geochemical measurements from DFDP-1 provide a rare opportunity to investigate the structural and hydraulic architecture of a large, active fault in its prerupture state. At Gaunt Creek, the central Alpine Fault has a ~2 m thick, very low permeability fault core; this is surrounded by an alteration zone extending >20 m from the PSZ and characterized by lower fracture densities and low permeability, which are in turn surrounded by an asymmetric damage zone wider than sampled in DFDP-1. We can draw a number of conclusions:

1. Fault zone composition—both DFDP-1 boreholes intersected the PSZ and we can relate petrophysical properties acquired using wireline logging tools to geological and geochemical data acquired from core. Electrical properties and natural gamma vary systematically with K2O and geochemical proxies for alteration (CIA) and hydrous phases (LOI). These relationships are consistent with potassium-rich phyllosilicates (clays and micas, particularly) providing the main compositional control on the fault rocks’ overall properties in the hanging wall, and with potassium feldspar dominating the potassium signal in the footwall.

2. Electrical—electrical resistivity and spontaneous potential vary systematically with proximity to the PSZ, in both the hanging wall and the footwall, and define trends consistent with progressive comminution and alteration superimposed on different protoliths. These processes are significant in the context of fault zone mechanics, as they imply that pervasive cementation and permeability reduction operate on scales much wider than the PSZ or fault core.

3. Acoustic—we observe very low P wave impedance close to the PSZ (<1 m) and a bulk impedance contrast across the fault with the footwall having lower P wave velocities (by ~0.5 km/s) and density (by ~0.1 g/cc). The density of fractures detected using acoustic televiewer data decreases downward and across the PSZ at 128 m in DFDP-1B: in other words, the high-velocity, low-permeability hanging wall is more fractured than the low-velocity, high-permeability footwall.

The complementary data sets used here reveal strong correlations between lithological and petrophysical observations—to the extent that the wireline logging data can be used to predict lithology with moderate precision. This redundancy gives us confidence in our interpretations of fault zone processes at 0.1–100 m scales, and provides a basis for drawing petrophysical conclusions at much larger scales from regional geological mapping.

At Gaunt Creek, the Alpine Fault is demonstrably asymmetric on 0.1–100 m scales, or larger, in at least four senses: geometric asymmetry imposed by its nonvertical dip; hydraulic asymmetry governed by lower fracture densities and low permeability, which are in turn surrounded by an asymmetric damage zone wider than sampled in DFDP-1.
by the low permeability of the alteration zone; elastic asymmetry associated with the bulk properties of the wallrocks and superimposed fracturing; and topographic asymmetry reflecting the fault’s large-scale structure and tectonic role. The competing effects of these asymmetries on rupture propagation have yet to be fully explored, but exemplify the difficulty of relating material properties produced by repeated earthquakes to the processes involved in individual rupture events. While the scales at which we have studied the fault in DFDP-1 are much smaller than can be explored using crustal-scale geophysical methods, they are comparable to the scales involved in models of earthquake rupture and seismic radiation.

[87] The P wave velocity and fracture data sets described here imply the opposite sense of elastic asymmetry at 0.1–100 m scales in the near-surface to that exhibited by the orogen as a whole at kilometer and larger scales. On the basis of the fault-zone scale asymmetry, the bimaterial interface model of preferred earthquake rupture directions implies a northeastward direction of preferred Alpine Fault rupture. Further characterizing the structural and hydraulic architecture of the fault zone will improve our understanding of the relationship between in situ conditions, earthquake rupture processes and the documented quasi-periodicity of large Alpine Fault earthquakes, and the hazards posed by future earthquakes.

Acknowledgments

[88] We thank Horizon Drilling, Alex Pyne, the Department of Conservation, White Heron Sanctuary Tours, and the Whataroa community for technical and logistical assistance during DFDP-1. Norio Shigematsu and an anonymous reviewer provided valuable comments on an early version of this manuscript. The wireline logging data set was acquired with the assistance of Gareth Hayes (Subsurface Imaging Ltd.), whose expertise and enthusiasm we gratefully acknowledge. Many of the figures in this manuscript were constructed using Generic Mapping Tools (Wessel and Smith, 1991) and Stereonet V8 (Allmendinger et al., 2012). Funded by: GNS Science; Victoria University of Wellington; the University of Otago; the University of Auckland; the University of Canterbury; Deutsche Forschungsgemeinschaft and the University of Bremen; Natural Environment Research Council and the University of Liverpool; and the Marsden Fund of the Royal Society of New Zealand.

References


Ben-Zion, Y., and J. P. Ampuero (2009), Seismic radiation from regions sustaining material damage, Geophys. J. Int., 178(3), 1351–1356.

Ben-Zion, Y., and C. G. Sammis (2013), Shear heating during distributed faulting and pulverization of rocks, Geology, 41(2), 142.


compaction around active faults, J. Struct. Geol., 22(10), 1395–1407.


